Palaeogeography, Palaeoclimatology, Palaeoecology, 78 (1990): 241–286
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

Chronology of expansion and contraction of four Great Basin lake systems during the past 35,000 years

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(Received February 24, 1989; accepted September 6, 1989)

ABSTRACT

Benson, L. V., Currey, D. R., Dorn, R. I., Lajoie, K. R., Oviatt, C. G., Robinson, S. W., Smith, G. I. and Stine, S., 1990. Chronology of expansion and contraction of four Great Basin lake systems during the past 35,000 years. Palaeogeogr., Palaeoclimatol., Palaeoecol., 78: 241–286.

During the past 35,000 years, Lake Lahontan, Lake Bonneville, Lake Russell, and Lake Searles underwent a major period of lake-level change. The lakes were at moderate levels or dry at the beginning of the period and seem to have achieved highstands between about 15,000 and 13,500 yr B.P. The rise of Lake Lahontan was gradual but not continuous, in part because of topographic constraints (intrabasin spill). Lake Lahontan also had an oscillation in lake level at 15,500 yr B.P. Radiocarbon-age estimations for materials that were deposited in the lake basins indicate that Lake Bonneville rose more or less gradually from 32,000 yr B.P., and had major oscillations in level between 23,000 and 21,000 yr B.P. and between 15,250 and 14,500 yr B.P. Lake Russell and Lake Searles had several major oscillations in lake level between 35,000 and 14,000 yr B.P. The timing and exact magnitude of the oscillations are difficult to decipher but both lakes may have achieved multiple highstand states. All four lakes may have had nearly synchronous recessions between about 14,000 and 13,500 yr B.P. After the recessions, the lakes seem to have temporarily stabilized or experienced a minor increase in size between about 11,500 and 10,000 yr B.P. These data provide circumstantial evidence that the Younger Dryas Event affected climate on at least a hemispheric scale. During the Holocene, the four lakes remained at low levels, and small oscillations in lake level occurred. An important aspect of the lake-level data is the accompanying expansion of lake-surface area at the time of the last highstand. Lake Bonneville and Lake Lahontan had surface areas about 10 times larger than their mean-historical reconstructed areas whereas Lake Russell and Lake Searles had surface areas about 5 times larger than their mean-historical reconstructed areas. Differences in the records of effective wetness may have been due to the locations of the basins relative to the position of the jetstream, or they may have resulted from lake/atmosphere feedback processes.

Introduction

During the late Quaternary, nearly 100 basins in the Great Basin of the western United States contained lakes (Snyder et al., 1964; Mifflin and Wheat, 1979; Williams and Bedinger, 1984). Four paleolake systems (Lake Lahontan, Lake Bonneville, Lake Russell, and Lake Searles; Fig.1) have been intensively studied (Broecker and Orr, 1958; Flint and Gale, 1958; Smith, 1962; Morrison, 1964; Stuiver, 1964; Broecker and Kaufman, 1965; Kaufman and Broecker, 1965; Morrison, 1965;

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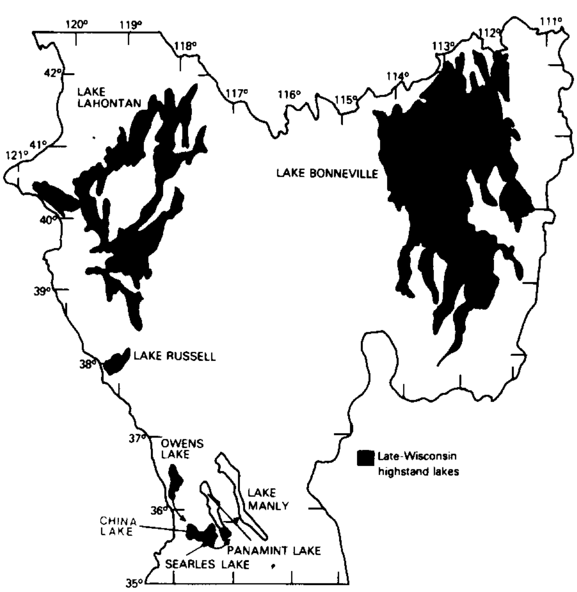

Fig. 1. Four principal paleolake systems of the late Quaternary.

Morrison and Frye, 1965; Lajoie, 1968; Smith, 1968; Born, 1972; Benson, 1978; Davis, 1978; Peng et al., 1978; Smith, 1979; Stuiver and Smith, 1979; Davis 1982; Lajoie, et al. 1982; Scott et al., 1983; Currey et al., 1984; Smith, 1984; Spencer et al., 1984; Stine, 1984; Currey and Oviatt, 1985; Thompson et al., 1986; Benson and Thompson, 1987a; Oviatt, 1988). Time correlation of the four major lake systems with each other and with mountain and continental glaciations was largely presumptive (Russell, 1885; Gilbert, 1890; Blackwelder, 1931; Antevs, 1945) prior to the development and application of accurate numerical methods of age estimation (Libby, 1955; Broecker and Orr, 1958; Kaufman and Broecker, 1965). This assumption of regional- and hemispheric-scale time-correlative changes in major components of the hydrologic balance continues to be made even though unequivocal numerical-age estimations that would confirm this assumption have not been published.

The principal objective of this paper is the assignment of patterns of lake-level and lake-surface area variation to four closed-basin lake systems for the past 35,000 yr. These patterns are assigned depending on the reliability of the age data and the ability to infer lake depth from a specific sample type. In assigning alternative chronologies to the lake systems, the senior author has invoked the working hypothesis that the lakes may have responded synchronously to large synoptic-scale changes in climate. Secondary questions addressed in this paper include: (1) What measure of lake-size variation (lake level, lake area, or lake volume) is the best indicator of change in the hydrologic balance? (2) Were lake-size variations synchronous across the Great Basin? (3) When comparing lake-size variations, what normalization procedure should be used? (4) Were synchronous changes in lake size of the same hydrologic magnitude? (5) What changes in global or local climate were associated with the rise and fall of the lakes?

Methods and limitations of radiocarbon-age estimates

Most of the numerical-age data compiled in this paper were obtained using the radiocarbon method. Since the introduction of accelerator mass spectrometry (AMS), the radiocarbon method can be applied to samples that are less than about 75,000 yr old (Elmore and Phillips, 1987). The accuracy of the radiocarbon method primarily remains limited by sample reliability. Most of the numerical-age data in this paper consist of radiocarbon-age determinations on carbon-bearing materials deposited either above the surface of a lake (soil, rock varnish, packrat-midden macrofossils, gastropods) or below (ostracode valves, lake sediment, tufa, mollusc shells, oolites, marl, wood). To assess the reliability of radiocarbon-age estimates for each of these materials, processes that limit the accuracy of radiocarbon ages need to be discussed.

Radiocarbon-age estimates of soils

Soils in the Great Basin contain inorganic and organic forms of carbon. The inorganic (calcareous) component often consists of carbonate precipitates or accumulations of wind-blown carbonate particles. The organic component usually consists of decaying vegetation. Radiocarbon ages of the calcareous component will be too young if precipitation of modern radiocarbon occurs in void spaces within the older calcareous component; radiocarbon ages of the calcareous component will be too old if the wind-blown component is derived from sediment or rock that contains 'dead' carbon. Chen and Polach (1986) have reported that radiocarbon ages of soil carbonate average about 3600 yr older than radiocarbon ages of coexisting organic carbon. Radiocarbon ages of organic materials from soils usually are more reliable than the ages obtained from inorganic materials. Age determinations on soil organics can be affected by contamination with material derived from younger root systems and by incorporation of fossil carbon dioxide that was formed during the decomposition of soil carbonates (Geyh, 1970; Luders et al., 1970). Incomplete isotopic equilibration of soil water with atmospheric carbon dioxide can occur if the depth of carbonate leaching exceeds several meters (Reardon et al., 1980).

Radiocarbon-age estimates of rock varnish

Rock varnish forms on stable landform surfaces in arid and semiarid environments. Rock varnish mostly consists of clay minerals, manganese and iron oxides (Potter and Rossman, 1977); but it also contains carbon-bearing organic matter (Dorn and Oberlander, 1982; Dorn and DeNiro, 1986). The organic matter consists of: (1) microorganisms present at the surface of the varnish (Dorn and Oberlander, 1982), (2) organic debris incorporated in the varnish as it accretes (Dorn and DeNiro, 1986), (3) undifferentiated organic matter that is loosely bound to clay-mineral ion-exchange sites (Dashman and Stostsky, 1982; Hedges and Hare, 1987), and (4) undifferentiated organic matter that is tightly bound to manganese and iron oxides. The object of AMS measurement of the radiocarbon content of rock varnish is estimation of the time of exposure of the rock surface that underlies the varnish. The organic matter submitted for age estimation should consist of material incorporated in the bottom-most layer of the varnish when it accreted to the rock surface. This material consists of organic matter that is tightly bound to the manganese and iron hydroxides and plant and pollen debris.

The radiocarbon age of basal varnish represents a minimum age for the exposure of a rock surface for several reasons: (1) the onset of varnish formation lags surface exposure by 60–100 yr in arid and semiarid regions (Dorn et al., 1987), (2) the basal layer of varnish may form earlier on one spot than on another, (3) varnish may undergo chemical or physical erosion that resets the varnish clock, and (4) the carbon-bearing basal layer submitted for radiocarbon analysis often composes $\sim 10\%$ of the total varnish thickness and, therefore, represents a certain period of accumulation. For these reasons, varnish ages are about 5–10% younger than the time of surface exposure (Dorn et al., this issue).

Radiocarbon-age estimates of packrat-midden macrofossils

Packrats are rodents that collect various plant materials for food or construction of a den. The dens of packrats in dry rock shelters can become indurated by urine, which forms hard, well-preserved organic deposits that are termed 'middens' (Van Devender et al., 1985). The presence of a water-soluble midden in a lake basin indicates that lake level has remained below the elevation of the midden. Prior to the availability of AMS for routine dating, combined specimens of individual plant species and unwashed pieces of midden matrix were used to provide a sample that contained sufficient radiocarbon for conventional age dating. The assumption of contemporaniety of plants that compose the dated assemblage is correct if the following criteria are met: (1) the sample was collected from a stratigraphically discrete unit, (2) the outer weathering rind was removed, and (3) ancient packrats collected non-fossil material (Van Devender et al., 1985). Radiocarbon ages that are too old by several hundred years can occur if a packrat collects residual (fossil) woody material. Radiocarbon ages that are too young can result if the dated plant assemblage is contaminated with younger plant material or by burrowing organisms.

Radiocarbon-age estimates of lake sediment

Lake sediment consists of: (1) material carried to the lake by streamflow and aeolian transport (allochthonous material), (2) material that originates from processes that occur within the water column (autochthonous material), and (3) material derived from the physical and chemical alteration of allochthonous and autochthonous (diagenetic or decomposed material). Allochthonous forms of organic carbon include: (1) pollen, (2) plant material derived from living vegetation, and (3) fossil plant material derived from older lake sediments and soils. Autochthonous forms of organic carbon include: (1) floating algae, (2) phytoplankton, (3) zooplankton, (4) fish, (5) substrate-attached algae, (6) substrate-attached vascular macrophytes, (7) benthic animals, and (8) bacteria. Allochthonous forms of inorganic carbon include: (1) graphite, (2) wind-borne fossil carbonate that is derived from glacial flour, and playa and soil surfaces, and (3) fossil carbonate that is washed in from the surrounding drainage basin. Autochthonous forms of inorganic carbon include: (1) carbonate precipitates (gaylussite, pirssonite, magnesium calcites, aragonite, monohydrocalcite, dolomite) and (2) ostracodes valves.

The dilution of lake water with dead carbon results in low initial $^{14}C/^{12}C$ ratios of autochthonous carbonates. Surface and ground water that supplies Great Basin lakes frequently come into contact with carbonate rocks of pre-Quaternary age. Dissolution of the carbonate rock decreases the initial $^{14}C/^{12}C$ ratio of the dissolved-carbonate chemical species. The $^{14}C/^{12}C$ ratio for a specific lake is dependent on the $^{14}C/^{12}C$ ratio of carbonate chemical species in water that supplies the lake, the concentration of carbonate species dissolved in the lake, and the rate of exchange of inorganic carbon across the air–water interface (Broecker and Walton, 1959). If the exchange rate is not rapid enough to enable $^{14}C/^{12}C$ equilibration between the atmosphere and the lake, the radiocarbon age of the water will be too old (Broecker and Kaufman, 1965); this phenomena is termed the hardwater or reservoir effect. Smith et al. (1987) reported that prior to and just after crystallization of carbonate salts in Owens Lake, major discrepancies existed between the partial pressures of dissolved and atmospheric CO_2. Just prior to crystallization, CO_2 partial pressure was about six times atmospheric pressure; at the end of crystallization, CO_2 partial pressure was about 2% of atmospheric partial pressure.

In terms of the carbon exchange rate, the $^{14}C/^{12}C$ ratio of carbonate species in a lake is directly related to the area/volume ratio of the lake (Broecker and Walton, 1959). Because the area/volume ratio for most Great Basin lakes increases with lake depth, departure of the initial $^{14}C/^{12}C$ ratio of in-situ carbonates from the atmospheric $^{14}C/^{12}C$ ratio is minimized during highstand periods. Addition of modern carbon during subaerial exposure of the sediment also may lessen the reservoir effect. Data for the Lake Lahontan basin indicate that highstand tufas sometimes yield dates that are about 1500 yr too young (see following section on age of Lahontan carbonates).

The reservoir effect is not limited to inorganic autochthonous material. The $^{14}C/^{12}C$ ratio of organisms that use dissolved carbonate in their life cycles also will indicate the $^{14}C/^{12}C$ depletion of dissolved carbonate. If an organism uses fossil allochtonous (detrital) material, its $^{14}C/^{12}C$ ratio will be depleted and the organism's radiocarbon age will be too old. Fossil detrital material can be supplied by the erosion of older lake sediments during the rise or fall of a lake. As a lake advances or recedes, erosion and transport of older lake sediment that lines the stream channels occurs, which results in the transport of fossil organic and inorganic carbon to the lake basin.

Diagenetic alteration of carbonate minerals is a common process. During the chemical evolution of a closed-basin lake, the initial precipitation of carbonates that are enriched in calcium relative to magnesium results in an ever-increasing magnesium/calcium ratio in the lake water. This process often results in the sequential precipitation of the following carbonate minerals: low-magnesium calcite, magnesium-enriched calcites, aragonite, and sometimes monohydrocalcite (Jones and Bowser, 1978). These minerals and calcitic ostracodes that live at the sediment–water interface are not necessarily in equilibrium with the lake water from which they precipitate, nor are they in equilibrium with the interstitial water that surrounds them after burial. The interstitial and water-column chemistries of lake water continually evolve through time; therefore, a tendency exists for the metastable carbonate precipitates to recrystallize into more stable forms. When recrystallization occurs, the $^{14}C/^{12}C$ ratio of the surrounding fluid is partitioned into recrystallized or newly precipitated carbonate minerals. The apparent radiocarbon age of these secondary carbonate minerals will not necessarily be the same as the time of formation of the original mineral.

Decomposition of allochthonous and autochthonous particulate organic carbon occurs as organic material falls through the water column and as the material accumulates at the bottom of the lake. Large particles (leaf and algal detritus) are colonized by bacteria and fungi and decompose at a rate of about 1% per day. Fine particles of phytoplankton and zooplankton decompose at a rate of about 10% per day (Saunders, 1976). Residual organic matter that accumulates on the bottom of a lake undergoes aerobic decomposition by macroscopic organisms, fungi, and bacteria. As oxygen becomes depleted in the interstitial water, anaerobic decomposition becomes the predominant process. The quantity and type of organic matter preserved in lake sediment depends on the bulk rate of sedimentation and the resistance of the organic material to decomposition.

Bacterial contamination of wet-sediment cores can result in unreliable radiocarbon-age estimates. Geyh et al. (1974), in a study of marine-sediment cores, reported that bacterial activity at 4°C resulted in as much as 10% contamination by modern carbon. The degree of contamination is dependent on air temperature, humidity, infiltration of air, the migration rate of the bacteria into the core, the quantity of organic carbon in the core, and time. Contamination of a sample by bacterial activity after collection can be avoided by: (1) deep-freezing, (2) drying, or (3) immediate sample processing after core recovery (Geyh et al., 1974).

Radiocarbon-age estimates of tufa, oolites, molluscs, and chara

Tufa, oolites, molluscs, and *Chara* are biologically mediated precipitates of calcium carbonate. These materials usually form in the shallow benthos of a lake and can be used to provide minimum estimates of lake level. Except for gastropods, which consist of aragonite, these materials usually consist of low-magnesium calcite. However, one form of tufa (thinolite) forms at low temperatures as ikaite ($CaCO_3{\cdot}6H_2O$) and later recrystallizes to calcite ($CaCO_3$) (Shearman and Smith, 1985). These carbonates are susceptible to the same processes that affect other autochthonous carbonates (low initial $^{14}C/^{12}C$ ratio and diagenetic alteration after burial). Other chemical and physical processes can affect the radiocarbon ages of the biologically mediated carbonates: (1) preferential fractionation of the heavier isotopes of carbon during precipitation (biological-isotope effect; Nier and Gulbransen, 1939), (2) ingestion and incorporation of dead carbon by molluscs, and (3) incorporation of dead carbon by molluscs and tufas from springs that discharge through the lake bottom (Riggs, 1984). Within-lake contamination of porous carbonates, such as the dendritic varieties of tufa, also can occur by incorporation of secondary carbon-bearing sediment or secondary carbonate cement (Benson, 1978). Addition of secondary inorganic carbon by a dissolution/reprecipitation process also can occur in the subaerial environment. If contamination occurs on the submicroscopic level, detection of the contaminant is made extremely difficult (Benson and Thompson, 1987a).

Radiocarbon-age estimates of wood from deltaic and lagoonal environments

Wood from deltaic and lagoonal environments provides extremely reliable material for radiocarbon-age estimates. The radiocarbon age of the outermost tree ring provides the age of the death of the tree or the age of the uprooting of the tree. Processes that delay transport of the woody material to its final depositional site may limit the usefulness of the radiocarbon-age estimate; e.g., exhumation of fossil wood previously buried at some upstream site.

Limitations of materials used to estimate lake levels

With the possible exceptions of carbonate-cemented beach sediment and rock varnish that coats erosional terraces, the materials discussed above cannot be used to exactly determine lake level. Most often these materials only provide some minimum or maximum constraint on lake level. Tufa can form over a wide depth range, and materials that form in shallow-water environments (gastropods, oolites, and small tufas) can be transported to deeper environments by wave action. Soils and packrat-midden macrofossils can form at elevations substantially higher than that of lake level. Lake depth can only be approximately deduced from the fabric and texture of lake sediment. The use of diatom and ostracode assemblages to infer water depth (or lake chemistry) remains at best a semi-quantitative procedure (Forester, 1987; Bradbury, 1988; De Deckker and Forester, 1988) when applied to large closed-basin lakes. These limitations indicate that lake-level curves constructed from available data are, at best, approximations of reality.

Lake Lahontan

The Lahontan basin consists of seven subbasins separated by sills of varying altitude (Fig.2). Six rivers terminate in the subbasins; four (Truckee, Carson, Walker and Humboldt) now contribute 96% of the total gaged surface inflow (Benson, 1986). The Truckee, Carson, and Walker Rivers have their headwaters in the Sierra Nevada on the western boundary of the Lahontan basin. The Humboldt River drains mountain ranges on the northeast (Figs.2 and 3). Subsurface inflow to surface-water bodies in the Lahontan basin is small compared to total water input as runoff and precipitation (Everett and Rush, 1967; Van Denburgh et al., 1973).

Russell (1885) did the first comprehensive study of lake deposits of Pleistocene age in the Lahontan basin. He identified a Lower Lacustral Clay and an Upper Lacustral Clay which are separated by a Medial Gravel. Morrison (1965) renamed the Lower Lacustral Clay the Eetza Formation, the Medial Gravel the Wyemaha Formation, and the Upper Lacustral Clay the Sehoo Formation. The Sehoo Formation is considered to have been deposited during the last lake cycle.

Broecker and Orr (1958) and Broecker and Kaufman (1965) attempted to assign radiocarbon age estimates to these deposits, but Morrison and Frye (1965) reported that certain radiocarbon ages of tufas were reversed when compared to Morrison's (1964) stratigraphic assignments. Benson (1978, 1981) developed an alternative lake-level chronology for the Pyramid and Walker Lake subbasins by using a selection procedure that tended to eliminate problems caused by the introduction of secondary carbon into tufa samples. Several types of tufa samples were rejected including: (1) tufas that probably formed by the mixing of ground water and lake water, (2) large porous tufas that were probably contaminated with younger detrital carbonate, and (3) tufas that consisted of prismatic crystals (thinolite) that may have recrystallized from a metastable carbonate phase. Age estimations of tufas that did not evidence microscopic-scale contamination and unrecrystallized-aragonitic gastropods formed the basis of the lake-level chronology developed in these studies. Prior to analysis, carbonate samples were acid leached to remove surficial contaminants.

Thompson et al. (1986) presented a revised Lake Lahontan chronology for the Pyramid and Winnemucca Dry Lake subbasins. The revision was primarily based on radiocarbon-age estimations of leaves and twigs preserved in packrat middens in caves of the Winnemucca Dry Lake subbasin. The radiocarbon ages of the plant material provided conclusive evidence that the last highstand of Lake Lahontan terminated before $12{,}070 \pm 210$ yr B.P and not at $11{,}430 \pm 160$ yr B.P. as deduced by Benson (1978). These observations indicated the possibility that tufas can be contaminated on a scale that is virtually impossible to detect by using conventional microscopic techniques. Dorn et al. (this issue) obtained a minimum-limiting radiocarbon age of $12{,}680 \pm 105$ yr B.P. on the basal layer of rock-varnish organic matter from the high terrace at the northern end of Pyramid Lake. Three tufa samples (samples I-9326, I-10001, I-10002, Table 1), located 4–9 m below the Lahontan highstand (1330 m), have radiocarbon ages less than 12,680 yr B.P. These radiocarbon ages are too young or they indicated the initial recession of lake level.

Varnish dates may be as much as 1500 yr younger than the time of surface exposure of the

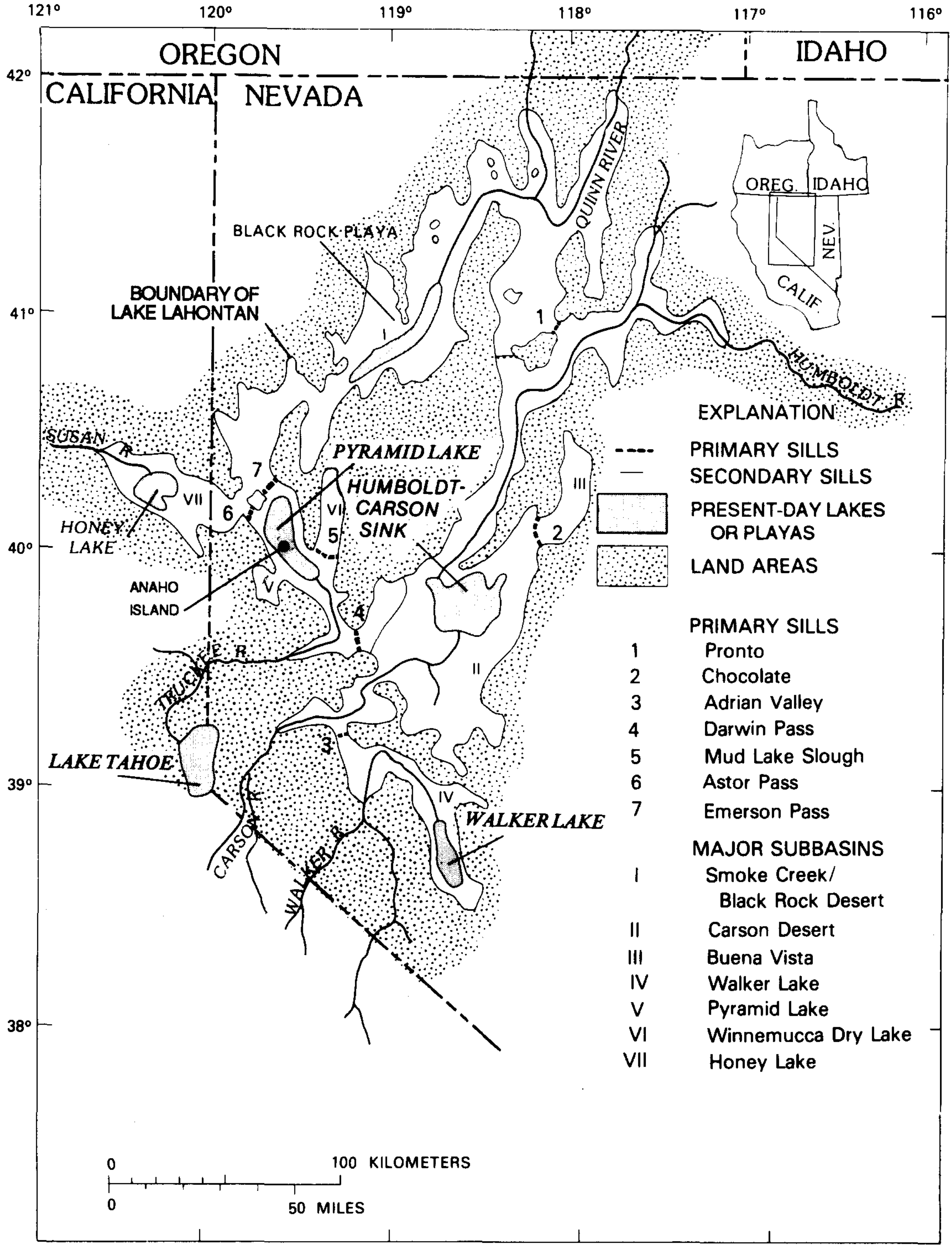

Fig.2. Rivers, subbasins, and sills of the Lahontan basin.

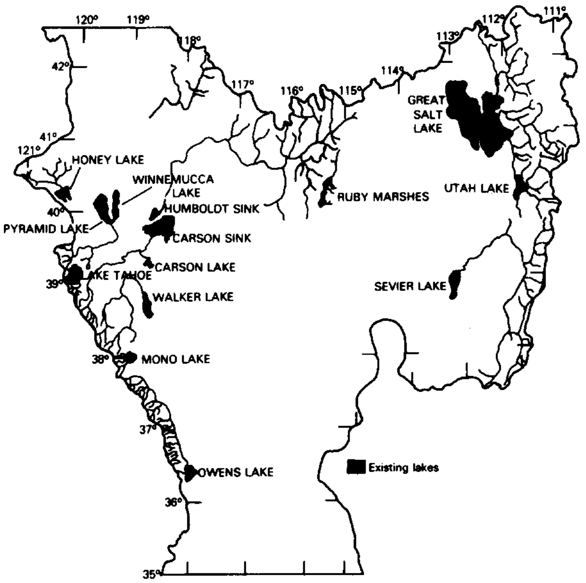

Fig.3. Drainage systems of the Great Basin.

material on which they form (Dorn et al., 1990). Therefore, the fall of Lake Lahontan may have occurred as early as 14,000 yr B.P. (which would indicate that the radiocarbon ages of samples I-10028, USGS-2168, I-10003, I-10026, I-10004, I-10000, I-9481, LDGO-1705A, LDGO-1750B, I-9344, I-9325, USGS-2169, and I-9992 in Table 1 are too young). Certain samples in the preceding set are considered reliable and indicate a fall in lake level as late as 13,260 $\pm$ 200 yr B.P. For example, sample I-9481 is a collection of unaltered aragonite gastropods (age = 13,260 $\pm$ 200 yr B.P.) found coating a lithoid tufa (sample I-9344, age = 13,430 $\pm$ 200 yr B.P.). The radiocarbon ages of these samples are in the right stratigraphic order and are similar in magnitude. These materials are characterized by differences in crystal structure, minor-element chemistry, and physical geometry (porosity and permeability). If contaminated, the samples would be expected to have yielded dissimilar and even reversed apparent radiocarbon ages. These materials might well date the time of the last highstand of Lake Lahontan, which indicates sufficient varnish for AMS age estimation accumulated in about 600 yr, a lag time consistent with many known varnish-age/bedrock-age pairs (Dorn et al., this issue).

The data in Table 1 have been used to develop a model chronology of lake-level fluctuation for the Pyramid Lake, Winnemucca Dry Lake, and Smoke Creek/Black Rock Desert subbasins (hereafter referred to as the western subbasins) (Fig.4). Two possible lake-level chronologies (solid and dashed lines in Fig.4) are depicted. These data have been supplemented with uranium-series age determinations on tufas from the Black Rock Desert (sample BR 84-1,2,3; age ~ 29,000 $\pm$ 9000 yr B.P.; altitude = 1214 m; Lao and Benson, 1988) and with a date on tephra from deltaic and alluvial sediments in the Black Rock Desert (age ~ 23,400 yr B.P.; altitude = 1256–1260 m; Davis, 1983).

Tufa-based age determinations for 50,000–20,000 yr B.P. are sparse. This sparseness may be primarily due to the absence of large lakes in the western subbasins during this time. Tufas and other carbonates that formed in shallow lakes would generally have been restricted to altitudes that were characterized by alluvial substrates. These substrates are very mobile in aquatic and exposed environments and do not lend themselves to preservation of carbonate coatings. In addition, evidence of lake sediments older than about 25,000 yr B.P. is lacking in the Smoke Creek/Black Rock Desert subbasins. The Trego Hot Springs tephra which has a radiocarbon age of 24,300 yr, crops out at an altitude of 1256 m at the base of the Upper Lacustral Clays described by Russell (1885) and at 1260 m in alluvial sediment. Based on these observations, small- or moderate-size lakes seem to have existed in the western subbasins between 49,000 $\pm$ 2000 yr B.P. and 29,000 $\pm$ 9000 yr B.P. (Lao and Benson, 1988)[1]. By about 24,000 yr B.P., lakes in the western subbasins had coalesced and risen to an altitude of at least 1260 m. Between 24,000 and 16,000 yr B.P., lake level may have remained between 1260 and 1270 m (Fig.4). Benson and Thompson (1987a) suggested that constancy of lake level at this altitude implied spill across the Darwin Pass Sill into the Carson Desert subbasin. Tufas in the altitude range of 1265–1300 m are the porous reeflike variety that are susceptible to contamination and, hence, have not been dated by the radiocarbon method (Benson, 1978).

[1]T.W. Stafford Jr. has recently obtained an AMS radiocarbon age of ~ 25,500 yr B.P. on protein extracted from a Camelops metacarpal found on the northwest shore of Pyramid Lake (altitude: 1162 m).

In the Astor Pass area, located at the northwestern end of the Pyramid Lake subbasin, two thick layers of *Chara* and diatoms are separated by a reddish soil. The radiocarbon age of organic material that was extracted from the soil is $15{,}660 \pm 150$ yr B.P. This oscillation in lake level is bracketed by radiocarbon ages of tufas from higher altitudes and also by radiocarbon ages of *Chara* located above and below the soil (Fig.4). The ages of *Chara* samples USGS-2171, 2173, and 2174 are not plotted because the degree of contamination of the *Chara* with younger carbon could not be evaluated. After $15{,}660 \pm 150$ yr B.P., Lake Lahontan rose to a highstand. Between $13{,}260 \pm 200$ and $12{,}680 \pm 100$ yr B.P., Lake Lahontan began to fall. This decline indicates that some tufas from altitudes > 1300 m are contaminated with young carbon or that lake level initially fell slowly.

Currey (1988) demonstrated that the final shallow-lake oscillation in the Carson Desert subbasin occurred about 11,000 yr B.P. Benson (1989) used Currey's (1988) data to demonstrate that the level of Pyramid Lake should have been stabilized by spill over Astor Pass Sill (1207 m) (Fig.2) during this time. In the Pyramid Lake subbasin, there is an erosional terrace on Anaho Island at an altitude of 1202–1220 m (Benson and Paillet, 1989). This terrace is free of tufa in contrast to the occurrence of large reef-like varieties of tufa above and below the terrace. The absence of tufa may indicate the presence of a moderate-sized lake in the Pyramid, Winnemucca Dry Lake, and Snake Creek/Blackrock Desert subbasins at about 11,000 yr B.P. Rock varnish from this terrace has a radiocarbon age of 9620 ± 95 yr B.P. (BETA 31065/ETH 5265). Varnish radiocarbon ages are about 5–13 % younger than the true radiocarbon ages of the exposed surface (see Fig.4 of Dorn et al., this issue). Therefore, the fall of Lake Lahontan from the 1202–1220 m terrace on Anaho Island is estimated to have occurred between 10,900 and 10,100 yr B.P.[1]

By 10,000 yr B.P. Pyramid Lake had fallen to the altitude of Mud Lake Slough Sill (1177 m), which connects Pyramid Lake and Winnemucca Dry Lake subbasins (Fig.2). A 6000-yr gap in the lake-level record occurs between 9000 and 3000 yr B.P. Benson and Thompson (1987a) have argued that Pyramid Lake could not have fallen below 1153 m during the Holocene. Their argument is based on the persistence through the Holocene of endemic fish (cui-ui and emerald trout) that were not able to spawn after the lake fell to the 1153-m level in 1967.

Problems associated with using radiocarbon ages of tufa samples that have been contaminated (Benson, 1978) or that have not been checked for contamination (Broecker and Kaufman, 1965) are indicated in Fig.5. The apparent shift in timing of the last highstand is due to the incorporation of secondary radiocarbon. The fall of lake level seems to occur at about 9500 yr B.P. instead of at about 13,000 yr B.P.

Lake Bonneville

The Bonneville basin consists of three main subbasins that are separated by sills of varying altitude (Fig.6). The Bear, Logan Ogden, Weber, Provo-Jordan, and Spanish Fork Rivers terminate in the Great Salt Lake subbasin and the Sevier and Beaver Rivers empty into the Sevier Lake subbasin. These rivers have their headwaters in mountains that form the eastern boundary of the Lake Bonneville basin (Figs.1 and 3). No major river empties into the Great Salt Lake Desert subbasin. Today, surface discharge constitutes $\sim 66\%$, precipitation $\sim 31\%$, and ground water $\sim 3\%$ of the average influx of water to Great Salt Lake, the major modern lake in the Bonneville basin (Arnow, 1984). Sevier Lake was dry from 1880 until 1982. In 1984 to 1985, the lake expanded to a 20th-century high as a result of abnormally high snow-melt runoff to the Sevier River (Oviatt, 1988).

The radiocarbon chronology of Lake Bonneville was first attempted by Broecker and Orr (1958) and Broecker and Kaufman (1965). As with their study of Lake Lahontan, carbonate samples were not examined for the presence of secondary carbon. Scott et al. (1983) attempted a revision of the chronology of the last lake cycle (Bonneville lake cycle) rejecting: (1) dates for tufa and marl, (2) dates in excess of 20,000 yr for shells, and (3) dates of shells that had ^{230}Th ages substantially greater than radiocarbon ages (Kaufman and Broecker,

[1] L.V. Benson and R.I. Dorn have recently obtained an AMS radiocarbon age of $10{,}460 \pm 80$ yr B.P. from subvarnish pockets of organic matter from boulders found at 1220 m on Anaho Island.

TABLE I

Radiocarbon ages, sample, and locality data for samples from the Pyramid Lake, Winnemucca Dry Lake, and Smoke Creek/Black Rock Desert subbasins of the Lahontan basin

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
Organic-samples deposited above lake level					
I-14059	*Neotoma* dung	1235	400 ± 100	1	Fishbone Cave No. 5
I-14060	*Atriplex*	1235	450 ± 120	1	Fishbone Cave No. 3B
I-14012	*Neotoma* dung	1235	1030 ± 90	1	Fishbone Cave
I-14015	*Atriplex* and *Neotoma* dung	1240	1910 ± 80	1	Crypt Cave No. 2
I-14061	Debris and *Neotoma* dung	1240	2100 ± 100	1	Crypt Cave No. 4
L-289II	Basketry	1240	2400 ± 200	4	Crypt Cave
I-14062	Debris and *Neotoma* dung	1230	2620 ± 110	1	Guano Cave No. 1
I-14063	Debris and *Neotoma* dung	1288	2950 ± 100	1	Kramer Cave No. 1
I-14064	Debris and *Neotoma* dung	1288	3070 ± 110	1	Kramer Cave GU
L-356B	Wood	1230	3200 ± 130	4	Guano Cave
I-14065	Debris and *Neotoma* dung	1288	3230 ± 120	1	Kramer Cave GM
L-364BI	*Neotoma* dung	1235	4150 ± 150	4	Fishbone Cave
L-289FF	Matting	1225	5970 ± 150	4	Cowbone Cave
L-596	Twigs	1250	6500 ± 150	5	Winnemucca Caves
L-289KK	Net	1235	7830 ± 350	4	Fishbone Cave
L-676B	*Neotoma* dung	1288	8300 ± 200	5	Kramer Cave
UCLA-672	Matting	1276	8380 ± 120	6	Shinners Site A
UCLA-675	Basketry	1276	9540 ± 120	6	Shinners Site A
L-676A	Plants and *Neotoma* dung	1288	10,500 ± 500	7	Kramer Cave
L-245	Roots and bark	1235	11,200 ± 250	4	Fishbone Cave
I-14009	*Neotoma* dung	1296	11,270 ± 170	1	Falcon Hill No. 1
A-3699	*Juniperus*	1230	11,580 ± 290	7	Guano Cave No. 11
I-14011	*Neotoma* dung	1296	11,770 ± 250	1	Falcon Hill No. 2
I-14016	*Artemisia*	1240	11,810 ± 230	1	Crypt Cave No. P3A
A-3696	*Juniperus*	1230	11,810 ± 230	7	Guano Cave No. 7B1
I-14014	Debris and *Neotoma* dung	1230	11,850 ± 170	1	Guano Cave No. 2
A-3695	*Juniperus*	1230	11,890 ± 250	7	Guano Cave No. 6A
A-3489	*Juniperus*	1296	12,020 ± 470	7	Falcon Hill No. 2
A-3698	*Juniperus*	1230	12,060 ± 260	7	Guano Cave No. 10
A-3697	*Juniperus*	1230	12,070 ± 210	7	Guano Cave No. 9
I-14013	*Juniperus* and *Neotoma* dung	1240	12,130 ± 180	1	Crypt Cave No. 84-2B
I-14008	*Juniperus* and *Neotoma* dung	1240	12,240 ± 180	1	Crypt Cave No. 1
AA-759	*Equus*	1235	12,280 ± 520	1	Fishbone Cave
I-14010	*Juniperus*	1240	12,350 ± 180	1	Crypt Cave No. 84-2A
PL 85-3S	Soil	1253	15,660 ± 150	1	Soil between marl units at Astor Pass
Organic-carbon samples deposited at or below lake level					
WIS-363	Wood	1174	670 ± 55	2	Truckee River delta
I-8195	Wood	1177	1025 ± 85	11	Truckee River delta

IS-364	Wood	1173	1110±55	2	Truckee River delta
WIS-378	Wood	1172	2270±55	2	Truckee River delta
WIS-375	Wood	1166	2690±65	2	Truckee River delta
WIS-361	Wood	1160	2710±60	2	Truckee River delta
WIS-376	Wood	1174	2890±50	2	Truckee River delta
WIS-374	Wood	1168	8800±90	2	Truckee River delta
WIS-377	Wood	1169	9720±100	2	Truckee River delta
I-8194	Wood	1177	9780±130	11	Truckee River delta
I-8193	Wood	1177	9970±140	11	Truckee River delta
Carbonate samples deposited below lake level; tufas thin-sectioned; gastropods X-rayed; all samples acid leached prior to analysis					
Unknown	Tufa/beachrock	1170	875±75	13	PL 40B
LDGO-1705C	Tufa	1165	10,350±130	12	PL 85-1; 1 m below land surface east of the Pyramid
I-10002	Tufa	1321	12,540±190	10	PL 103
I-10001	Tufa	1321	12,570±190	10	PL 102
I-9326	Tufa	1325	12,610±180	13	PL 21
I-10028	Tufa	1326	12,770±190	10	PL 113
USGS-2168	Tufa	1332	12,850±600	1	BR 85-2
I-10003	Tufa	1321	12,850±190	10	PL 104
I-10026	Tufa	1303	12,890±190	10	PL 112
I-10004	Tufa	1324	13,050±190	10	PL 105
I-10000	Tufa	1312	13,130±190	10	PL 101
I-9481	Gastropod shells	1311	13,260±200	13	PL 41G
LDGO-1705A	Tufa	1230	13,310±150	12	PL 87-2A; Sehoo Formation, Nixon Amphitheater
LDGO-1705B	Tufa	1230	13,430±170	12	PL 87-2B; Sehoo Formation, Nixon Amphitheater
I-9344	Tufa	1311	13,430±200	13	PL 41
I-9325	Tufa	1311	13,550±200	13	PL 20
USGS-2169	Tufa	1306	13,810±600	1	BR 85-1
I-9992	Tufa	1312	13,820±200	10	PL 100
USGS-4240157	Tufa	1270	14,090±190	1	BR 84–8
I-9331	Tufa	1230	15,140±250	13	PL 15
USGS-4240154	Tufa	1238	15,510±170	1	BR 84-5
I-9328	Tufa	1267	16,510±250	13	PL 18
USGS-4240156	Tufa	1254	16,900±270	1	BR 84-7
Unknown	Gastropod shells	1230	17,170±270	1	WDL 84-2G
I-10019	Tufa	1256	17,300±200	10	PL 110
USGS-4254144	Gastropod shells	1230	17,800±640	1	WDL 84-3G
USGS-4240155	Tufa	1245	18,030±470	1	BR 84-6
USGS-4254143	Gastropod shells	1230	18,030±300	1	WDL 84-4G
USGS-4240186	Tufa	1253	18,260±230	1	WDL 84-1
I-9329	Tufa	1260	18,580±310	13	PL 17
USGS-4240153	Tufa	1231	19,520±380	1	BR 84-4
I-9342	Tufa	1260	19,530±350	13	PL 23
I-9482	Gastropod shells	1260	19,620±360	13	PL 22G
I-9991	Tufa	1242	19,820±340	10	PL 109
I-10018	Tufa	1235	19,990±380	10	PL 108

TABLE I (*continued*)

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
Carbonate samples deposited below lake level; samples are contaminated or have not been thoroughly examined for contamination					
L-288F	Oolites	1177	1100±200	5	
L-766U	Shells	1158	1700±200	5	
L-766M	Shells	1164	1800±200	5	
L-288H	Shells	1173	2100±200	5	
L-289R	Shells	1161	3200±250	5	
L-364CE	Tufa	1186	8500±200	5	
L-364AA	Tufa	1329	9500±200	5	
L-356H	Tufa	1284	9700±200	5	
L-289G	Tufa	1318	9700±200	5	
L-356G	Tufa	1318	10,000±220	5	
Unknown	Tufa	1276	10,370±140	13	PL 11
Unknown	Tufa	1319	10,700±150	13	PL 42
L-437E	Tufa	1332	11,150±250	5	
L-289I	Tufa	1329	11,250±350	5	
L-437F	Tufa	1325	11,350±200	5	
Unknown	Tufa	1328	11,430±160	13	PL 43
L-289L	Tufa	1277	11,600±250	5	
L-289M	Tufa	1317	11,700±200	5	
L-289C	Tufa	1234	11,700±500	5	
L-289N	Tufa	1332	11,800±200	5	
L-364CF	Tufa	1186	12,150±150	5	
Unknown	Tufa	1302	12,270±180	13	PL 12
L-364CG	Tufa	1186	12,300±200	5	
Unknown	Tufa	1277	12,390±180	13	PL 19
Unknown	Tufa	1260	12,460±180	13	PL 24
Unknown	Tufa	1228	12,650±280	13	PL 6
L-289H	Tufa	1161	12,700±300	5	
L-364AM	Tufa	1222	12,700±350	5	
L-289S	Tufa	1220	12,900±350	5	
L-364DA	Tufa	1284	13,000±400	5	
Unknown	Tufa	1239	13,300±200	13	PL 8
Unknown	Tufa	1249	13,300±200	13	PL 9
Unknown	Tufa	1262	13,580±200	13	PL 10
L-364AN	Tufa	1222	13,700±300	5	
USGS-2171	*Chara*	1254	14,090±1600	1	PL 85-2C
L-364CI	Tufa	1186	14,500±400	5	
L-289D	Tufa	1234	14,800±500	5	
L-289P	Shells	1234	15,130±550	5	

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
L-364CK	Tufa	1186	15,350 ± 400	5	
L-364CJ	Tufa	1186	15,500 ± 350	5	
L-2890	Shells	1234	15,670 ± 700	5	
L-289K	Tufa	1244	16,130 ± 750	5	
L-264CQ	Shells	1210	16,500 ± 300	5	
Unknown	Tufa	1229	16,520 ± 310	13	PL 7
L-364CR	Marl	1225	16,800 ± 600	5	
L-772Q	Ostracode valves	1219	16,800 ± 500	5	
L-772G	Shells	1250	16,800 ± 500	5	
L-364CQ	*Chara*	1210	16,800 ± 300	5	
L-772F	Shells	1250	17,000 ± 300	5	
L-772A	*Chara*	1234	17,100 ± 400	5	
Unknown	Tufa	1260	17,100 ± 340	13	PL 44A
Unknown	Tufa	1155	17,170 ± 270	13	PL 38
Unknown	*Chara*	1256	17,250 ± 270	13	PL 4C
L-772L	Marl	1225	17,300 ± 400	5	
L-772B	Shells	1234	17,300 ± 400	5	
L-483P	Tufa	1207	17,400 ± 500	5	
L-364CS	Shells	1225	17,500 ± 600	5	
L-7720	Ostracode valves	1225	17,600 ± 400	5	
L-364AL	Marl	1219	17,600 ± 650	5	
L-772K	Shells	1225	17,900 ± 800	5	
L-772NB	*Chara*	1225	17,900 ± 600	5	
USGS-2173	*Chara*	1252	18,130 ± 800	1	PL 85-4C
L-772H	Shells	1250	18,400 ± 300	5	
L-364BR	Shells	1250	18,700 ± 700	5	
USGS-2174	*Chara*	1251	18,970 ± 1,000	1	PL 85-5C
L-772M	Marl	1225	19,000 ± 500	5	
L-772NA	Shells	1225	19,100 ± 700	5	
L-364BS	Ostracode valves	1250	19,750 ± 650	5	
Unknown	*Chara*	1257	17,810 ± 280	13	PL 5C
Unknown	*Chara*	1254	18,910 ± 340	13	PL 3C
Unknown	Tufa	1238	21,370 ± 420	13	PL 16
Unknown	Tufa	1203	22,090 ± 450	13	PL 13
L-289J	Tufa	1210	28,900 ± 1400	5	

Key to references: 1 = Benson and Thompson (1987a), 2 = Born (1972), 4 = Broecker and Orr (1958), 5 = Broecker and Kaufman (1965), 6 = Hattori (1982), 7 = Thompson et al. (1986), 10 = Benson (1981), 11 = Prokopovich (1983), 12 = This report, 13 = Benson (1978).

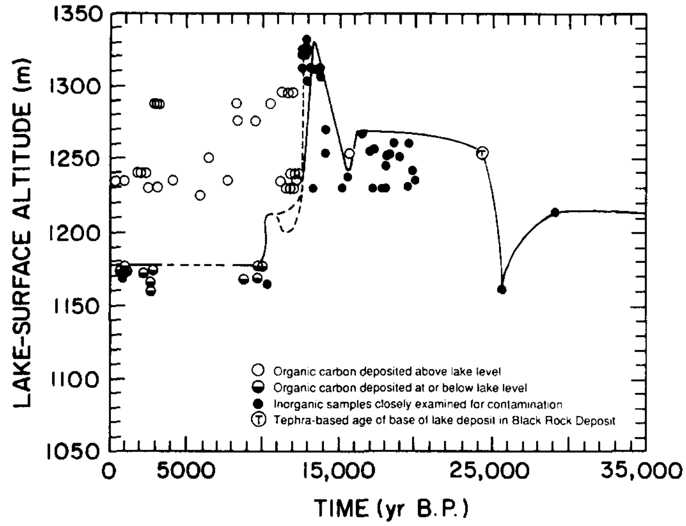

Fig.4. Model chronology for Lake Lahontan derived using data from the Pyramid Lake, Winnemucca Dry Lake, and Smoke Creek/Black Rock Desert subbasins. Dashed and solid lines between 25,000 and 10,000 yr B.P. indicate alternative chronologies. Dashed line between 8000 and 4000 yr B.P. indicate time for which little data are available.

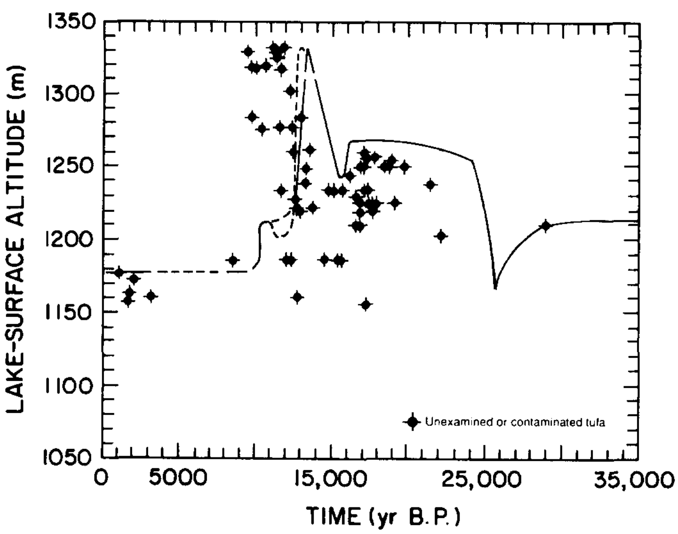

Fig.5. Lake-level data for Lake Lahontan derived by using samples that are contaminated and samples that were never checked for contamination.

1965). Scott et al. (1983) also reported new radiocarbon ages for wood and charcoal from lagoonal, shoreline, and alluvial deposits.

Currey and Oviatt (1985) made a synthesis of the Bonneville lake cycle that incorporated all radiocarbon-age data for wood and most of the compiled radiocarbon-age data for wood and plant material from Scott et al. (1983). Currey and Oviatt (1985) rejected radiocarbon-age data for shells, tufa, and certain organic samples for which the stratigraphic context or altitude of these samples was in question. Radiocarbon-age data from the literature for ostracode valves and gastropod shells deposited in stratigraphic association with the Pavant Butte Ash were selected and used to estimate the age of the ash. In addition,

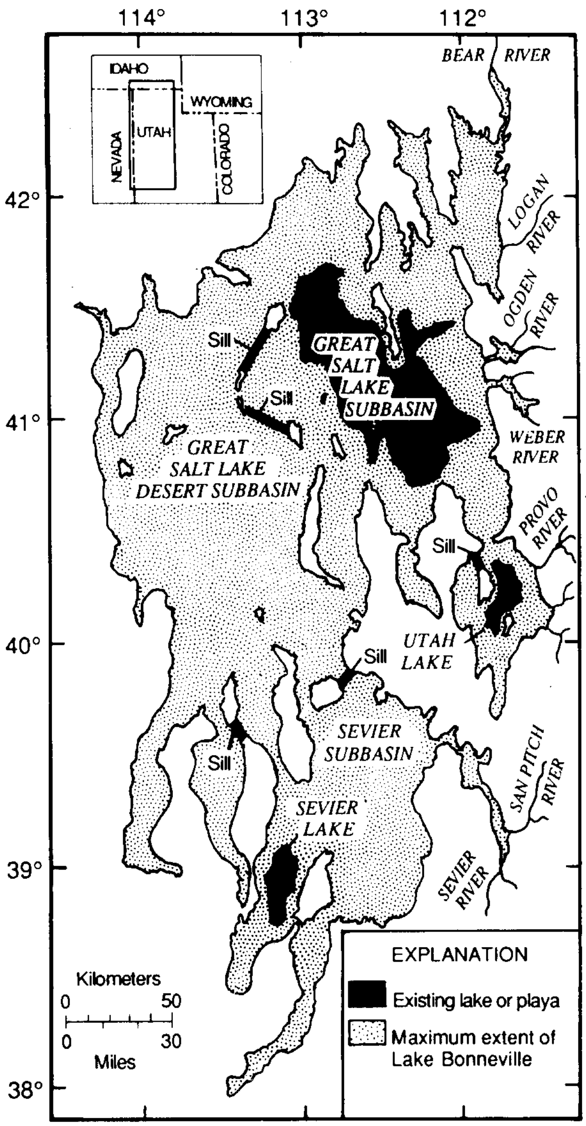

Fig.6. Major subbasins and sills in the Lake Bonneville basin (modified from Currey et al., 1984).

radiocarbon-age data for about 20 carbonate samples for which the stratigraphic context was known also were used in the synthesis of the Bonneville lake cycle (Table 2). Since 1985, new radiocarbon ages and stratigraphic analyses have permitted refinements to the Currey and Oviatt (1985) history of Lake Bonneville. These refinements are discussed below, but are not shown in Fig.7.

Radiocarbon ages of wood samples provide excellent documentation of change in the level of Lake Bonneville from 26,000 to 15,250 yr B.P. (a charcoal sample — BETA 23174/ETH 3518 — collected from just below the highest Lake Bonneville shoreline has a radiocarbon age of 15,250 ± 160 yr B.P.) (D. R. Currey and C. G. Oviatt, pers. comm., 1989) and between 13,150–10,300 yr B.P. (Fig.7). The radiocarbon ages of ostracodes, gastopods, and charcoal (Oviatt and Nash, 1989) which were deposited in association with the Pavant Butte Ash indicate a major drop in lake level (Keg Mountain Oscillation) (Currey et al., 1984) after about 15,250 yr B.P., but before the Bonneville Flood that occurred about 14,500 yr B.P. This basaltic ash crops out in the Sevier Desert and was deposited during a transgressive phase when the lake was 15 m below its highest level. Fluctuations in lake level having a maximum amplitude of about 45 m occurred between about 23,000 and 21,000 yr B.P. and are collectively known as the Stansbury Oscillation.

Dorn et al. (1990) have obtained a radiocarbon age of 14,050 ± 130 yr B.P. from rock varnish that coats boulders that were deposited from the Bonneville Flood that downcut the outlet at Zenda by 108 m, forming a new pass (Red Rock Pass) (Malde, 1968). This age is consistent with other radiocarbon data that indicate the flood occurred about 14,500 yr B.P. (Currey and Burr, 1988). After about 14,200 yr B.P., climatic forcing caused the lake to regress from the Provo shoreline (Fig.7). The apparent time of initial regression is constrained mainly by a single radiocarbon age of 13,900 ± 400 yr B.P. on mollusc shells (sample W-899) from the youngest part of the post-flood paleodelta of the Bear River (Table 2). The possibility of contamination by secondary carbon and the lack of ^{13}C adjustment indicate that the reported age of sample W-899 is a minimum for the start of the regression. However, sample W-899 also is composed of shells that may have been reworked from slightly older deposits within the post-flood fluviodeltaic system.

Between about 13,000 and 12,000 yr B.P., Great Salt Lake was at very low levels as indicated by extensive pre-Gilbert red beds that contain desiccation cracks and desert pavement (Currey et al., 1988a) and by deposition of as much as 13 m of mirabilite ($Na_2SO_4 \cdot 10H_2O$). Slightly after 11,000

TABLE II

Radiocarbon ages, sample and locality data for samples from the Lake Bonneville basin

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
Organic-carbon samples deposited above lake level					
GX-4737	Soil	1550	26,080±1,150	8,27	Predates Bells Canyon glacier and last deep lake
W-1338	Plant fragments	1442	12,090±300	9	Postdates Provo overflow at Swan Lake
C-609	Ovis dung	1314	11,450±600	10,26	Postdates last inundation of Danger Cave
BETA-12988	Disseminated organic carbon	1387	7930±110	24	Alluvial mud above lake level Sevier subbasin
BETA-8011	Disseminated organic carbon	1386	5930±220	23	Alluvial mud above lake level Sevier subbasin
Organic-carbon samples deposited at or below lake level					
W-4893	Wood	1314	26,000±600	1	Transgressive delta
W-4898	Wood	1365	22,500±300	1	Transgressive gravel-bar complex
I-698	Wood	1357	22,000±600	2	Transgressive delta (?)
W-4897	Wood	1417	20,900±250	1	Transgressive lagoon/bar complex
L-775N	Wood	1442	20,800±300	3	Transgressive shore zone
W-876	Wood	1442	20,600±500	4	Transgressive shore zone
SI-4124	Wood	1438	20,500±200	5	Transgressive lagoon/bar complex
W-941	Wood	1442	20,300±300	6,3	Transgressive shore zone
W-4421	Wood	1445	19,700±200	1	Transgressive lagoon/bar complex
W-4445	Wood	1442	19,580±280	1	Transgressive lagoon/bar complex
SI-4041C	Wood	1487	18,980±160	5	Transgressive lagoon/bar complex
W-4695	Wood	1463	18,800±180	1	Transgressive lagoon/bar complex
W-4693	Wood	1494	18,600±150	1	Transgressive shore zone
W-4687	Wood	1524	18,000±150	1	Transgressive shore zone
W-4451	Charcoal	1535	17,580±170	1	Transgressive shore zone
W-4896	Wood	1553	16,770±200	1	Transgressive shore zone
I-697	Wood	1367	13,150±270	2	Post-Provo regressive delta
W-1824	Wood	1313	12,290±350	7	Regressive (?) sequence
I-696	Wood	1290	10,300±280	2	Overlain by sand of Gilbert transgression
Carbonate samples deposited above lake level; samples have not been thoroughly examined for contamination					
BETA-8344	Soil carbonate	1455	22,500±370	11	Predates transgression of last deep lake
BETA-12987	Gastropod shells	1387	9570±430	24	Alluvial mud above lake level Sevier subbasin
BETA-17878	Gastropod shells	1390	9340±160	25	Alluvial mud above lake level Sevier subbasin
BETA-17882	Gastropod shells	1385	2560±80	25	Alluvial mud underlying Sevier River terrace Sevier subbasin

Carbonate samples deposited below lake levels; samples have not been thoroughly examined for contamination					
BETA-11489	Oolitic sand and carbonate mud	1232	31,660 ± 740	12	Shallow saline lake
BETA-13997	Clean oolitic sand	1299	27,180 ± 430	12	Transgressive beach
I-4409	Oolitic marl	1292	26,700 ± 900	13	Marsh inundated by shallow lake
BETA-8343	Marl	1356	24,870 ± 410	12	Transgressive offshore zone
BETA-5038	*Sphaerium* shells	1378	23,190 ± 1,360	14	Transgressive nearshore zone
BETA-5566	*Amnicola* shells	1362	20,710 ± 310	15	Nearshore zone
W-982	*Stagnicola* shells	1478	18,980 ± 500	6,28	Transgressive nearshore zone (?)
L-711C	Gastropod shells	1507	17,500 ± 400	16	Transgressive nearshore zone
L-363G	Tufa	1585	16,100 ± 350	17	Near Bonneville shoreline
L-483BB	Tufa	1585	16,050 ± 300	18	Near Bonneville shoreline
L-711B	Ostracode valves	1427	15,600 ± 400	16	Below Pavant Butte Ash
L-774H	Gastropod shells	1480	15,500 ± 500	16	Above Pavant Butte Ash
L-774N	Gastropod shells	1525	15,400 ± 300	16	Brackets Pavant Butte Ash
BETA-10389	Ostracode valves	1420	15,220 ± 140	12	Base of Pavant Butte Ash
L-774F	Gastropod shells	1433	15,000 ± 300	16	Within Pavant Butte Ash
SI-4277C	Tufa	1579	14,730 ± 100	5	Youngest occupation of Bonneville shoreline
W-899	Shells	1433	13,900 ± 400	19,28	Bear River paleodelta
W-491	Tufa	1494	13,380 ± 400	20	Provo shoreline
W-2000	Shells	1424	12,860 ± 400	21,29	Regressive nearshore zone (?)
W-943	Gastropod shells	1440	12,780 ± 350	6	Regressive nearshore zone (?)
BETA-8348	Anodonta shells	1396	12,490 ± 130	30	Regressive nearshore zone Sevier subbasin
L-774Q	Ostracode valves	1382	11,900 ± 300	16	Regressive nearshore zone
BETA – 17883	Anodonta shells	1385	11,270 ± 110	25	Lacustrine mud at distal end of gravel spit Sevier subbasin
W-4395	Gastropod shells	1289	10,920 ± 150	22	Pre-Gilbert shoreline marsh
GX-6776	Anodonta shells	1385	10,360 ± 220	8	Lacustrine mud interfingering with spit Sevier subbasin
GX-6949	Gastropod shells	1284	10,300 ± 310	15	Lagoonal marsh at Gilbert shoreline
GX-6614	Gastropod shells	1294	10,280 ± 260	15	Lagoonal marsh at Gilbert shoreline
BETA-19455	Anodonta shells	1385	10,070 ± 130	25	Lacustrine mud Sevier subbasin

Key to references: 1 = Scott et al. (1983), 2 = Trautman and Willis (1966), 3 = Morrison (1965), 4 = Rubin and Alexander (1960), 5 = R. Stuckenrath, pers.comm. (1979), 6 = Ives et al. (1964), 7 = Ives et al. (1967), 8 = Currey (1980), 9 = Bright (1966), 10 = Libby (1955), 11 = Currey et al. (1984), 12 = Currey and Oviatt (1985), 13 = Mehringer (1977), 14 = Oviatt (1984), 15 = Currey et al. (1983), 16 = Broecker and Kaufman (1965), 17 = Broecker et al. (1957), 18 = Olson and Broecker (1961), 19 = Rubin and Berthold (1961), 20 = Rubin and Alexander (1958), 21 = Smith et al. (1968), 22 = Miller (1980), 23 = Simms and Isgreen (1984), 24 = Simms (1985), 25 = Oviatt (1988), 26 = Jennings (1957), 27 = Madsen and Currey (1979), 28 = Bright (1963), 29 = Marsters et al. (1969), 30 = Isgreen et al. (1984).

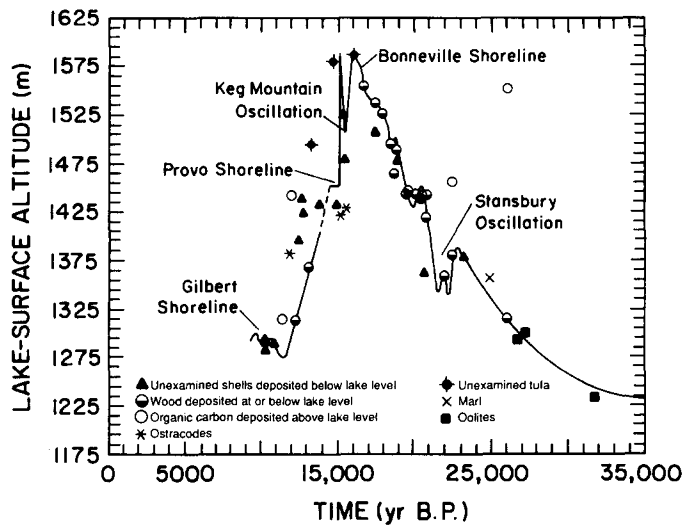

Fig.7. Model chronologies for Lake Bonneville (modified from Currey and Oviatt, 1985).

yr B.P., the lake had expanded to the Gilbert shoreline (1295 m). A subsequent regression had begun about 10,500 yr B.P. (Currey and Oviatt, 1985). Between 13,000 and 11,000 yr B.P., the lake in the Sevier subbasin (Lake Gunnison) apparently did not recede (Fig.8), but instead continued to spill across the Old River Bed threshold/sill (1400 m) to the Great Salt Lake subbasin until about 10,000 yr B.P. (Oviatt, 1988).

Studies of cored sediment from the Great Salt Lake subbasin indicate that, during much of Holocene time the lake in that subbasin stood at altitudes almost equal to the historic average of 1280 m. The Holocene highstands of 1287 m (Currey et al., 1988b) may have occurred about 3500 or 2300 yr B.P. (Spencer et al., 1984). During the early and middle Holocene, the lake in the Sevier subbasin (Sevier Lake) remained below an altitude of 1381 m (Oviatt, 1988). During the late Holocene, Sevier Lake expanded and contracted a number of times; one expansion deposited a prominent beach ridge with a crestal altitude of 1382 m. Oviatt (1988) has correlated this beach ridge with a terrace that is underlain by fine-grained alluvium. Gastropods collected from fine-grained fill below the terrace yielded a radiocarbon age of 2560±80 yr B.P. (sample BETA-17882, Table 2). Oviatt (1988) interpreted this date as the time of deposition of the highest late Holocene beach in the Sevier Lake subbasin (Fig.8).

Lake Searles

During the late Wisconsin, Searles Lake was third in a chain of lakes that received water from the Owens River (Fig.9). When Owens Lake filled to a depth of about 60 m (altitude = 1146 m), it overflowed into the China Lake basin. Whien China Lake filled to about 12 m (altitude = 666 m), it overflowed into the Searles Lake basin. When lake levels in the two adjoining basins reached 678 m, the two lakes coalesced (Lake Searles). Lake Searles overflowed to the Panamint Lake basin when it reached a depth of about 200 m (altitude = 695 m).

The pluvial lake in Searles, Panamint, and Death Valleys were first indicated on plate 1 of Russell (1885). Bailey (1902) was the first to correctly portray the hydrologic relation between these basins. The identification of Searles Lake in 1912 as a potential domestic supply of potash that would be needed if European sources were threatened by the outbreak of war stimulated a series of geologic and engineering studies of Searles Lake. Of these studies, Gale's (1914) was the most complete. It contained data about the Owens,

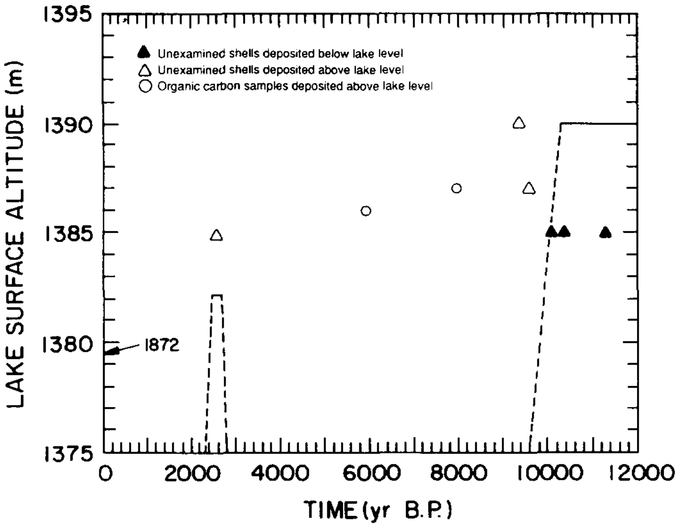

Fig.8. Holocene chronology for Lake Gunnison and Sevier Lake (modified from Oviatt, 1988).

China, and Panamint basins, including a systematic record of shorelines in the basins and the relation of the shorelines to the altitudes of the sills that separated the basins.

Flint and Gale (1958) were the first to attempt a radiocarbon-based chronology of the upper four major stratigraphic units observed in cores obtained from the Searles Lake basin. These informal units from the top downward are: (1) Upper Salt, (2) Parting Mud, (3) Lower Salt, and (4) Bottom Mud. Flint and Gale (1958) reported that the Bottom Mud/Lower Salt and Parting Mud/Upper Salt pairs represented relatively deep lakes that subsequently evaporated to become shallow, very saline lakes or perhaps to complete dryness. The Parting Mud contains fossil fishes, pollen, detrital clay, and disseminated organic matter. The mud itself was found later to consist primarily of calcite ($CaCO_3$), aragonite ($CaCO_3$), dolomite [$CaMg(CO_3)_2$], and halite (NaCl). Disseminated crystals of gaylussite ($CaCO_3 \cdot Na_2CO_3 \cdot 5H_2O$) and pirssonite ($CaCO_3 \cdot Na_2CO_3 \cdot 2H_2O$) were concentrated near the upper and lower contacts with the bordering salt units. The rarity of these moderately soluble minerals in the middle part of the Parting Mud indicated to Flint and Gale (1958) that "... net evaporation reached a minimum around a midpoint in the history of the freshwater lake". However, Smith and Haines (1964) and Smith (1979) believed that the minerals gaylussite and pirssonite were virtually all secondary in origin. Hand specimens and thin sections of the laminated muds revealed penetration textures that indicated the megascopic fraction of the gaylussite and pirssonite crystals grew after sediment compaction. Large crystals cut across bedding planes; where the crystals cut aragonite laminae, the laminae remain undistorted. Smith and Haines (1964) and Smith (1979) hypothesized that crystal growth was a diagenetic process that resulted from diffusion of a sodium-laden brine through the mud layers, where it reacted with microcrystalline calcium carbonate to form gaylussite and pirssonite.

Flint and Gale (1958) believed that the clay fraction of the Parting Mud was transported to the Searles Lake basin by spill from the Owens River; therefore, the top of the Parting Mud represented the time of cessation of overflow from China Lake. The subsequent history of the lake was one of continued evaporation and precipitation of evaporites. The origin of thin layers of mud within the Upper Salt was not well understood. The Bottom Mud/Lower Salt pair were considered similar to the Parting Mud/Upper Salt pair in many respects; however, the Lower Salt contained more numerous and thicker partings of mud than did the Upper Salt.

Radiocarbon-age dating of extractable organics

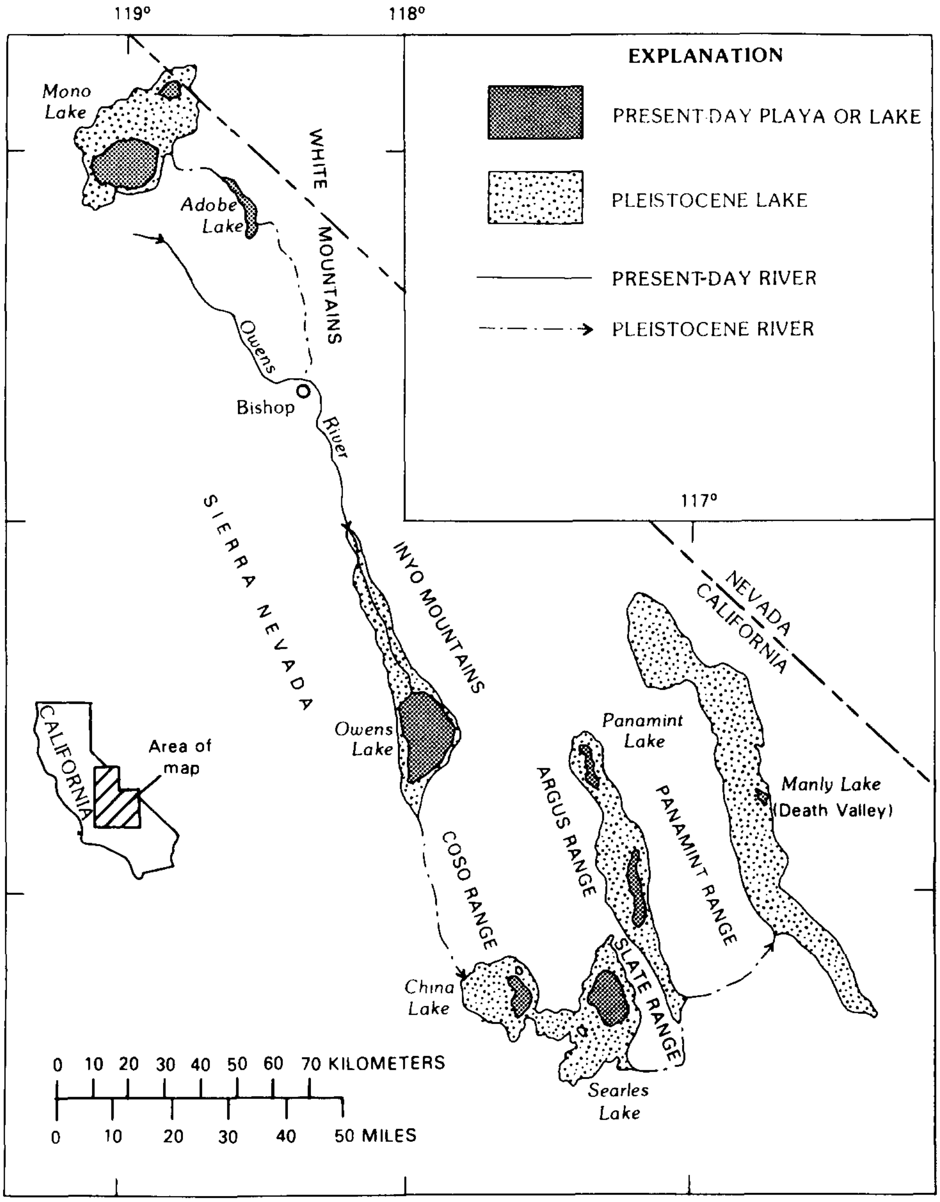

Fig.9. Lakes in the Owens River system (modified from Smith, 1979).

from the Parting Mud generally were stratigraphically consistent and indicated that inflow of water to the Searles Lake basin began about 23,000 yr B.P. and ceased about 10,000 yr B.P. (Flint and Gale, 1958). Radiocarbon-age determinations on samples of organic carbon and carbonate from the Upper and Lower Salts yielded less consistent results.

Stuiver (1964) continued the work of Flint and Gale (1958) and presented a more precise and detailed age analysis of samples from Lake Searles (Table 3). Although the Overburden Mud, the mud layer that overlies all parts of the Upper Salt, was not studied by Flint and Gale (1958), Stuiver (1964) briefly mentioned it, stating that it probably was formed by sediment washed in from the sides of the basin. Stuiver (1964) also concluded that the radiocarbon age of 3520-yr-old fossil twig found at a depth of 2.4 m in the Overburden Mud had no apparent connection with the desiccation of the lake.

Smith's (1979) study of the Overburden Mud indicated that it actually covers the entire surface of the present-day Searles Lake playa, extending a kilometer or more beyond the upper surface of the Upper Salt. The basal contact of the Overburden Mud is gradational and the composition of the Overburden Mud is quite different from the other mud units that were deposited in the basin. Clastic fragments are coarser and compose a larger percentage of this unit. The Overburden Mud unit consists entirely of fine-grained mud at the edge of the basin but grades basinward into interbedded layers of salt and fine-grained mud. Unlike the Parting Mud, laminae do not occur in the Overburden Mud. Aragonite and calcite are not present, although small quantities of dolomite have been detected. Smith (1979) correlated the Overburden Mud with an outcropping unit that can be mapped to levels up to 55 m above the basin floor. Stuiver and Smith (1979) also concluded that the 3520-yr-old twig could be used to estimate the age of the Overburden Mud.

Stuiver (1964) demonstrated that the total quantity of salts contained in the Upper Salt is sufficient to saturate a body of water three times the volume of the Upper Salt. This volume of water indicates that precipitation of the Upper Salt began when the surface of Lake Searles was about 50 m above the former basin floor and implies that deposition of the Parting Mud occurred in a lake deeper than 50 m. Because the annual evaporation rate in the Searles Lake basin would have been about 1–2 m, the Upper Salt probably was desposited in less than 100 yr if inflow ceased.

Radiocarbon ages of samples of organic carbon extracted from mud layers in the Upper Salt are essentially the same between 4.5 m and the bottom of the unit (18.1 m) (Table 3). Stuiver (1964) concluded that this similarity in ages indicated that at least 80% of the Upper Salt was rapidly deposited in agreement with his previous calculation. Stuiver (1964) indicated that if overflow from Owens Lake had produced the mud partings in the Upper Salt, the age of the material should have been less than the 9850 yr B.P. age of the lower section of the Upper Salt (Table 3, samples Y-1206 and Y-1207). However, the ages of the disseminated organic carbon samples (samples Y-1202B and Y-1204B) are somewhat older, indicating the partings from which the samples came consisted of reworked and redeposited older sediment.

Stuiver's (1964) general description of formation of the Upper Salt was as follows: (1) precipitation of the Upper Salt began soon after 10,200 yr B.P. (after cessation of inflow from the Owens Lake system), (2) desiccation of the lake was very rapid, and four-fifths of the salt was deposited within about 100 yr of the onset of salt precipitation, and (3) between 9700 and 6800 yr B.P., and possibly for some time thereafter, only shallow, intermittent lakes existed as the result of local runoff.

Radiocarbon ages of organic samples that were extracted from the Parting Mud (Stuiver, 1964) agree with the results of Flint and Gale (1958), which indicate that deposition of the Parting Mud occurred between about 24,200 and 10,200 yr B.P. (Table 3). Stuiver (1964) reported that radiocarbon ages of disseminated organic-carbon samples (samples Y-1208B-1 and Y-1208B-2, Table 3) extracted from the upper 0.8% of the Parting Mud had older ages (10,900±90 and 10,680±90 yr B.P.) than the sample from the upper 0.8 to 1.4% of the unit. These ages indicate that reworking of older sediment had begun at least by 10,200 yr B.P.

Radiocarbon ages of paired organic carbon and secondary carbonate samples from the Parting Mud usually were different; the carbonate sample was younger than the organic sample in six out of seven sample pairs taken from depths in excess of 7 m (Table 3). In Table 3, samples from depths less than 7 m are actually from the Overburden Mud, a unit not recognized by Stuiver (1964). The only sample pair that had an older carbonate age was the uppermost sample pair from the Parting Mud (samples Y-1211, and Y-1211B, Table 3). Stuiver (1964) suggested the older carbonate age indicated that older carbon had been diagenetically added or exchanged from lower in the section. Stuiver (1964) reported that exposure of the Parting Mud to the atmosphere for a period of three weeks resulted in an exchange (or addition) of 5.6% modern carbon. Stuiver (1964) hypothesized that contamination with younger carbon (either from the laboratory

TABLE III

Radiocarbon ages, sample, and locality data for samples from Lake Searles (Samples arranged by decreasing altitude within each stratigraphic unit)

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
Organic-carbon samples deposited below lake level in outcropping sediments of the Searles Lake Formation					
W-1418	Disseminated organic carbon (tufa)	510	11,720 ± 500	1	Unit C
Y-unknown	Disseminated organic carbon (marl)	561	13,750 ± 160	1	Unit C
Carbonate samples deposited below lake level					
Y-unknown	Molluscs shells	515	16,200 ± 300	1	Unit CD, sample contaminated or reworked
W-1317	Tufa	625	12,000 ± 400	1	Unit C
W-1327	Oolites	619	11,730 ± 350	1	Unit C
W-1325	Tufa	597	12,110 ± 300	1	Unit C
USGS-2328A	Tufa	591	12,700 ± 50	1	Unit C
USGS-2328B	Tufa	591	12,910 ± 50	1	Unit C
W-1318	Tufa	510	12,200 ± 450	1	Unit C
USGS-33A	Tufa	585	14,300 ± 200	1	Unit C
USGS-33B	Tufa	585	12,800 ± 150	1	Unit C, duplicate of USGS-33A
Y-unknown	Mollusc shells	572	11,700 ± 160	1	Unit C
W-1679	Oolites	570	11,020 ± 400	1	Unit C
W-1680	Oolites	570	11,820 ± 400	1	Unit C, acid-leached duplicate of W-1679
Y-unknown	Mollusc shells	565	14,210 ± 200	1	Unit C
W-1893	Tufa	564	13,830 ± 500	1	Unit C
W-1890	Tufa	564	13,650 ± 500	1	Unit C, duplicate of W-1893
Y-unknown	Tufa	562	10,430 ± 100	1	Unit C
Y-unknown	Tufa	561	14,140 ± 120	1	Unit C
W-1201	Tufa	536	13,300 ± 500	1	Unit C
W-1323	Tufa	533	13,700 ± 500	1	Unit C
Y-unknown	Oolites	512	13,200 ± 200	1	Unit C
Y-unknown	Oolites	509	13,350 ± 200	1	Unit C
W-1894	Oolites	643	9070 ± 300	1	Unit B, sample contaminated or reworked
USGS-67	Marl	643	7700 ± 75	1	Unit B, same bed as W-1894, sample contaminated
USGS-2520	Tufa	616	13,520 ± 120	1	Unit B
USGS-2424	Marl	582	15,020 ± 80	1	Unit B
W-1904	Oolites	564	14,950 ± 500	1	Unit B
USGS-2321	Marl	561	20,210 ± 90	1	Unit B
Y-unknown	Oolites	561	16,970 ± 400	1	Unit B
Y-unknown	Oolites	558	16,820 ± 200	1	Unit B
USGS-2425	Marl	558	21,020 ± 200	1	Unit B
W-1324	Tufa	552	22,500 ± 600	1	Unit B
USGS-2102	Marl	543	30,510 ± 210	1	Unit B, sample contaminated or reworked
USGS-2423	Marl	543	26,140 ± 140	1	Unit B
USGS-2327(?)	Mollusc shells	533	18,600 ± 130	1	Unit B
USGS-2327(?)	Marl	533	18,560 ± 210	1	Unit B

USGS-2326	Marl	544	32,960±420	1	Unit B, sample contaminated or reworked
USGS-2329B	Marl	533	21,330±80	1	Unit B
W-1905	Marl	501	19,150±600	1	Unit B
USGS-2106	Marl	555	29,960±250	1	Unit AB6, sample contaminated or reworked
Y-unknown	Oolites	538	32,100±1000	1	Unit AB6, sample contaminated or reworked
USGS-2104	Marl	536	25,740±130	1	Unit AB6
USGS-2327(?)	Marl	546	18,560±210	1	Unit AB4, sample contaminated or reworked
Y-unknown	Tufa	530	28,800±400	1	Unit AB4
Y-unknown	Oolites	530	27,800±1200	1	Unit AB4
W-1922	Mollusc shells	524	27,400±800	1	Unit AB4
USGS-2105	Mollusc shells	558	26,650±860	1	Unit AB3
USGS-2426	Marl	546	29,130±160	1	Unit AB3
USGS-2322	Marl	543	27,880±210	1	Unit AB3
W-1575	Mollusc shells	529	29,200±2000	1	Unit AB3
Y-unknown	Mollusc shells	529	35,000±1600	1	Unit AB3, sample contaminated or reworked
USGS-2325	Marl	631	10,610±50	1	Unit AB2, sample contaminated or reworked
W-1322	Tufa	671	10,230±2200	1	Unit A, sample contaminated or reworked
W-1321	Tufa	552	32,500±2000	1	Unit A, sample comtaminated or reworked
Y-unknown	Mollusc shells	552	40,000±2500	1	Unit A
GX-11935	Tufa	549	>36,000	1	Unit A
USGS-2324B	Tufa	549	33,370±330	1	Unit A, sample contaminated or reworked
USGS-2323	Mollusc shells	543	30,900±260	1	Unit A, sample contaminated or reworked
		Core depth (m)			
Organic-carbon samples from cored stratigraphic units; relative positions of samples from parting mud are approximate; radiocarbon ages of mud partings in the Upper Salt					
Y-1200B	Disseminated organic carbon	1.60–1.65	6630±390	2	Testhole L-U-1
Y-1048	Twig	2.40	3520±190	2	Unknown testhole
W-892	Disseminated organic carbon	6.71–6.92	12,390±400	3	Corehole S-35
W-942	Disseminated organic carbon	6.71–6.92	11,800±1000	4	Corehole S-35, duplicate of W-892
Y-1202B	Disseminated organic carbon	9.45–9.48	11,010±150	2	Testhole L-U-1
Y-1204B	Disseminated organic carbon	12.80–12.83	11,510±150	2	Testhole L-U-1
		Core depth (relative %)			
Radiocarbon ages of the parting mud; depth in relative position (in %) between top and bottom of core					
Y-1208B-1	Disseminated organic carbon	0.0–0.4	10,680±90	2	Testhole L-U-1
Y-1208B-2	Disseminated organic carbon	0.4–0.8	10,900±90	2	Testhole L-U-1
Y-1209B	Disseminated organic carbon	0.8–0.4	10,230±80	2	Testhole L-U-1
Y-1210B-1	Disseminated organic carbon	0.0–2.7	10,060±90	2	Testhole X-23
Y-1210B-2	Disseminated organic carbon	0.0–2.7	10,410±120	2	Testhole X-23
Y-1211B	Disseminated organic carbon	6.2–6.9	10,590±110	2	Testhole X-23
Y-1212B	Disseminated organic carbon	12.8–14.9	11,800±130	2	Testhole X-20
Y-1213B	Disseminated organic carbon	27.7–29.8	13,710±270	2	Testhole X-20

TABLE III (*continued*)

Radiocarbon ages, sample, and locality data for samples from Lake Searles (Samples arranged bydecreasing altitude within each stratigraphic unit)

		Core depth (relative %)			
Y-1214B	Disseminated organic carbon	45.2–45.7	18,800 ± 240	2	Testhole X-23
Y-1215B	Disseminated organic carbon	61.0–62.4	19,970 ± 280	2	Testhole X-20
Y-1216B	Disseminated organic carbon	75.3–75.6	22,300 ± 280	2	Testhole X-23
Y-1217B	Disseminated organic carbon	95.9–97.9	24,630 ± 460	2	Testhole X-23
Y-1218B	Disseminated organic carbon	99.3–100	23,710 ± 320	2	Testhole L-U-1

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
Radiocarbon ages of mud layers in the lower salt					
Y-2230	Disseminated organic carbon		23,750 ± 300	5	Testhole L-31, top 0.05 m of M-7
Y-2231	Disseminated organic carbon		26,350 ± 350	5	Testhole L-31, bottom 0.05 m of M-7
Y-2232	Disseminated organic carbon		24,760 ± 300	5	Testhole L-31, Top 0.05 m of M-6
Y-2233	Disseminated organic carbon		28,880 ± 500	5	Testhole L-31, bottom 0.05 m of M-6
Y-2234	Disseminated organic carbon		27,550 ± 400	5	Testhole L-31, bottom 0.05 m of S-5
Y-2235	Disseminated organic carbon		28,380 ± 350	5	Testhole L-31, top 0.05 m of M-5
Y-2236	Disseminated organic carbon		29,040 ± 350	5	Testhole L-31, bottom 0.05 m of M-5
Y-2237	Disseminated organic carbon		28,620 ± 350	5	Testhole L-31, top 0.05 m of M-4
Y-2238	Disseminated organic carbon		30,160 ± 400	5	Testhole L-31, bottom 0.05 m of M-4
Y-2239	Disseminated organic carbon		30,510 ± 400	5	Testhole L-31, top 0.05 m of M-3
Y-2240	Disseminated organic carbon		30,270 ± 500	5	Testhole L-31, bottom 0.05 m of M-3
Y-2241	Disseminated organic carbon		30,280 ± 300	5	Testhole L-31, top 0.05 m of M-2
Y-2242	Disseminated organic carbon		32,620 ± 500	5	Testhole L-31, bottom 0.05 m of M-2
Y-340	Twig		26,700 ± 2000	6	Testhole GS-27, ~1 m above base of Lower Salt
Y-343	Disseminated organic carbon		29,500 ± 2000	6	Testhole GS-27, ~1 m above base of Lower Salt
GRO-1460	Disseminated organic carbon		32,300 ± 400	2	Testhole X-20, ~1.15 m above base of Lower Salt
Radiocarbon ages of top of bottom mud					
Y-1224B	Disseminated organic carbon		32,700 ± 800	2	Testhole X-23, top 0.05 m of Bottom Mud
Y-2243	Disseminated organic carbon		32,800 ± 600	5	Testhole L-31, top 0.05 m of Bottom Mud

		Core depth (m)			
Carbonate samples from cored stratigraphic units; relative positions of samples for parting mud are approximate; radiocarbon ages of mud partings in the Upper Salt					
Y-1200	Carbonates	1.60–1.65	6890 ± 140	2	Testhole L-U-1, paired with Y-1200B
Y-1201	Carbonates	4.57–4.62	9700 ± 180	2	Testhole L-U-1
Y-1202	Carbonates	9.45–9.48	11,100 ± 180	2	Testhole L-U-1, paired with Y-1202B
Y-1203	Carbonates	11.43–11.45	10,600 ± 100	2	Testhole L-U-1
Y-1204	Carbonates	12.80–12.83	10,460 ± 170	2	Testhole L-U-1, paired with Y-1204B
Y-1205	Carbonates	15.87–15.95	9720 ± 200	2	Testhole L-U-1
Y-1206	Carbonates	18.59–18.61	9850 ± 180	2	Testhole L-U-1
Y-1207	Carbonates	19.68–19.71	9840 ± 80	2	Testhole L-U-1
		Core depth (relative %)			
Radiocarbon ages of the parting mud; depth in relative position (in %) between top and bottom of core					
Y-1209	Carbonates	0.8–1.4	10.630 ± 180	2	Testhole L-U-1, paired with Y-1209B
Y-1210	Carbonates	0.0–2.7	11,470 ± 100	2	Testhole X-23, paired with Y-1210B-1
Y-1211	Carbonates	6.2–6.9	11,010 ± 110	2	Testhole X-23, paired with Y-1211B
Y-1212	Carbonates	12.8–14.9	10,440 ± 90	2	Testhole X-20, paired with Y-1212B
Y-1213	Carbonates	27.7–29.8	13,100 ± 230	2	Testhole X-20, paired with Y-1213B
Y-1214	Carbonates	45.2–45.7	17,710 ± 280	2	Testhole X-23, paired with Y-1214B
Y-1215	Carbonates	61.0–62.4	19,380 ± 250	2	Testhole X-20, paired with Y-1215B
Y-1216	Carbonates	75.3–75.6	21,380 ± 340	2	Testhole X-23, paired with Y-1216B
Y-1217	Carbonates	95.9–97.9	20,650 ± 210	2	Testhole X-23, paired with Y-1217B
Y-1218	Carbonates	99.3–100	22,450 ± 380	2	Testhole L-U-1, paired with Y-1218B
Radiocarbon ages of mud layers in the Lower Salt					
GRO-1802	Carbonates		31,100 ± 400	2	Testhole X-20, ~0.75 m above base of Lower Salt
Radiocarbon ages of the bottom mud					
Y-1224	Carbonates		32,300 ± 900	2	Testhole X-23, paired with Y-1224B

Key to references: 1 = this report, 2 = Stuiver (1964), 3 = Rubin and Berthold (1961), 4 = Ives et al. (1964), 5 = Stuiver and Smith (1979).

atmosphere or more probably from migrating interstitial water) caused the age discrepancy.

Stuiver (1964) also hypothesized that the rate of sedimentation of the Parting Mud was primarily dependent on the deposition of salts (including the moderately soluble carbonates pirssonite and gaylussite); more rapid sedimentation rates occurred during effectively dryer conditions associated with lower lake levels and increased salinities. Stuiver (1964) indicated that differing sedimentation rates could be used to distinguish three periods of effective wetness. Frolm 24,000 to 19,000 yr B.P., the sedimentation rate was 0.37 mm/yr; from 19,000 to 14,000 yr B.P. (the effectively wettest period), the sedimentation rate decreased to 0.12 mm/yr; and from 14,000 yr to 10,200 yr B.P., the sedimentation rate increased to about 0.29 mm/yr.

G. I. Smith (pers. comm., 1989) favors another hypothesis, wherein periods of rapid mud deposition also were periods of increased streamflow discharge. Approximately 35% of the Parting Mud is composed of acid-insoluble calcium–magnesium carbonates. G. I. Smith (per. comm., 1989) believes that an increase in the influx of calcium and magnesium, in part, controls the rate of sediment accumulation. In this hypothesis, an increase in sedimentation rate indicates an increase in the influx of dissolved solids, which indicates an increase in the rate of discharge.

Variations in abundance of gaylussite and pirssonite, in the chemical composition of organics, and in the spacing and color (mineralogy) of laminae were used by Mankiewicz (1975) to divide the Parting Mud into five units. The lower 40% of the Parting Mud consists mostly of massive black mud. Smith (1979) believed that this mud indicated a deep lake that was unstratified or stratified with a thick zone of fresher water that formed the upper layer. Mankiewicz (1975) studied the hydrocarbon chain length of organic samples from the Parting Mud. The unit deposited between ~24,000 and 17,100 yr B.P. contains macerated plant fragments, which indicate a relatively rapid rate of inflow. The upper 60% of the Parting Mud consists of four laminated units. Based on interpolation of radiocarbon ages, the contacts between the four units have estimated ages of 14,500, 12,300, and 11,200 yr B.P. The unit deposited between 17,100 and 14,500 yr B.P. contains closely spaced dolomite (?) laminae that Smith (1979) and Smith and Street-Perrott (1983) interpreted to have been deposited during the waning stages of a deep lake. The unit deposited between 14,500 and 12,300 yr B.P. was interpreted to indicate a persistent period of shallow water during which no salts were deposited. Widely spaced dolomite(?) laminae may indicate infrequent inflow of freshwater into a saline body of water. The two units that range in age from 12,300 to 11,200 yr B.P. and from 11,200 to 10,500 yr B.P. contain aragonite(?) laminae. Smith (1979) and Smith and Street-Perrott (1983) believed these two units were deposited in a deep lake that overflowed into the Panamint basin. A brief final moderate-size lake expansion is represented by about 50 cm of unlaminated mud found in many cores (but not in the core studied by Mankiewicz (1975).

Relative mineral abundances derived by X-ray diffraction of samples from the Parting Mud from cores KM-3 and L-12 (Fig.10, Smith, 1979) indicate the following: (1) gaylussite and pirssonite have a variable distribution but tend to be concentrated near the top and base, (2) dolomite is concentrated near or at the top and base of the unit, (3) aragonite and calcite are most abundant in the center of the unit, (4) halite is a major component of the entire unit, but decreases in abundance in those sections that have abundant gaylussite and pirssonite, and (5) fine-grained clastic silicate minerals are concentrated in the middle and upper-middle part of the unit. X-ray diffraction analysis of the Parting Mud in core GS-16 (Table 13 in Smith, 1979) indicates somewhat higher amounts of dolomite throughout most of the core.

The abundance distribution of some minerals in these cores is somewhat different from that reported by Mankiewicz (1975). Dolomite was near the base of the unit in cores KM-3 and L-12 (Fig.10); however, it only occurred in the top one-half of core B studied by Mankiewicz (1975). Calcite was not reported by Mankiewicz (1975) but was in the upper one-half of KM-3 and L-12, usually in association with aragonite. Difference in the vertical distribution of minerals in different cores from Searles Lake indicates areal variation in

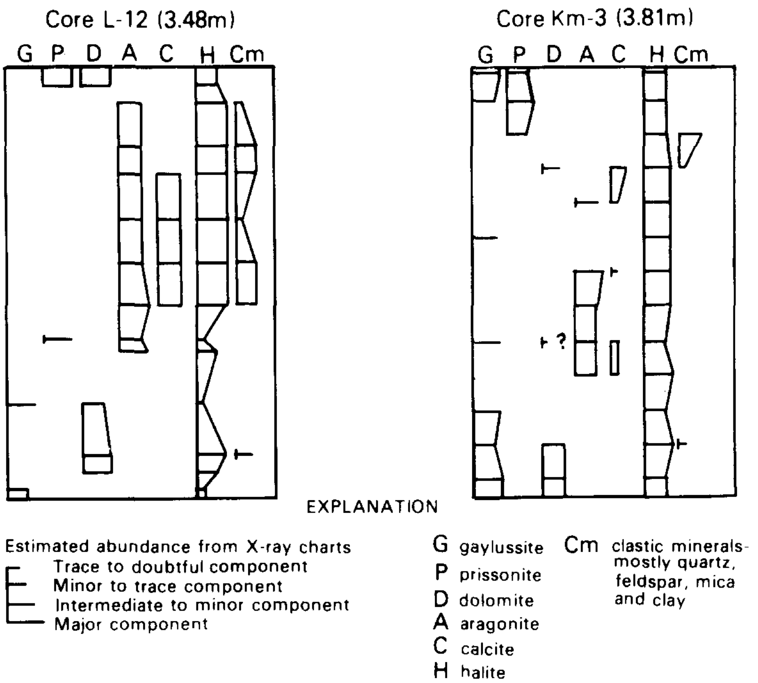

Fig.10. Mineral abundance in samples from two cores of Parting Mud. Samples from core L-12 represent silt- and clay-sized components; samples from core KM-3 also include megascopic components (modified from Smith, 1979).

the mineralogy of the Parting Mud or that the techniques used by Mankiewicz (1975) and Smith (1979) to identify certain minerals did not yeild equivalent results. For example, optical identification of minerals is biased toward those minerals larger than a certain size and x-ray diffraction peaks of one mineral may obscure the peaks of another.

The Lower Salt contains six layers of mud (M-2 through M-7) (Smith, 1962). Stuiver (1964) and Stuiver and Smith (1979) indicated that these mud layers differ from those in the Upper Salt in several ways: (1) they are much thicker, (2) the organic-carbon content is about the same as that in the Parting Mud and Bottom Mud, whereas the organic carbon content of the mud in the Upper Salt is much smaller, (3) successive mud layers in the Lower Salt are progressively younger, which indicates that washing in of large quantities of older sediments did not occur, and (4) most of the time represented by the Lower Salt is accounted for by the mud layers.

Stuiver (1964) and Stuiver and Smith (1979) indicated that radiocarbon ages of organics from the Lower Salt ranged from $32,620 \pm 500$ to $23,750 \pm 300$ yr B.P. Most of Stuiver's (1964) radiocarbon ages of materials from the Lower Salt have not been incorporated in Table 3 because of insufficient documentation of stratigraphic placement, but newer, stratigraphically precise radiocarbon-age data from Stuiver and Smith (1979) have been incorporated. Each of the mud units in the Lower Salt consists chiefly of gaylussite, pirssonite, halite (predominant mineral in M-4 and M-7 mud units), and some aragonite. The relative quantities of gaylussite and pirssonite seem to be related to site of deposition (Smith, 1979). The presence and style of bedding changes with time. In the oldest units (M-2 and M-3), laminar bedding is well developed; in units M-4 and M-5, laminar bedding is faint; and in the youngest units (M-6 and M-7), laminar bedding is indistinct or nonexistent. In units M-2 to M-5, the decrease in distinctive laminae with time indicates that successive mud units were deposited in lakes that were subjected to a higher degree of convective mixing; i.e., the degree of meromixis decreased in successively younger lakes and was probably absent during deposition of M-6 and M-7.

The Bottom Mud is about 30 m thick and consists of clay- to silt-sized, dark-green to black mud that contains megascopic crystals of gaylussite. The mud mostly consists of fine-grained carbonates (gaylussite, dolomite, aragonite, and calcite), evaporite minerals, and clastic silicate minerals. Gaylussite is a component in about 90% of the samples. The gaylussite crystals are subhedral to anhedral and generally cut across bedding, which indicates they grew after burial (Smith and Haines, 1964). The mineral distribution in the upper mud unit in cores 254 and L-30 (plate 2 in Smith, 1979) is similar to the distribution of minerals in the Parting Mud. Gaylussite and dolomite (instead of pirssonite) are the predominant minerals at the top and base of the Bottom Mud. Calcite, aragonite, and fine-grained clastic silicates are most abundant in the middle of the Bottom Mud; and halite is common throughout the Bottom Mud.

Radiocarbon ages of samples from the Lower Salt and Bottom Mud indicate that deposition of the Lower Salt began about 32,620 $\pm$ 500 yr B.P. and ended about 23,750 $\pm$ 300 yr B.P. (Table 3) (Stuiver and Smith, 1979). A plot of radiocarbon ages versus depth (fig.31 in Stuiver and Smith, 1979) indicates the relatively long periods of deposition represented by the muds. For example, units M-2, M-6, and M-7 account for about 6500 yr of the 9000-yr depositional history represented by the Lower Salt. The top of unit S-5 (radiocarbon age = 28,000 yr B.P.) may represent the only period of desiccation that occurred during deposition of the Lower Salt unit (as indicated by the large number of salt mineral types, the largest percentage of halite present, and the thickness of the unit) (Smith, 1979). The number of salt mineral types is not necessarily a diagnostic criterion for desiccation because some of the very soluble salts may dissolve in the waters of the next lake to form. The accuracy of the radiocarbon ages of the Lower Salt depends in a complicated way on the source of the organic carbon in Searles Lake (modern allochthonous, fossil autochthonous, or fossil allochthonous) and on the exchange rate of carbon across the air–water interface (Peng et al., 1978; Stuiver and Smith, 1979).

Peng et al. (1978) compared radiocarbon and ^{230}Th ages of the Lower Salt and subtracted 900 yr from Stuiver and Smith's (1979) radiocarbon ages of the tops and bottoms of the mud units. This procedure assumed that: (1) the organic carbon in the lake was totally autochthonous in origin, (2) lake area ranged between 500–600 km^2 during mud deposition, (3) the exchange rate was about 6 mol m^{-2} yr^{-1}, (4) the salt layers were deposited nearly instantaneously, (5) the initial ^{232}Th/^{230}Th ratio was the same for all salt deposits, and (6) the radiocarbon and ^{230}Th ages should be the same. Peng et al. (1978) suggested the following corrected salt-layer ages: 22,900 yr B.P. (S-7), 24,700 yr B.P. (S-6), 27,400 yr B.P. (S-5), 27,900 yr B.P. (S-4), 29,400 yr B.P. (S-3), 29,400 yr B.P. (S-2), and 31,800 yr B.P. (S-1).

In the preceding paragraph, some of these assumptions (1, 2, 5, and 6) are probably incorrect, and the others are only approximations. For that reason, the use of uncorrected radiocarbon ages is considered equally valid. Uncorrected salt-layer radiocarbon ages are: 23,700 yr B.P. (S-7), 25,500 yr B.P. (S-6), 28,600 yr B.P. (S-5), 28,800 yr B.P. (S-4), 30,300 yr B.P. (S-3), 30,500 yr B.P. (S-2), and 32,500 yr B.P. (S-1). In general, the age of the bottom 0.05 m of each mud unit is the same (within the counting error) as the upper 0.05 m of the next lower mud unit (Table 3). For some layers, the ages are stratigraphically reversed. This reversal indicates that deposition of the salt units occurred in a very short time and that some of the organic carbon dates are in error.

In a study of the pluvial history of the Panamint Valley, R. S. U. Smith (1978) indicated (based on the degree of weathering) that the last spill of Lake Searles produced a shoreline at an elevation of 335 m (depth = 38 m) in the south Panamint Valley subbasin. No datable material has been found in association with this shoreline.

Lake Searles chronologies for the past 35,000 yr are shown in Fig.11. The chronology shown in Fig.11a is based on the data interpretations of Smith (1979) and on unpublished correlations of sedimentary units in core with outcropping sedimentary units. The Overburden Mud is correlated

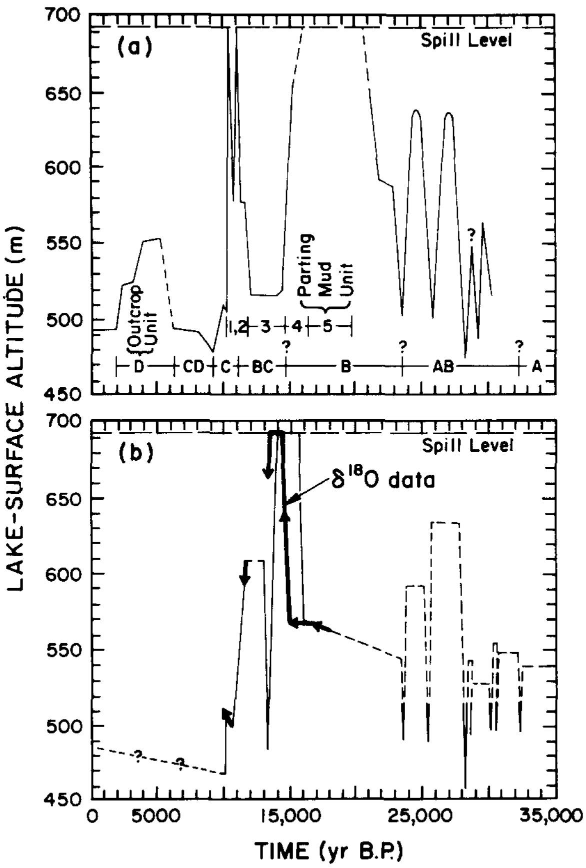

Fig.11. Model chronologies for Lake Searles. The solid line in (a) is based on unpublished data referenced in Smith and Street-Perrott (1093). The chronology in (b) is based on a reinterpretation of the data from Smith (1979), on rock-varnish ages from Dorn (this volume) and on unpublished oxygen-isotope data of Fred Phillips (pers. comm., 1989).

with lacustrine sediment that crops out about 55 m above the present-day basin floor. Lake-surface altitudes during deposition of the Parting Mud and of the muds of the Lower Salt were obtained by correlation of core-based mud units with outcropping lacustrine mud units. This chronology is nearly identical to that shown in fig.10-5 of Smith and Street-Perrott (1983).

An alternative lake-level chronology for Lake Searles is shown in Fig.11b. This chronology is based on the mineralogy of cores GS-16, KM-3, and L-12 (Fig.10), on different interpretations of the depositional environments of gaylussite and pirssonite, on rock-varnish radiocarbon ages of the overflow-high stand terrace, and on one possible interpretation of unpublished $\delta^{18}O$ analyses of carbonates (mostly dolomite) that were extracted from core KM-3 (Fred Phillips, New Mexico Inst. Mining and Tech., pers. comm., 1989). Altitudes of outcropping units of the Lower Salt and Bottom Mud mapped by G. I. Smith (unpublished data) have been used in the construction of Fig.11b. The senior author has not visited the outcrops nor seen the geologic maps based on correlations of outcrop and subsurface units. For that reason, he is unable to judge or attest to the accuracy of the correlations and has arbitrarily used the altitudes of the Lower Salt and Bottom Mud only as a background within which to illustrate that the amount of time represented by salt deposition may have been small (salt units in the Lower Salt are assumed to have been deposited in a brief time period — <200 yr — after the cessation of inflow from the China Lake basin) compared to the amount of time associated with mud deposition; i.e., the deposition of mud and salt is represented by a sawtooth pattern in Fig.11a and by a square-wave pattern in Fig.11b. Unadjusted radiocarbon ages of the salts were used in developing this chronology.

Dolomite present in the lower part of cores GS-16, KM-3, and L-12 may have formed in a relatively shallow lake of moderate-to-high salinity, which had a high magnesium/calcium dissolved-solids ratio (De Deckker and Last, 1988). The presence of dolomite in the lower part of the cores indicates that Lake Searles may have been relatively shallow between 24,000 and $\sim$20,000 yr B.P. In addition, microscopic crystals of gaylussite and pirssonite (core L-12 in Fig.10) may have precipitated directly from a shallow, briny lake, and megascopic crystals of gaylussite and pirssonite (core KM-3 in Fig.10) may have formed penecontemporaneously in a sodium-rich interstitial brine. The interpretation is consistent with the presence of a shallow lake between 24,000 and $\sim$20,000 yr B.P.; similar data (Fig.10) may also indicate the presence of a shallow lake between 13,000 and 11,000 yr B.P.

The presence of significant amounts of dolomite from intermediate depths in core GS-16 (table 13 in Smith, 1979) also may indicate relatively shallow lake depths between about 20,000 and 17,000 yr B.P. The relative lack of dolomite in samples PM 3 and PM 5 from depth intervals of 22.1–22.3 m and 22.6–22.9 m in core GS-16 (table 13 in Smith, 1979) may indicate the occurrence of relatively deep-lake conditions between about 17,000 and 13,000 yr B.P. The relatively few number of analyses and the necessity of age interpolation hinders precise interpretations of the lake level record.

The dolomite $\delta^{18}O$ data of F. Phillips (pers. comm., 1989) also permit a tentative reconstruction of lake-level in the Searles Lake basin from about 18,300 yr B.P. to about 10,500 yr B.P. (arrows in Fig.11b indicate lake-level trajectories bases on ^{18}O data). The data indicate the presence of a shallow lake at 18,300 yr B.P. ($\delta^{18}O = 40‰$ relative to SMOW). Between 18,200 and 17,700 yr B.P., Lake Searles rose to a moderately shallow or intermediate level ($\delta^{18}O = 37‰$) and remained there until ~16,000 yr B.P. Between ~16,000 and ~15,000 yr B.P., the lake again rose ($\delta^{18}O = 33‰$) and fluctuated about this highstand condition until about 13,500 yr B.P. at which time lake level began to decline. The lake seems to have been moderately deep ($\delta^{18}O = 35‰$) until ~11,300 yr B.P. During the following 300 yr, lake level again declined ($\delta^{18}O = 37‰$ at 11,000 yr B.P.). Between 11,000 and 10,500 yr B.P., lake level rose ($\delta^{18}O = 33‰$).

Rock varnish collected from the overflow-highstand terrace at Poison Canyon on the western side of Searles Lake basin has a radiocarbon age of 13,610 ± 110 yr B.P.; varnish from the Gold Bottom Mine locality on the eastern side of the basin has an age of 13,290 ± 120 yr B.P. (Dorn et al., 1990). These data indicate that lake-level recession began prior to 13,500 yr B.P. and that the lake did not rise to its spill point after this time.

The sediments of the Overburden Mud are unlike sediments from any of the deeper mud units. The lack of laminae and the presence of coarse clastic units indicates that the Overburden Mud may represent an alluvial deposit. This interpretation implies that lakes did not form in the Searles Lake basin during the Holocene as the result of spill from the Owens Lake basin (Fig.11b).

Radiocarbon ages of carbonates and of two samples of disseminated organic carbon that were extracted from five sedimentary units that crop out in the Searles Lake basin are listed in the first two sections of Table 3. Units that have double letters are correlated with relatively shallow lakes or alluvial facies that crop out at low altitudes in the basin. These correlations are depicted in Fig.11a.

Samples from the same outcrop unit are arranged by altitude. An age/altitude plot of samples from units B, C, and CD (Fig.12) indicates that some carbonate samples have been reworked and redeposited; others have been contaminated with older or younger carbon. For example, three samples from unit B have radiocarbon ages that are younger or are in the age range of samples from unit C (Table III). This stratigraphic reversal of ages indicates incorporation of secondary carbon.

Smith (1979) and unpublished data correlated the upper two laminated aragonitic units of the Parting Mud (radiocarbon ages on organic carbon = 12,300–10,500 yr B.P.) with outcrop Unit C (Fig.11a). However, 12 of the 20 radiocarbon ages of carbonate samples from Unit C are as much as 2000 yr older than 12,300 yr B.P. (Table 3; Fig.12). These ages may indicate that Lake Searles possessed a low initial $^{14}C/^{12}C$ ratio consistent with the need for age corrections of the Lower Salt discussed by Peng et al. (1978). However, two factors contradict this conclusion: (1) Searles Lake was supposedly at a highstand during the deposition of Unit C (Fig.11a) which would minimize (but not necessarily eliminate) the concentration of old carbon in the lake, and (2) radiocarbon ages of carbonates from other lake basins (Figs.4, 5, and 7) usually yield ages that are too young rather than too old. Because continuous tracing of the two laminated aragonitic units from core to outcrop is physically impossible, the possibility exists that unit C and the two laminated units are not time-stratigraphic correlatives. The age of the one sample from unit CD is too old and came from a sedimentary facies that contained mollusc shells probably reworked from an older unit.

Radiocarbon ages of some samples from older units are reliable but many are not. Units AB and A samples range in age from 10,230 + 2200 yr B.P. to > 36,000 yr B.P. (Table 3, not plotted in Fig.12). If correlation of units A and AB with the Bottom Mud and Lower Salt is correct, then about 60% of unit AB and about 33% of unit A samples have radiocarbon ages within the correct age range.

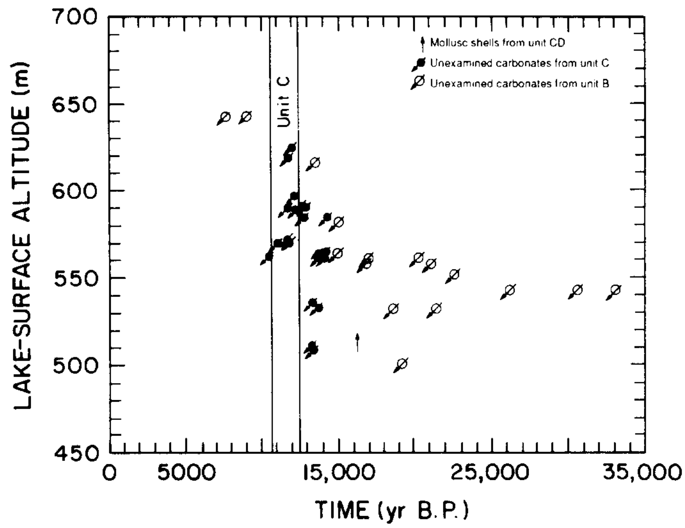

Fig.12. Radiocarbon age-altitude plot of Searles Lake basin carbonates.

Lake Russell

The Mono Lake drainage basin (Fig.13), which contained Lake Russell, is located on the western edge of the Great Basin (Fig.3). Mono Lake lies at the base of the steep eastern escarpment of the Sierra Nevada at an altitude of about 1950 m. The three principal streams that enter Mono Lake (Rush, Lee Vining, and Mill Creeks) originate in the high Sierra Nevada (Fig.13). The Mono Lake basin differs from the Lake Bonneville, Lake Lahontan, and Searles Lake basins in its small size, relative simplicity (only one subsidiary basin), and proximity to its catchment area.

Russell (1889) did the first systematic study of the lake that occupied the Mono basin during the late Pleistocene. Russell (1889) concluded that the lake had not spilled southward into the Owens River drainage during its most recent highstand but instead had spilled during its previous highstand. Putnam (1949) agreed with Russell's (1889) interpretation, but incorrectly correlated recessional lacustrine terraces in the Mono basin with recessional glacial moraines in three major canyon systems that drain the high Sierra Nevadas. Putnam (1949) referred to the Pleistocene lake in the Mono basin as Lake Russell in honor of the pioneering work of Russell.

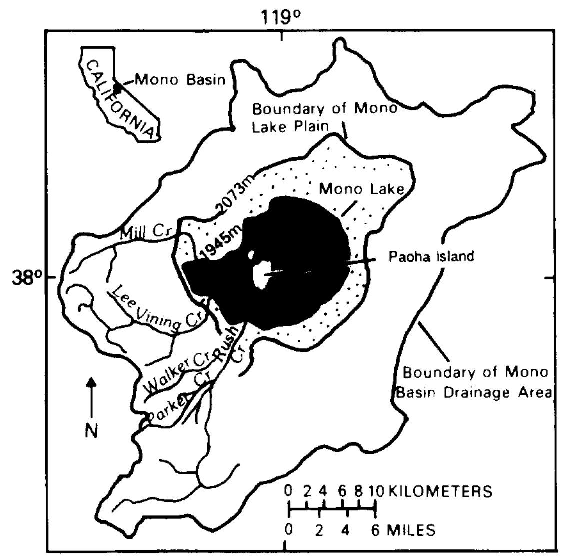

Fig.13. Drainage area of the Mono Lake basin (modified from Vorster, 1985). Stippled area is the Mono Lake plain.

Lajoie (1968) did the first detailed stratigraphic study in the Mono basin and concluded that two large delta terraces that crop out at altitudes of 2095 m and 2035 m were formed during the glacial maximum when the lake stood at intermediate levels. Lajoie et al. (1982) further concluded that the lake rose to its last highstand (2155 m) well after the ice had receded from the lower reaches of adjacent glacial valleys and that the lake did not spill to the Owens River drainage at that time.

Lajoie (1968) reported that most of the lacustrine deposits of Pleistocene age in the Mono basin are restricted to the area below 2075 m. The most widely distributed rock-stratigraphic unit is informally referred to as the Wilson Creek beds and has been named for its excellent exposures along the lower course of Wilson Creek. The Wilson Creek beds consist of 6–15 m of finely laminated to very thinly bedded, lacustrine clayey silts that were deposited in relatively deep water. In the type section, these beds directly overlie alluvial gravels. On Paoha Island in the center of Mono Lake, the Wilson Creek beds overlie at least 200 m of lacustrine silts that are informally referred to as the Paoha Island beds (Lajoie, 1968). The Wilson Creek beds contain 19 thin layers of volcanic ash that form 5 distinctive marker beds that provide precise stratigraphic control throughout the basin. The Wilson Creek beds laterally interfinger with deltaic and terrace sands that form the two prominent terraces at 2075 m and 2035 m. In the northeastern part of the basin, a similar, but thinner bed of sand in the lower part of the Wilson Creek beds, correlates with a distinctive marl unit in the type section. Both the lower and upper sand beds thicken to the north. The two coarse-grained units represent relative lowstands, and the three interbedded silt beds represent relative highstands of Lake Russell. The upper silt bed, which drapes over the two prominent constructional terraces, was deposited during the last highstand. Lajoie (1968) correlated the upper silt bed and the erosional terrace cut into bedrock at 2155 m in the western part of the basin with the last highstand of Lake Russell. Russell (1889) and Putnam (1949) also correlated this terrace with the last highstand.

Black Point, a flat-topped basaltic cinder cone that is located on the northern shore of Mono Lake, erupted subaqueously during the last highstand. A deltaic deposit, consisting entirely of this distinctive tephra, is located on the Sierran escarpment about 6 km west of Black Point. Lake level at the time of eruption is indicated by the altitude of this deposit and the altitude of the cinder cone.

Broecker and Walton (1959) reported that Mono Lake has a low $^{14}C/^{12}C$ ratio as a result of the slow rate of exchange of dissolved carbon with atmospheric CO_2 (apparent radiocarbon age $\sim$1700 yr). Modern carbonate samples that formed during the past several decades have radiocarbon ages that range from about 1100 to 5300 yr B.P. (Table 4). The degree of deviation from a modern age probably represents the quantity of dead carbon incorporated within the sample during its depositional history. If a tufa sample is deposited near a spring, the water of which contains dead carbon, the quantity of dead carbon incorporated within the sample will depend on the mixing ratio of spring and lake water. This process indicates the difficulty inherent in assessing the reliability of radiocarbon ages of carbonate and organic carbon samples deposited in the Mono Lake basin.

Lajoie (1968) originally based the age of the Wilson Creek beds (23,000–12,000 yr B.P.) on radiocarbon ages of four carefully cleaned fossil-ostracode samples obtained from modern exposures of the type section and from a nearby outcrop. More recent radiocarbon-age data on platy tufa, carbonate crusts, carbonate (tufa) nodules, and ostracode valves from various exposures of the Wilson Creek beds extend their age to about 36,000 yr B.P. (Table 4). The lake-level chronology for Lake Russell indicated by the solid line in Fig.14, is primarily based upon lithologic, stratigraphic, and geomorphic relations and not on a line connecting the altitudes of shallow-water deposits or on lake levels inferred from lake-sediment mineralogy. Radiocarbon ages of materials from the type section of the Wilson Creek beds provide the primary chronologic control. Radiocarbon ages of exposed and stratigraphically controlled platy tufas and tufa mounds from other parts of Mono basin substantiate the primary lake-level chronology, including the correlation of the last highstand with the erosional terrace at 2155 m (Fig.14).

Salinity estimates derived from nine fossil ostracode assemblages (R. Forester, pers. comm., 1988) to some extent substantiate the Lake Russell chronology (Fig.14). Lake level was estimated from three ostracode assemblages that could be stratigraphically associated with paleolake levels.

Microscopic (thin-section) examination of the marls and tufas in the Wilson Creek beds for the

TABLE IV

Radiocarbon ages, sample, and locality data for samples from Lake Russell

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
Organic-carbon samples deposited at or above lake level					
LDGO-16770W	Wood	1947	550 ± 30	1	Rooted tree stump killed by rising lake level
LDGO-1677IW	Wood	1950	570 ± 30	1	Rooted tree stump killed by rising lake level
USGS-1315	Wood	1943	600 ± 70	1	Non-rooted tree stump, dates rise from lowstand
USGS-1167	Wood	1947	640 ± 70	1	Rooted tree stump killed by rising lake level
USGS-1277	Wood	1951	770 ± 60	1	Rooted tree stump killed by rising lake level
LDGO-1677AW	Wood	1950	700 ± 50	1	Rooted shrub, dates rise from lowstand
USGS-1274	Wood	1948	850 ± 50	1	Rooted tree stump killed by rising lake level
LDGO-1677FW	Wood	1948	860 ± 60	1	Rooted tree stump killed by rising lake level
USGS-1276	Wood	1950	860 ± 60	1	Rooted tree stump killed by rising lake level
USGS-1275	Wood	1949	910 ± 60	1	Rooted tree stump killed by rising lake level
UCLA-118	Wood	1951	920 ± 90	1	Rooted tree stump killed by rising lake level
USGS-1704	Pine cone	1946	930 ± 40	2	In beach sand, dates rise from lowstand
USGS-1319	Wood	1942	940 ± 60	1	Rooted tree stump killed by rising lake level
LDGO-1677LW	Wood	1947	940 ± 20	1	Rooted tree stump killed by rising lake level
USGS-1487	Wood	1951	950 ± 20	1	Rooted tree stump killed by rising lake level
USGS-1705	Pine cone	1946	1590 ± 50	1	Fluvial gravels, dates lake stand
USGS-1524	Pine cone	1946	1625 ± 40	1	Fluvial gravels, dates lake stand
Organic-carbon samples deposited below lake level					
USGS-1317	Charcoal	1961	220 ± 65	1	In lake sediments, dates rise or highstand
USGS-1320	Charcoal	1966	220 ± 60	1	In bottomset deltaic sediments, dates highstand
USGS-1316	Pine cone	1950	675 ± 50	1	In shallow-water sands, dates littoral zone
USGS-1706	Pine cone	1951	760 ± 50	2	Lower Lee Vining Creek
USGS-1310	Pine cone	1943	890 ± 60	1	In transgressive sediments, dates rise from lowstand
USGS-1321	Marsh grass	1945	920 ± 40	1	In beach sand
USGS-1704	*Artemisia*	1951	930 ± 40	2	Lower Rush Creek
USGS-1692	Wood	1951	1380 ± 50	2	Lower Lee Vining Creek
USGS-1694	Wood	1951	1450 ± 170	2	Lower Lee Vining Creek
USGS-1314	Pine cone	1946	1495 ± 45	1	In transgressive sediments, dates rise
USGS-1705	Pine cone	1951	1590 ± 50	2	Lower Lee Vining Creek
BETA-5115	Charcoal	1977	3490 ± 90	1	In bottomset beds, dates highstand
USGS-1539	Disseminated organic carbon	1943	$14{,}320 \pm 110$	3	Silt containing Ash Layer II-6, Negit Causeway
Modern carbonate samples					
USGS-275	Tufa (punky)	1943	1105 ± 30	3	Modern (?) tufa tower on south lake shore
USGS-225	Tufa (punky)	1954	1830 ± 60	3	Modern (?) tufa from tower on Wilson Creek fan
USGS-539	Tufa	1945	1840 ± 40	3	Tufa coating modern stick
USGS-532	Tufa	1945	1940 ± 60	3	Modern tufa tower
USGS-70	Tufa	1951	2060 ± 60	3	Tufa coating historical-age wood

TABLE IV (*continued*)

Sample number	Sample type	Alt. (m)	^{14}C age (yr B.P.)	Ref.	Stratigraphic data
USGS-346	Tufa (punky)	1954	2120 ± 30	3	Modern tufa tower, Wilson Creek fan
USGS-1168	Tufa	1947	3240 ± 50	1	Tufa coating tree stump (USGS-1167)
USGS-274	Carbonate-cemented sand tube	1950	4680 ± 70	3	Tufa from south lake shore, same as USGS-1017
USGS-1017	Carbonate-cemented sand tube	1950	5310 ± 60	3	Sand tube in modern beach, same as USGS-274
Carbonate samples deposited below lake level that have not been covered by younger sediment: samples have not been thoroughly examined for contamination					
USGS-254	Tufa (punky)	1957	6960 ± 60	3	Tufa tower on Wilson Creek fan
USGS-302	Tufa (thinolite)	2017	8540 ± 70	3	Tufa mound west of Wilson Creek
USGS-348	Tufa	1965	8810 ± 40	3	Outer coating on tufa tower, inner part = USGS-368
USGS-349	Tufa	1964	8930 ± 40	3	Tufa tower on crest of fault scarp
USGS-368	Tufa (punky)	1965	9280 ± 50	3	Inner core of tufa tower, coated by USGS-348
USGS-253	Tufa (punky)	1965	9520 ± 60	3	Tufa tower on Wilson Creek fan
USGS-304	Tufa (thinolite)	2005	10,220 ± 45	3	From Black Point cinder pit, coated by USGS-303
USGS-235	Tufa (thinolite)	1982	10,650 ± 230	3	Coating on Black Point cinders
USGS-347	Tufa	2111	11,790 ± 50	3	Tufa tower north of lower DeChambeau terrace
USGS-345C	Tufa	2097	12,030 ± 60	3	Outer part of tufa from lower Dechambeau terrace
USGS-300	Tufa (punky)	2114	12,770 ± 100	3	From summit of Black Point
USGS-180	Tufa	2098	12,780 ± 140	3	On walls of Black Point Crevasse
USGS-1564	Tufa	2155	12,910 ± 70	3	Tufa at altitude of highstand terrace on northwest Mono Craters
USGS-345A	Tufa (crystalline)	2097	13,030 ± 70	3	Inner core of tufa from lower DeChambeau terrace
USGS-345B	Tufa	2097	13,410 ± 45	3	Middle of tufa from lower DeChambeau terrace
USGS-179	Tufa	2097	14,080 ± 200	3	Tufa tower on DeChambeau terrace
USGS-129	Tufa	2030	16,870 ± 210	3	Tufa tower on Cowtrack Mountain terrace
USGS-1827	Tufa	2037	19,230 ± 100	3	Coating on bedrock cobble near Bridgeport Creek
Carbonate samples deposited at Black Point locality					
USGS-295	Tufa	2019	10,630 ± 50	3	Above Black Point Ash, Black Point cinder pit
USGS-301	Tufa (thinolite)	1982	10,640 ± 90	3	Above Black Point Ash, Black Point cinder pit
USGS-303	Tufa	2005	11,970 ± 60	3	Above Black Point Ash, Black Point cinder pit
USGS-296	Tufa	2073	12,330 ± 60	3	Above Black Point Ash, north slope of Black Point
Carbonate samples deposited within the Wilson Creek type section; ostracodes and tufa nodules examined under microscope					
USGS-369	Tufa	2047	10,980 ± 80	3	20 cm above Ash No. 2, Fish Hatchery Rd., same horizon USGS-273B
USGS-260	Tufa	1964	11,730 ± 100	3	Probably above Ash No. 1, Bodie Junction
USGS-273A	Tufa	2047	11,830 ± 120	3	25 cm above Ash No. 2, Fish Hatchery Road
USGS-273B	Tufa (thinolite)	2047	11,980 ± 50	3	55 cm above Ash No. 2 from same horizon as USGS-273A, USGS-261
USGS-361	Tufa (thinolite)	2047	11,990 ± 90	3	55 cm above Ash No. 2 from same horizon as USGS-273B
USGS-261	Tufa	1975	12,610 ± 100	3	Probably above Ash No. 1, Bodie Junction
USGS-602	Carbonate crust	1996	12,800 ± 60	3	5 cm above Ash No. 1, 6.98 m above base
USGS-600	Carbonate crust	1996	13,050 ± 80	3	5 cm below Ash No. 1, 6.88 m above base
L-1167A	Ostracode valves	1985	13,700 ± 500	3	Between Ash beds No. 2 and No. 3, 6.5 m above base
USGS-1435	Ostracode valves	1983	19,720 ± 160	3	Below 7-cm thick sand in upper part of section
USGS-1436	Ostracode valves	1973	22,610 ± 600	3	Between Ash No. 11 and No. 12

USGS-835	Carbonate nodules	1974	23,000 ± 200	3	39 cm below Ash No. 7, 3.94 m above base
L-1167C	Ostracode valves	1973	23,700 ± 300	3	Between Ash No. 11 and 12, 2.99 m above base, dupl. of USGS-1436
USGS-833	Carbonate nodules	1973	24,850 ± 190	3	7 cm below Ash No. 8, 3.18 m above base
USGS-834	Carbonate nodules	1973	28,000 ± 300	3	5 cm below Ash No. 14, 2.87 m above base
USGS-805	Carbonate nodules	1972	28,150 ± 190	3	20 cm below Ash No. 15, 2.27 m above base
USGS-362	Tufa	1970	28,600 ± 350	2	Below Ash No. 19, base of type section
USGS-282B	Carbonate-cemented silt	1971	28,720 ± 290	3	7 cm below Ash No. 16, 0.81 m above base, duplicate of USGS-282A
USGS-636	Carbonate nodules	1980	29,100 ± 420	3	8 cm below Ash No. 15, 2.39 m above base
USGS-1013	Carbonate nodules	1980	30,400 ± 300	3	63 cm below Ash No. 15, 1.85 m above base
USGS-637	Carbonate nodules	1980	31,500 ± 250	3	20 cm above Ash No. 16, 1.07 m above base
USGS-282A	Carbonate-cemented silt	1971	33,760 ± 470	3	7 cm below Ash No. 16, 0.81 m above base
USGS-1011	Carbonate nodules	1970	34,700 ± 250	3	20 cm below Ash No. 17, 0.43 m above base
USGS-1012	Carbonate nodules	1970	34,900 ± 250	3	33 cm below Ash No. 17m 0.31 m above base
USGS-276	Tufa (crystalline)	1984	39,600 ± 1000	3	Between Ash No. 17 and No. 19, near base
Carbonate samples deposited within the Wilson Creek beds in other localities					
USGS-1015	Tufa	2033	15,190 ± 90	3	Between Markers A and C, upper deltaic sands, Wilson Creek
USGS-1014	Tufa	2027	17,930 ± 130	3	Between Markers A and C, middle deltaic sands, Wilson Creek
USGS-1016	Tufa	2024	18,850 ± 100	3	Between Markers A and C, lower deltaic sands, Wilson Creek
USGS-1385	Tufa	2027	20,000 ± 90	3	Littoral sand and gravel below Marker A, West Dry Creek
L-1167F	Ostracode valves	2028	23,200 ± 700	3	Ash No. 8 equivalent (?), Sta. E, upper Wilson Creek
L-1167G	Ostracode valves	1981	32,300 ± 1300	3	Below Ash No. 19, Rush Creek

Key to references: 1 = Stine (1987), 2 = collected by Michael Perkins (pers. comm., 1983), 3 = this report.

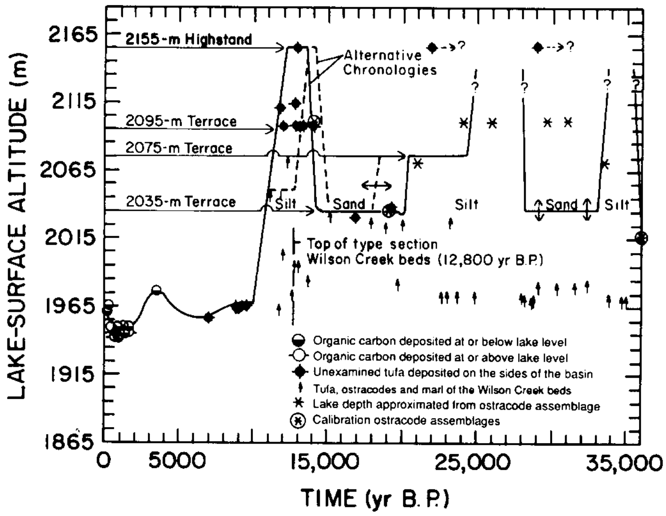

Fig. 14. Model chronologies for Lake Russell. The solid line is the chronology preferred by Lajoie (Ken Lajoie, pers. comm., 1989). An alternative chronology based on possible errors within the set of radiocarbon ages is shown by dashed-line segments.

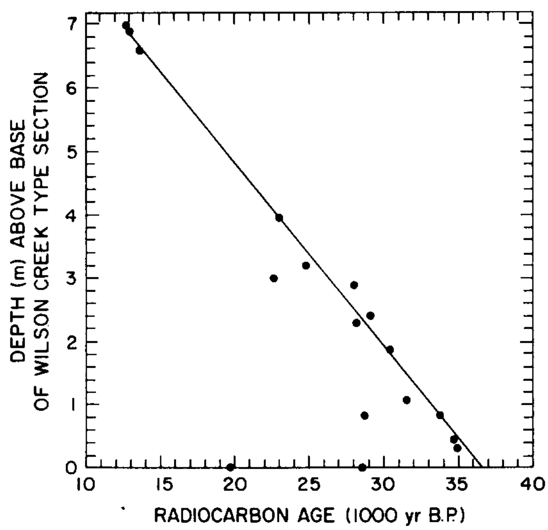

Fig. 15. Age of sediments of the Wilson Creek type section as a function of distance above base of the unit based on radiocarbon ages of carbonate samples.

presence of secondary carbon was not routinely done. However, most of the dated materials from the type section were examined for the presence of contaminants. Ostracode valves that had secondary coatings of carbonate were not submitted for radiocarbon-age estimation. The ages of the ostracode assemblages were approximated by plotting the radiocarbon ages of carbonate samples in the type section of the Wilson Creek beds as a function of their distance above the base of the unit (Fig. 15). The increase in the scatter of radiocarbon ages with increasing depth in the section indicates the incorporation of younger secondary carbon. The radiocarbon age (21,900 ± 600 yr B.P.) (Hubbs et al., 1965) of a tufa sample from the highest sharply marked terrace in the Mono basin and the radiocarbon age (29,000 ± 1000 yr B.P.) (S. Stine, pers. comm., 1989) of a tufa sample from 2 m below the highstand sill also are plotted in Fig. 14. In other lake basins, tufas that have radiocarbon ages in excess of 20,000 yr B.P. have yielded unreliable ages (Lao and Benson, 1988). For this reason, the radiocarbon ages of these samples should be considered minimum estimates of their true ages.

Lajoie (1968) reported that lithoid tufa of Wilson Creek age often was coated with thinolite tufa. He concluded that the thinolite tufa was deposited during the last regressive phase of the lake. However, the radiocarbon ages of the thinolite samples may have been reset by recrystallization; e.g., lithoid sample USGS-303 has a radiocarbon age 1750 yr older than the thinolite sample (USGS-304) it coats. An additional layer of dense lithoid tufa coats the thinolite tufa layer in lower parts of the lake plain (< 2033 m) and may represent a minor pre-Holocene transgression. The late Holocene history of lake-level fluctuation in the Mono Lake basin was reconstructed from radiocarbon ages of wood and other forms of organic carbon (Table 4) (Stine, 1981, 1984, 1987).

Two lake-level chronologies for Lake Russell are shown in Fig.14. In the primary chronology (Fig.14, solid line), the radiocarbon ages of samples (except thinolite tufas) listed in Table 4 and lake-level estimates based on ostracode salinity tolerances are accepted as essentially correct. In this chronology, which resembles one previously suggested by Lajoie et al. (1982):

(1) Lake Russell was at a low level prior to 35,000 yr B.P. and began to rise at 36,000 yr B.P.

(2) Lake Russell remained at high or intermediate levels between 35,000 and 21,000 yr B.P. Lajoie et al. (1982) believed that two beds of silt that are interbedded with terrace sand and gravel were deposited about 34,000 and 26,000 yr B.P. and indicate the presence of deep lakes (see fig.11 in Forester, 1987).

(3) Lake Russell fell prior to 20,000 yr B.P. to 2035 m where it remained for about 6000 yr during the deposition of deltaic sands that were derived from glacial outwash (Lajoie, 1968).

(4) Lake Russell rose between 15,000 and 13,000 yr B.P. to its highest level after which it fell to about 1965 m by 10,000 yr B.P.

(5) Lake Russell remained at relatively low levels during the Holocene.

An alternative chronology (indicated by the dashed line in Fig.14) that is somewhat similar to the chronologies of Lakes Lahontan and Bonneville is based on possible errors inherent in some sample subsets of radiocarbon ages. The first subset consists of tufas there were deposited after the last highstand at an altitude of about 1965 m. Based on the quantity of dead carbon incorporated in tufas that are forming in the modern lake, the radiocarbon ages of the early Holocene tufas probably are more than 1000 yr too old.

The second subset consists of radiocarbon ages of tufas that were deposited during the last highstand. In the other lake basins, carbonates deposited during the last highstand have radiocarbon ages that are consistently too young (Fig.5). Therefore, the time of recession from the last highstand may have occurred prior to 12,800 yr B.P.

The third subset consists of carbonate samples that were deposited within the sediments of the type section of the Wilson Creek beds. These samples are from a very recent exposure and contain carbonates that may not have had as much radiocarbon contamination as tufa samples that were deposited along the sides of the basin. The top of the type section of the Wilson Creek beds has a radiocarbon age of 12,800 ± 60 yr B.P. This age is older than the ages of several highstand tufas from altitudes in excess of 2090 m (Fig.14) and indicates that carbonate samples from older exposures and outcrops have probably been contaminated with younger carbon (Fig.14; Table 4). The youngest Wilson Creek sediments may have been deposited in a moderately deep (> 2045 m) post-highstand lake.

The fourth sample subset consists of tufas that were deposited in deltaic sands (USGS-1014, 1015, and 1016) and littoral sands and gravels (USGS-1385) (Table 4). These porous sediments provide conduits for carbon-bearing ground water that discharges from the lake bottom. Today, similar environments produce tufas with radiocarbon ages 1100–5300 yr too old (Table 4). Sedimentation in a shallow-water deltaic environment also would be expected to occur at a much faster rate than sedimentation in a deep-water environment. Therefore, the moderate sized lake that existed during the deposition of the deltaic sands may have lasted less than the 6000 yr that are indicated in the primary chronology (Fig.14).

The fifth subset consists of two tufa samples that were deposited at an altitude of about 2155 m (Hubbs et al., 1965; S. Stine, pers. comm., 1989). Based on the probability of contamination by younger carbon (Lao and Benson, 1988), Lake Russell probably stood at 2155 m long before 22,000 yr B.P.

Comparison of lake-level chronologies

Comparison of lake-level chronologies of the four lake systems (Figs.4, 7, 11, and 14) indicates certain similarities and certain differences:

(1) Lake Bonneville and Lake Lahontan were at low levels about 35,000 yr B.P.

(2) Lake Russell and Lake Searles were at

moderate levels at about 35,000 yr B.P. By 33,000 yr B.P., levels of both lakes had fallen.

(3) A decrease in lake level occurred at about 15,500 yr B.P. in the Lake Lahontan basin and between about 15,250 and 14,500 yr in the Lake Bonneville basin and also may have occurred in the Mono basin. One interpretation of the Searles Lake record (Fig.11a) indicates a decrease in lake level between 15,000 and 12,500 yr B.P.

(4) Highstands occurred before 13,500 yr B.P. in all four basins.

(5) Recession from the last highstand may have occurred nearly synchronously between 14,000 and 13,500 yr B.P. in all four basins.

(6) In the Lake Bonneville and Lake Lahontan basins, the lakes either stabilized at or rose to moderate levels between about 11,500 and 10,000 yr B.P. In the Searles Lake basin, lake levels rose between 12,500 and 11,500 yr B.P. and perhaps again at 10,500 yr B.P. Carbonate samples that may have been deposited during this oscillation in the Mono basin remain undated.

(7) During the last 10,000 yr, lakes in the Bonneville, Lahontan, and Mono Lake basins remained at low levels. However, a 55-m lake may have occupied the Searles Lake basin for a large part of the Holocene (Fig.11a). If a lake this size existed in the Searles Lake basin then the Holocene climate of the lower Owens River drainage area was strikingly different from the climate of the Lake Bonneville, Lake Lahontan and Lake Russell drainage areas.

Differences are evident in the lake-level chronologies (Figs.4, 7, 11, and 14) between 35,000 and 14,000 yr B.P. During this time, Lake Bonneville and Lake Lahontan had one major oscillation (highstand). In contrast, Lake Russell and Lake Searles seem to have had several major oscillations. The cause of more frequent oscillations of the smaller lakes is discussed in the next section.

Surface-area variation and its normalization

The proper gage of lake response to change in the hydrologic balance is neither depth (level) nor volume but instead surface area (Hutchinson, 1957; Mifflin and Wheat, 1979; Smith and Street-Perrott, 1983; Stine, 1987; Benson and Paillet, 1989). This concept can be demonstrated by examining the mass-balance equation for a closed-basin lake in which the flux of ground water is negligible.

$$E_l A_l = P_l A_l + V_d \tag{1}$$

Where E_l = lake evaporation rate; A_l = surface area of the lake; P_l = precipitation on the lake; and V_d = volume of streamflow discharge. Rearrangement of Eq.1 indicates that change in the rates of streamflow discharge and of exchange of water across the air-water interface are manifested in change of lake-surface area.

$$A_l = V_d/(E_l - P_l) \tag{1a}$$

For a given basin, the dependence of lake area, lake volume, or lake depth (lake-surface altitude) is determined by basin topography. Under the same steady-state hydrologic balance, lakes that have different depth–area relations will obtain the same equilibrium surface area. To demonstrate this concept, consider the following thought experiment. Suppose there exist two conical-shaped lake basins that have different apices (topographies) and identical drainages (Fig.16). Assume that identical climatic change occurs over both lake basins; the influx of stream-flow discharge (V_d) becomes 1 km^3/yr, and the exchange of water across the surface of the lake ($E_l - P_l$) becomes

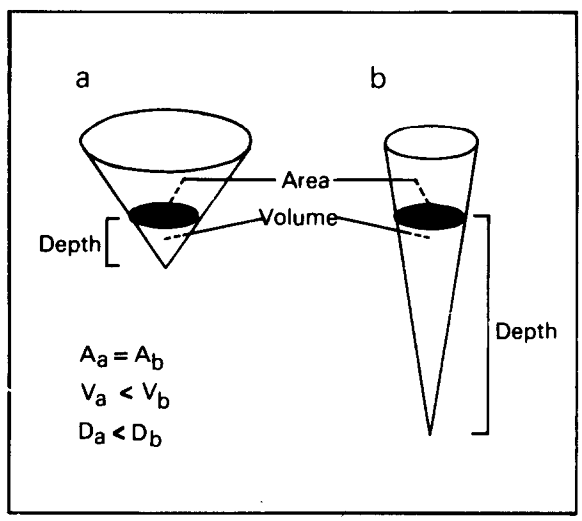

Fig.16. Two hypothetical lake basins that have conical shapes and different apices.

1 m/yr. To satisfy these identical changes in the mass balance, both lakes will expand or contract depending on their initial states. Because in our thought experiment, $(E_l - P_l)$ and V_d are identical, the equilibrium-surface area eventually achieved by each lake (A_l) will be identical (10^3 km^2). Because their basin topographies differ, the two lakes will achieve different depths and different volumes (Fig.16). In addition, the time it takes each lake to reach its equilibrium-surface area will differ because the kinetics of lake-level change is a function of the shape and size of the lake basin (Hutchinson, 1957; Benson and Thompson, 1987a; Stine, 1987). However, if the net flux of ground water across the sediment–water interface becomes significant relative to other components of the hydrologic balance, surface area will no longer be a completely precise measure of climatic change.

Normalization of surface areas is necessary when comparing surface areas of lakes in basins that have different topographies. In this paper, we adopt 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 (Table 5).

Causes of lake-size variation

The surface-area estimates (Table 5) indicate that, during the last highstand, Lake Bonneville and Lake Lahontan surface areas increased in size by about a factor of 10 relative to their mean-historical surface areas, whereas Lake Russell and Lake Searles surface areas increased by about a factor of 4–6. Two hypotheses may explain these differences in the effective wetness of climate during the last highstand: (1) either the mean track of the jetstream passed over Lake Bonneville and Lake Lahontan but passed north of Lake Russell and Lake Searles, or (2) Lake Bonneville and Lake Lahontan continued to grow beyond a factor of 5 because of lake–atmosphere feedback processes such as lake-effect storms. There is no way at this time to assess the relative importance of the two hypotheses.

The principal argument for the second hypothesis is that the surface areas of Lake Bonneville and Lake Lahontan were always much greater than surface areas of Lake Russell and Lake Searles. In modern times, lake-effect storms (Eubank and Brough, 1980; Braham and Dungey, 1984) and other lake–atmosphere feedback processes (Benson and Thompson, 1987b, p. 255) have altered the climate in the vicinity of lakes as large or larger than Great Salt Lake.

Between about 11,500 and 10,000 yr B.P., lake-surface area increased in three of the lake basins by about a factor of 2–3 relative to mean-historical reconstructed surface areas (Table 5). The timing of this change in effective wetness and the relatively uniform increase in surface area of the three lake systems indicates that a different type of climate forcing other than that associated with continental ice sheet size may have been responsible for this oscillation. Benson and Thompson (1987b) and Lao and Benson (1988) have hypothesized that major Great Basin paleolake cycles were primarily associated with the changing mean position of the jetstream. Kutzbach and Wright (1985) and Kutzbach (1987) have used atmospheric general circulation models (AGCMs) to simulate climatic patterns for the past 18,000 yr. These studies indicate that the jetstream moved in response to the size and shape of the continental ice sheet during the last glacial/deglacial cycle; by 11,000–10,000 yr B.P., the effect of the ice sheet on the wind field had diminished and the winter and summer positions of the jetstream were much like those of today. Oscillations in lake level occurring between 11,500 and 10,000 yr B.P. may be associated with the Younger Dryas climatic event that produced cooling over western and central Europe (Jensen, 1938; Iversen, 1954). To the extent that this association is correct, it indicates that climatic change associated with the Younger Dryas climatic event was at least hemispheric in scale. If Lake Searles spilled into the Panamint basin between 11,500 and 10,000 yr B.P. (Fig.11a), then the total surface area of the lake chain in the Owens River drainage basin increased by a factor of 5.5 (Table 5). If Lake Searles did not spill to the Panamint basin 11,000 yr B.P. (Fig.11b), then the surface area of the lake chain in the Owens River drainage increased by more than a factor of 2.7

TABLE V

Surface areas of Lake Bonneville, Lake Lahontan, Lake Russell, and Lake Searles for: the historical period, between 11,500 and 10,000 yr B.P. (as hypothesized in this paper), and for the time of the last highstand (normalized areas are shown in parentheses).

Lake name	Historical surface area (km^2)	11,500–10,000 yr B.P. surface area (km^2)	Highstand surface area (km^2)
Bonneville	5140[1]	17,100 (3.3)[6]	51,300 (10.0)[1]
Lahontan	2420[2]	–	22,300 (9.2)[10]
	1260[3]	2950 (2.3)[6]	–
Russell	190[4]	440 (2.3)[7]	790 (4.2)[11]
Searles	315[5]	1550 (4.9)[8]	1725 (5.5)[12]
		850 (2.7)[9]	

[1]Combined surface areas of present-day Great Salt Lake and Utah Lake (Currey et al., 1984).
[2]Calculated from data of Benson and Thompson (1987b).
[3]Surface area of lake in Carson Desert subbasin; calculated from data of Benson and Thompson (1987b).
[4]Reconstructed present-day surface area of Mono Lake calculated from data of Voorster (1985).
[5]Surface area of Owens Lake calculated from data of Milne (1987).
[6]From Currey (1988).
[7]Calculated from approximate height of lithoid coating (2133 m, Lajoie, 1968) and altitude/surface area curve for Mono Lake basin (Lajoie, unpublished data).
[8]Combined surface areas of Owens Lake, China Lake, and a 200-m, deep Searles Lake.
[9]Combined surface areas of Owens Lake and China Lake.
[10]From Benson and Mifflin (1986) and Benson and Thompson (1987b).
[11]Calculated from height of highest tufa and altitude/surface area curve for Mono Lake basin.
[12]Calculated using cumulative altitude/surface curve for lakes of the Owens River drainage (N. Jannik, pers. comm., 1989) and assuming that Lake Searles was spilling and a 38-m deep lake had formed in the Panamint Lake basin (Smith, 1978).

(Owens Lake barely spilling) and less than a factor of 4.9 (Lake Searles spilling).

The high-frequency record of lake-level change in the Mono Lake basin is due in part to the topography of this basin. Lake levels in such small, relatively steep-sided basins respond rapidly to changes in moisture storage and tend to be excellent gages of low-amplitude high-frequency climatic events (Benson and Thompson, 1987b). The high-frequency record of lake-level change in the Searles Lake basin is due to its position in a chain of lakes. When upstream lake basins were nearly full, relatively small increases in the hydrologic balance of the Owens River drainage area led to large changes in the level of Lake Searles. Lakes formed in the Searles Lake basin only did so during relatively wet climatic periods.

Summary and conclusions

From the study of the chronology and expansion of Lakes Lahontan, Bonneville, Searles, and Russell during the past 35,000 yr, the following conclusions are drawn:

(1) Because most carbon-bearing materials deposited in lake basins only provide some minimum or maximum constraint on lake level, lake-level curves constructed using the radiocarbon ages and altitudes of these materials are, at best, approximations.

(2) Numerous chemical processes can affect the reliability of a radiocarbon-age. Thin sections of carbonate samples should be examined for the presence of carbon-bearing contaminants. Samples contaminated with secondary carbon should not be submitted for radiocarbon-age determination.

(3) To insure the reliability of an assigned age, different types of carbon-bearing material (aragonite gastropods, calcite tufa, disseminated organic carbon, etc.) from the same horizon or site should be submitted for radiocarbon-age determination.

(4) Carbonate samples that have radiocarbon ages in excess of 20,000 yr are often contaminated with small amounts of secondary carbon and should be considered unreliable. Uranium-series age estimates should be obtained on carbonate samples with radiocarbon ages in excess of 20,000 yr.

(5) Lake-level chronologies for Lake Lahontan, Lake Bonneville, Lake Searles, and Lake Russell for the past 35,000 yr can be ranked in terms of reliability.

(a) The lake-level chronology for the western Lake Lahontan subbasins is very reliable. The chronology is based on radiocarbon ages of a variety of carbon-bearing materials (calcite tufas, aragonite gastropods, packrat-midden macrofossils, wood, and rock varnish). In addition, carbonate materials from the Lahontan basin have been examined for the presence of carbon-bearing contaminants. However, the record of lake-level fluctuation in the western subbasins is complicated by the effects of spill across sills that separate the Pyramid Lake subbasin from adjoining subbasins.

(b) The lake-level chronology for Lake Bonneville is highly reliable. Radiocarbon ages of wood samples provide excellent documentation of change in the level of Lake Bonneville from 26,000 to 15,250 yr B.P. and between 13,150 and 10,300 yr B.P. The radiocarbon ages of ostracodes, gastropods, and charcoal which were deposited in association with the Pavant Butte Ash, indicate a major drop in lake level between 15,250 and 14,500 yr B.P. However, the Lake Bonneville chronology is complicated by the Bonneville Flood that downcut the pass at Zenda; i.e., the ensuing 108-m drop in lake level was not forced by climatic change.

(c) The lake-level chronology for Lake Searles is moderately reliable. Radiocarbon ages of some carbonates from sediments that crop out in the Searles Lake basin are stratigraphically inconsistent. The inconsistency indicates that some carbonate samples have been reworked and redeposited; others have been contaminated with secondary carbon. The radiocarbon ages of disseminated organic carbon extracted from the Parting Mud are very reliable except for samples extracted from the upper 0.8% of the unit. The accuracy of radiocarbon ages of disseminated-organic carbon from the Lower Salt remains in question. Different interpretations of the depositional environments (depth of water) of chemical sediments in the Lower Salt, Parting Mud, Upper Salt, and Overburden Mud units do not permit the construction of a unique lake-level chronology. In addition, stratigraphic correlation of core-based mud units with outcropping mud units is complicated by the fact that the core-based units have not been physically traced to outcrop.

(d) The lake-level chronology for Lake Russell is moderately reliable. Most of the dated material from the type section of the Wilson Creek beds has been examined for the presence of contaminants. In addition, the radiocarbon ages of carbonate samples from the type section plot as a simple function of depth in the type section. However, modern carbonate samples that formed during the past several decades have radiocarbon ages that range from about 1100–5300 yr B.P. The magnitude of this reservoir effect is not known as a function of time or lake size, indicating the difficulty inherent in assessing the reliability of radiocarbon ages of carbonate and organic carbon samples from the Mono Lake basin.

(6)During the past 35,000 yr, all four lakes underwent a major period of lake-level change. This period of change includes the following elements:

(a) The lakes were at moderate levels or dry at the beginning of the period and seemed to have achieved highstands between about 15,000 and 13,500 yr B.P.

(b) Lake Russell and Lake Searles had several major oscillations in level between 35,000 and 14,000 yr B.P.

(c) All four lakes may have had nearly synchronous recessions between about 14,000 and 13,500 yr B.P. After the recession, the lakes seem to have temporarily stabilized or experienced a minor increase in size between about 11,500 and 10,000 yr B.P. These data provide circumstantial evidence that the Younger Dryas Event affected climate on at least a hemispheric scale.

(d) During the last 10,000 yr, three of the four lakes (Bonneville, Lahontan, and Russell) maintained small surface areas relative to their highstand areas.

(7) Most workers have discussed variation in lake size in terms of variation in lake level. However, the proper gage of lake response to change in the hydrologic balance is neither lake level (depth) nor lake volume but lake-surface area.

(8) Normalization of surface area is necessary when comparing surface areas of lakes in basins having different topographies. As a first approximation, normalization can be accomplished by dividing the palaeosurface area of a lake by its mean-historical, reconstructed surface area.

(9) A comparison of surface-area data for the four lake systems indicates that, during the last highstand, Lake Lahontan and Lake Bonneville increased in size by a factor of about 10, whereas Lake Russell and Lake Searles increased by a factor of about 4–6. Two hypotheses may explain these differences:

(a) The mean track of the jetstream passed over Lake Lahontan and Lake Bonneville but passed north of Lake Russell and Lake Searles during the last highstand.

(b) Surface areas of Lake Lahontan and Lake Bonneville continued to grow beyond a factor of 4–6 because of lake–atmosphere feedback processes.

Acknowledgments

The authors express their gratitude to a number of colleagues who provided important insight into the processes that govern the biological, chemical, and hydrochemical evolution of closed-basin lakes. These colleagues include: Michael Binford of Harvard University, Wallace Broecker of the Lamont-Doherty Geological Observatory, Fred Phillips of the New Mexico Institute of Mining and Technology, Phillip Meyers of the University of Michigan, and James Bischoff, Richard Forester, Blair Jones, and James LaBaugh of the U.S. Geological Survey.

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