Introduction

More than a hundred closed basins in the western United States contained lakes during the Late Wisconsin, 25,000 to 10,000 yr B.P., but only about 10% of the lakes are perennial and of substantial size today (Figure 10-1). Changes in one or more elements of the climate—precipitation, temperature, evaporation, wind, cloud cover, and humidity—were responsible for most of these dramatic oscillations. The climatically enlarged lakes are referred to by most investigators in the United States as “pluvial lakes,” although the relative importance of increased rainfall (the literal meaning of the word pluvial) versus changes in the other climatic controls is not specifically implied and is still being debated. The large lakes left geomorphic, stratigraphic, paleontologic, and archaeological evidence of their former existence; their careful study, along with evidence from dry phases intervening between times of lake expansion, provides a paleoclimatic record of alternating pluvial and interpluvial periods over an area that represents about a third of the conterminous United States.

Most pluvial lakes formed in basins that are still topographically closed and have internal drainage. The largest group was within the Great Basin area of Nevada, Utah, and California, but clusters of pluvial lakes also existed in the peripheral parts of the Basin and Range province in Oregon, California, Arizona, New Mexico, and northern Mexico. Numerous small lakes also occurred in the southern High Plains of Texas and New Mexico and in some externally drained valleys in California. Most of these basins in the western United States are products of late-Cenozoic tectonic activity, although some are the results of deflation or of damming by lava or landslides. The intensity of these geologic processes in the arid western states explains the concentration of closed basins in this area, especially in the Basin and Range physiographic province. Where climates are not now arid, most closed depressions contain lakes that are constantly filled and overflowing. Former more-intense pluvial conditions produced no increase in these lake levels, and more-arid periods left records that are now concealed by water. Lakes in less-arid areas also tend to have shorter lives; they fill with sediment more rapidly or erode their outlets to the valley-floor level and drain.

The value of fluctuations in lake levels as an indicator of climate has been recognized for more than 250 years (Halley, 1715). In 1776, Velez de Escalante (translation, 1943) speculated that shells near Utah Lake recorded a former large lake, the one now known as Lake Bonneville. Pluvial lakes in the area were first well documented about a century later in a series of classic studies by Whitney (1865), Simpson (1876), King (1878), Russell (1883, 1884, 1885, 1889), and Gilbert (1885, 1890) on the two largest pluvial lakes—Lake Bonneville in Utah, Nevada, and Idaho, and Lake Lahontan in Nevada, California, and Oregon—as well as on smaller lakes in California and Oregon. The pluvial chain of lakes that existed in Owens, China, Searles, Panamint, and Death Valleys in California was later documented by Gale (1914); other former lakes in Oregon, Nevada, California, Arizona, and New Mexico were identified by Waring (1908, 1909), Meinzer (1911), Meinzer and Kelton (1913), Meinzer and Ellis (1915), and Clark and Riddell (1920). The regional paleoclimatic implications of these extinct lakes were summarized by Meinzer (1922), who first compiled a map of pluvial lakes in the Basin and Range province, and the inferred relations between these lakes and glacial events in other parts of the world were discussed by Antevs (1925). Since then, several maps showing the locations of pluvial lakes have been published, mostly with accompanying texts and lists of references; examples are those by Hubbs and Miller (1948), Feth (1961), Snyder and others (1964), Morrison (1965c), and Mifflin and Wheat (1979).

Besides being records of paleoclimate, the datable stratigraphic and geomorphic records found in many pluvial-lake basins provide the means of determining the recency of faulting, the rates of tectonic tilting, and the kinetics of landform development, and of obtaining detailed records of past variations in the positions and polarities of the Earth’s magnetic field. Pluvial-lake shorelines are also favorable sites for finding traces of early human cultures. The derived paleoclimatic records also provide inferential methods of dating or correlating se-
Consequences of other climate-related geologic phenomena, such as glaciation or soil and terrace development.

Factors Affecting Pluvial-Lake Levels

CLIMATIC FACTORS

The amount of precipitation falling on the drainage area of closed basins is the most significant factor affecting the supply of water to many lakes, although the amount falling directly on a lake’s surface can be significant. Substantial differences in the percentage of the tributary’s precipitation that reaches the lake, however, are caused by variation in any of several factors. The most important of these are the seasonal distribution and intensity of the precipitation, the proportions of snow and rain, the nature of the vegetative and soil cover, the mean annual and seasonal temperatures, the topographic relief and slope angle, and the amount of annual variability. Langbein and others (1949: Figure 2), Snyder and Langbein (1962: Figure 3), Dury (1965: C15-C40), Schumm (1965: Table 1), Galloway (1970: 250-52), Brakenridge (1978: 30-34), Mifflin and Wheat (1979: 37-49), and Benson (1981: 394-400) propose ways to estimate the net effect of these variables, but some factors are treated differently or are omitted and the results differ markedly. A major source of uncertainty in these attempts is the fact that in presently arid regions, drainage from areas lying at higher elevations and receiving annually more than about 40 cm of precipitation is required to maintain perennial lakes, and there is no present analogue of the pluvial runoff from the more arid lower sectors. Estimates of past increases in runoff that were
necessary to support maximum-sized pluvial lakes range from about double (Galloway, 1970: Figure 5, Table 1) to an order of magnitude (Dury, 1965: C22; Smith, 1976a: Table 1) (Table 10-2).

One of the most important influences on evaporation from lakes is water temperature, which determines vapor pressure. In the 0°-to-30°C range, water-temperature reductions of 5°C reduce the vapor pressure to about 72% of the original value; reductions of 10°C, about the maximum lowering of air temperatures indicated by most data on Pleistocene paleoclimates in temperate latitudes, produce vapor pressure reductions to about 50% of the original value. However, water temperature is determined by a complex interaction among about a dozen energy sources and sinks, and these vary in magnitude according to the temperature and absolute humidity of the air, the duration and intensity of solar radiation, and the velocity of the wind. Most of these factors become less effective as air temperatures become lower (Kohler, 1954: 132-36; Kohler et al., 1955: 9-19; Harbeck, et al., 1955; 1958; Kohler et al., 1959: 43-50; Meyers, 1962; Helvy et al., 1966: 9-19) and as elevations increase (Blaney, 1956: Figures 1 and 2; Harding, 1965; Mifflin and Wheat, 1979: Figure 24). During cooler pluvial periods, however, the changes in the factors other than the temperature that influence evaporation tend to cancel each other: evaporation (1) is increased as a result of lower salinities and probably higher wind velocities, (2) is possibly decreased as a result of higher humidity and cloudiness, and (3) is partially offset by precipitation on the enlarged lake surface.

In saline lakes, the evaporation rate and the degree of fractionation of stable isotopes during evaporation are reduced in proportion to the percentage of dissolved ions as a result of the hydration energies of those ions (Langbein, 1961; Stewart and Friedman, 1974; Friedman et al., 1976). In saturated brines dominated by sodium and potassium, evaporation rates are reduced to 60% to 80% of those of fresh water (Harbeck, 1955); saturated brines dominated by divalent magnesium and calcium cations, however, can have rates reduced to 10% of fresh water (Turk, 1970). Water color and depth, as well as current velocities and patterns, also affect surface-water temperatures, and evaporation rates are also affected by these factors.

OTHER FACTORS

Nonclimatic phenomena can also change the levels of lakes in closed basins. Tectonic events such as the raising or lowering of catchment areas and outlet sills by faulting or crustal warping can gradually lead to new lake-basin dimensions and hydrologic regimes. Erosion of the outlet barrier can quickly lower or drain a lake in a closed basin, local subaerial erosion of surrounding exposures can fill a lake basin to the level of its sill, and headward erosion of nearby streams can lead to stream capture and to a rapid increase or decrease in the inflow to a lake. Volcanic eruptions and landslides can abruptly dam water channels and temporarily or permanently block or divert flow to downstream lakes.

Most of these processes, however, result in changes in lake levels that are long lasting compared to fluctuations caused by climatic change. The Bonneville Flood, for example, apparently occurred after the capture of the externally flowing Bear River by a tributary of Lake Bonneville as a result of lava damming (Bright, 1967: 4-5); the increased tributary flow caused the lake to rise to a previously unoccupied spillway level, which was eroded so rapidly through unconsolidated sediments down to bedrock that it produced downstream flooding with an estimated discharge of more than 400,000 m³ per second (Malde, 1968). As another example, the initial formation of pluvial Lake Las Vegas 30,000 years ago may have been tectonic (Haynes, 1967: 82), and its draining 14,000 years ago could have been a result of erosion at its outlet, although Mifflin and Wheat (1979: 27) conclude that a spring-fed marsh, rather than a lake, occupied that site. Overall, however, few changes in the levels and histories of lakes in the western United States appear to have been caused by these nonclimatic processes during the Late Wisconsin. Nevertheless, the two examples illustrate why caution must be used in translating pluvial-lake changes, especially the longer and more "permanent" changes, into climatic histories.

Processes of Pluvial-Lake Sedimentation

CLASTIC SEDIMENTATION

The sizes and shapes of pluvial lakes in the western United States were first reconstructed on the basis of the record left by nearshore clastic sedimentation. Very few studies have been made, however, of the shoreline depositional environments of fluctuating pluvial lakes, although a large literature exists on erosion and sedimentation processes along the shorelines of coasts and large stable lakes (e.g., Sly, 1978: 72-76). Gilbert (1855: 98-100, Plate XII) inferred that the V-shaped terraces and bars in the Lake Bonneville area of Utah were a result of the lake's fluctuating character; Born and Ritter (1970) attribute river terraces leading to Pyramid Lake, Nevada, to fluctuating lake levels; and Born (1972: 66-90) describes the late-Quaternary and modern depositional environments in the delta portion of the constantly changing Pyramid Lake.

Clastic sediments in deeper-water and other lower-energy environments of lakes generally consist of well-sorted fragments derived from weathered bedrock in the drainage area. Compared to freshwater-lake sediments, pluvial-lake sediments may contain larger percentages of chemical precipitates (discussed later), and diagenesis commonly is more extensive. However, many of the processes by which sediment is introduced from outside the lake and by which various organic and inorganic components are absorbed or precipitated from the lake water are essentially the same as those in freshwater lakes, and these processes are partially controlled by the specific clastic mineral phases present, especially the clay species (Jones and Bowser, 1978). Sly (1978: 77-83) discusses the interaction between energy source and sediment supply in freshwater lakes. He also notes that lake sediments are finer grained than marine sediments deposited in water of comparable depth because of the lower-energy environments of lakes; lake-sedimentation rates, however, are usually higher. The most diagnostic and frequently used physical, mineralogical, petrographic, and textural characteristics of closed-basin and freshwater lacustrine sediments are summarized by Picard and High (1972). The limnology and the bottom sediments of several present-day Lahontan Basin lakes are described by Hutchinson (1937), and the organic and inorganic composition of Pyramid Lake bottom sediments, by Swain and Meader (1958).

CHEMICAL SEDIMENTATION

Even though pluvial lakes were (by definition) larger than the lakes presently occupying the same closed basins, many had no outlet, and so they accumulated salts and deposited chemical sediments. Carbonate minerals are the most abundant precipitates in sediments deposited in relatively fresh water, although differentiating chemical-
ly precipitated carbonate from detrital, skeletal, and diagenetic carbonates can be difficult (Kelts and Hsu, 1978). Deposition of carbonate minerals from fresh water generally requires decreasing the partial pressure of carbon dioxide as a result of exsolution or removal by microorganisms, increasing the chemical activity of calcium, warming the waters, or mixing chemically dissimilar waters (Jones and Bowser, 1978: 211-15; Smith, 1979: 79-82).

Pluvial lakes that were constantly resupplied with the required components precipitated substantial percentages of aragonite, calcite, protodolomite, or gypsum along with the clastic fraction—in effect, undergoing the first stages of brine evolution (Hardie and Eugster, 1970; Eugster and Hardie, 1978: 243-47). For example, cores of pluvial Lake Bonneville sediments, interpreted as having been deposited during the latest pluvial period in a relatively freshwater environment, contain 18% to 64% total carbonate (Eardley and Gvosdetsky, 1960: Figure 1). In moderately saline, unstratified pluvial lakes, a continuous rain of chemical sediments probably occurred at rates proportional to seasonal changes in the evaporation/inflow balance (Hardie et al., 1978: 23-27). In stratified lakes, denser saline bodies of brine were overlain by lighter layers of less saline water that flowed into the basin and brought new components; these combined, on mixing, with components from the lower layer, and chemical sedimentation rates probably became more proportional to freshwater inflow volumes. Aragonitic laminae in pluvial-lake deposits 24,000 to 10,500 years old from Sears Lake cores are inferred to be products of seasonally replenished stratified perennial lakes; they now contain, on the average, about 55% carbonate minerals as a mixture of primary and diageneric species that includes some carbonate introduced later by downward-moving brines (Smith, 1979: 111, Table 13).

Pore waters of sediments deposited in closed basins are commonly saline, either because of a brine layer along the floor of the lake at the time of initial deposition or because of a subsequent downward migration of denser brines formed during later desiccations. Reaction of these brines with both chemical sediments and clastic silicate sediments commonly leads to the development of one or more diageneric minerals and to the partial or total destruction of some original phases. The chemical principles involved in these reactions are summarized by Berner (1971: Chapters 6-10); examples of Quaternary lake deposits that document these processes are cited by Sheppard and Gude (1968: 32-36), Eugster and Hardie (1978: 252-55), and Smith (1979: 100-106).

Late-Quaternary Geology of Selected Pluvial-Lake Areas

The geologic record left by late-Pleistocene pluvial lakes in the western United States can be divided into two broad categories, geomorphic and stratigraphic. The geomorphic evidence, such as gravel bars and beach deposits, and the erosional shorelines required as the sources of such sediments, are evident in most of the basins that contained lakes. The distinctive shapes, lithologies, and horizontal alignments of these features provided early investigators with nearly indisputable evidence of the former presence of standing water at higher levels. The classic studies of Russell (1885, 1889), Gilbert (1890), and Gale (1914) were based largely on features of these types, although the studies were supplemented by stratigraphic considerations. The more recent shoreline mapping of Nevada’s pluvial lakes by Mifflin and Wheat (1979: 10-37) and the study of Lake Bonneville shoreline zones by Currey (1980) illustrate the extent to which features of these types, used with little stratigraphic corroboration but in conjunction with radiocarbon dating, still provide a reliable basis on which to reconstruct the size, distribution, and contemporaneity of pluvial lakes and from which to infer the climatic regimes that produced them. In many basins, in fact, nearshore features provide the only exposed evidence of former lakes.

Extensive bars, deeply cut shorelines, and thick beach deposits in pluvial-lake basins provide the best-documented portions of the geomorphic record, and they are usually inferred to represent periods of lake stability caused by an escape of excess water via a spillway (which prevented the lake level from fluctuating as short-term variations in the pluvial intensity occurred), by an unusual basin configuration, or by a stable climate. Between these periods of stability, however, periods of instability probably characterized most pluvial lakes, and the less-pronounced nearshore features that developed during those times now provide a correspondingly more obscure geomorphic record of that part of a lake’s history.

The stratigraphic record left by pluvial lakes can be more nearly complete and less ambiguous. However, there are geologic problems. The sedimentary record exposed around the edges of lake basins is commonly fragmentary and many have hiatuses caused by erosion or nondeposition, and the record provided by cores from areas of nearly continuous sedimentation rarely allows the reconstruction of the water depths over the site, unless salts were being deposited, in which case the deposition rate and age estimates based on it become uncertain. To solve these problems, most investigators collect stratigraphic data from numerous settings in order to gain a complete record. But the most difficult aspect of stratigraphic studies based on outcropping lacustrine sediments—and one of the most serious obstacles to the correct reconstruction of pluvial-lake histories from this type of evidence—is the reliable correlation among widely separated outcrops. Outcrops commonly consist of nearshore facies of lacustrine deposits, and the character of the clastic fraction of these sediments varies greatly, even over short distances, making long-distance lithologic correlations very difficult. The most reliable means of making correlations are distinctive volcanic ash layers, but many basins have few or none. Basins with outcrops containing some horizons characterized by abnormally high (or low) percentages of chemical sediments—usually calcite, aragonite, or gypsum—generally can be correlated over long distances, because the abnormal horizons represent a basin-wide chemical anomaly in the lake waters. The soils developed on the exposed lacustrine sediments or overlying alluvium during low stands have been used extensively by some investigators for correlating sections of lake sediments.

The clastic fraction in deep-water facies found in cores commonly shows little stratigraphic change, and correlations over long distances are also difficult. However, as in outcrops, the lithologic variations of perennial-lake deposits caused by changes in water chemistry or lake structure are generally widespread enough to provide a reliable basis for intra-basin correlation and possibly for interbasin correlation if they are caused by regional climatic events. When pluvial lakes shrink, saline layers are deposited in the basin centers, but lateral changes in the mineralogy and texture of saline beds make distant correlations difficult; correlation of the perennial-lake sediments above and below such beds usually provides a better basis.

Reconstructing lake histories from any of these types of data requires reconstructing the lake level through time. Dated geomorphic
records provide reliable histories of the stable periods and elevations of high lake levels. Dated stratigraphic records of outcropping sediments permit the reconstruction of both stable and fluctuating lake periods, and the elevation range of each lacustrine and nonlacustrine stratigraphic unit allows at least minimum estimates of lake-surface elevation change. Studies of cores permit the unequivocal separation of perennial-lake periods from saline- or dry-lake periods. Stratigraphic studies of both cores and outcrops also allow supplemental data to be derived from their chemical and mineralogical nature, textural character, and fossil content.

LAKE BONNEVILLE SYSTEM

Lake Bonneville, which covered about 51,640 km² at its highest stage, was the largest of the late-Pleistocene pluvial lakes in western North America. (Areas and other data on individual lakes are mostly from Mifflin and Wheat [1979, appendix] or Snyder and others [1964].) At its highest stage, it formed an irregular oval body of open water that contained about 20 relatively small islands; its longest dimension, nearly 500 km, was oriented north-south. Water covered the lake floor to a maximum depth of about 335 m, and its total volume was nearly 7500 km³. The weight of this water caused isostatic depression of the crust; and, since the lake’s desiccation about 10,000 yr B.P., the once-level Bonneville shoreline has rebounded isostatically so that the shorelines on islands near the middle of the lake are now about 65 m above the elevation of the shoreline near the edge of the lake (Crittenden, 1965a: 5520, Figures 1 and 2; 1963b).

Lake Bonneville is probably the most studied of the pluvial lakes in western North America, and so its history from 25,000 to 10,000 yr B.P. should be the best known, but substantial differences in the details of interpretation still exist. Gilbert’s classic study (1890) expressed the lake’s history in terms of shoreline levels, and he named the three most prominent benches as the Bonneville (1565 m elevation), Provo (1470 m), and Stansbury (1350 m) levels. (The elevations are nominal; they vary from place to place because of the differential isostatic rebound that followed evaporation of the lake water.) Ives (1951) named two lower shorelines as the Timpie (1295 m) and the Dunway (1320 m); Eardley and others (1957) renamed the Timpie as the Gilbert level. All these shorelines are considered by Currey (1980: Figure 8) to have been occupied within the time of interest here. Through the use of stratigraphic and radiometric dating methods, other reconstructions of Lake Bonneville’s history have been made by Antevs (1945, 1948, 1952, 1955), Ives (1951), Williams (1962), Hunt and others (1953), Eardley and others (1957, 1973), Broecker and Orr (1958), Bissell (1963), Morrison (1961b, 1965a, 1965b), Richmond (1961), Broecker and Kaufman (1965), and Scott and others (1980). Morrison (1965c: Figure 2) summarizes his own (1965a) and seven preceding interpretations of Lake Bonneville’s history. Figure 10-2 shows Morrison and Frye’s (1965) interpretation, plus four others that are not included in Morrison’s summary or have been published since. Two of the lake histories illustrated in Figure 10-2 have age controls based on radiocarbon dates, one on both radiocarbon and amino acid dating methods, and one on geomorphic methods. Much of the age discrepancy among the histories is attributable to the geochemical problems associated with radiocarbon dating. (See the section on Lake Lahontan.)

As might be expected with this many investigators studying Lake Bonneville over a period encompassing a century, the history of the nomenclature for the shorelines and the sediments deposited in Lake Bonneville has become complex. Early investigators based their conclusions on combinations of geomorphic and stratigraphic criteria, and usages developed in which shore features and the sediments inferred to be their offshore equivalents were assigned ages and stratigraphic positions more on the basis of their elevation within the basin than on conventional stratigraphic criteria, such as superposition or lateral correlation of exposed sections. Morrison (1965c: 273-76) summarizes his own and previous usages; Scott (1980) presents an updated review and a significantly different version of the stratigraphic record. For brevity, no attempt is made here to explain the interrelation among the stratigraphic units defined or used by the various investigators; details about the past and present usage of the terms “Alpine Formation,” “Bonneville Formation,” “Draper Formation” and “Provo Formation,” which appear to include sediments deposited partially or entirely within the time period here under discussion, can be found elsewhere (International Union for Quaternary Research, 1977: 89-94).

Figure 10-2. Interpretations of the history of Lake Bonneville during the period 30,000 yr B.P. to present. (A modified from Broecker and Orr, 1958: Figure 9; B modified from Broecker and Kaufman, 1965: Figure 4; C modified from Morrison and Frye, 1965: Figure 2; D modified from Scott et al., 1980; and E modified from Currey, 1980: Figure 8.)
Recent studies by Currey (1980) and Scott and others (1980) probably come the closest to reconstructing a correct history for Lake Bonneville over the period 25,000 to 10,000 yr B.P., because they use the largest number of stratigraphically controlled radiocarbon dates and other methods that serve as tests of those ages. Scott’s (1980) review of the stratigraphic complexities in the basin, however, illustrates the difficulties involved in compiling dates from widely spaced localities to create a lake chronology; isolated sections of lake deposits must be correlated by criteria that require assumptions, depositional environments are sometimes difficult to reconstruct with certainty, and rapid facies changes in nearshore environments make even short-range stratigraphic correlations in these areas uncertain.

When Lake Bonneville fell from the Bonneville level at 1565 m to the Provo level at 1470 m, the Escalante Desert, Rush Valley, and White Valley arms were detached, and the lakes in those arid valleys probably shrank or desiccated. Continuing recession past levels below the Stansbury level at 1350 m detached the arms of the lakes in Sevier Valley, Provo Valley, and Cache Valley. These three basins, however, continued to receive water from the perennially large Sevier, Provo, and Bear Rivers; and, during all except interpluvial periods, one or all of the basins were presumably full, overflowing, and contributing their excess water to the lower lake. Evaporation from their surfaces prior to overflow, however, effectively acted on the overall hydrologic budget as if they were all part of the same lake. The fall of Lake Bonneville below 1286 m disconnected the Great Salt Lake Desert arm; this represented 48% of the evaporated area of the full lake (Eardley et al., 1957; Figure 1), and without major inflow the arm desiccated and formed the Bonneville Salt Flats.

The area of present-day Great Salt Lake at 1280 m elevation is about 8290 km², and Utah Lake covers about 345 km²; their combined area is about a 17th of the maximum size of Lake Bonneville. It appears from this that the net effects of changes in the inflow rates, precipitation on the surface, and evaporation rates during pluvial periods promoted lakes that were at least 17 times larger than lakes permitted by the present climate. This figure may be too low in that it does not take into account the water volume lost by the enlarged lake from its overflow or the reduced evaporation rate caused by the present lake’s salinity. As a measure of relative inflow volumes, it is clearly too high because it does not take into account the precipitation directly on the increased surface area of the lake, which could offset 25% to 35% of the evaporation from it. Even if these considerations perfectly offset each other, this figure would not imply a seventeen-fold increase in the discharges of the rivers that drain from the Wasatch Range into present-day Great Salt Lake. Proposed evaporation rates during pluvial periods range from 10% to 45% below present rates (Table 10-1). With the largest estimate of evaporation reduction and with precipitation on the lake offsetting 35% of the remaining evaporation, inflow requirements would be reduced to about 36% of apparent volumes, and this would reduce the input requirements to about 6 times present river volumes. Also, during the Pleistocene, other rivers in the drainage area probably flowed into Lake Bonneville. There is no quantitative way to reconstruct the pluvial flow of the now-dry tributaries, but present runoff from arid areas north, west, and south of the former Lake Bonneville are about 5% or less of the runoff from high areas of the Wasatch Range (Langbein et al., 1949: Plate 1). The areas drained along those rela-

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<th>Table 10-1. Published Estimates of Full-Glacial Precipitation and Evaporation over Closed-Basin Lakes</th>
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<td><strong>Paleolake or Area</strong></td>
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<td>Estancia, New Mexico</td>
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<td>Brakenridge, 1978a</td>
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<td>Galloway, 1970</td>
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<td>Leopold, 1951</td>
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<td>Warner, Oregon</td>
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<td>Luhontan, Nevada</td>
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<td>Broecker and Orr, 1958</td>
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<td>Mifflin and Wheat, 1979c</td>
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<td>Spring Valley, Nevada</td>
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<td>Brakenridge, 1978a</td>
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<td>Nevada (entire state)</td>
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<td>Southern High Plains, Texas</td>
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<td>Reeves, 1966a</td>
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Note: ΔTs and ΔTs are the changes in annual and summer temperatures, respectively; ΔP is the change in annual precipitation; ΔE the change in annual evaporation.

a Recalculated.
b Average of two values from Weide’s Table 1.
c Calculated from Figure 24, Table 8, and p. 44.
d Calculated from p. 43.
tively arid shores were not large, however, and it seems unlikely that Pleistocene runoff from them would have constituted more than 10% of the total lake inflow. Thus, these considerations imply that the pluvial-age rivers draining the Wasatch Range had flow volumes at least 5 times their present volumes, and possibly much greater.

LAKE LAHONTAN SYSTEM

Lake Lahontan was the second-largest pluvial lake in the western United States. In contrast to Lake Bonneville, it was more an interconnected series of long, narrow lakes than an open body of water. Its 22,900-km² area and 280-m maximum depth made it slightly less than half as large as Lake Bonneville, and the geometry of its basins and its multiple sources of water made it different in several hydrologic aspects. Its highest shoreline averaged 1330 m elevation.

On shrinking, Lake Lahontan was fragmented into nine smaller basins (Benson, 1978: Figure 1); five of these sustain lakes or marshy areas today, and they probably existed independently as lakes of significant size during pluvial transitions and semipluvial periods.

Evidence of pluvial Lake Lahontan was first noted by a geologist with the Simpson expedition in 1858-1859, and the lake was later mapped by King (1878) and named for Baron LaHontan, an explorer. The classic study by Russell (1885) set new standards for geologic investigations of pluvial lakes by integrating data from stratigraphy, geomorphology, and water chemistry, and it used tufa morphology as a supplementary criterion. Russell identified a "lower lacustral clay" and an "upper lacustral clay" separated by a "medial gravel." Morrison (1965c: Figure 4) summarizes the lake histories inferred by Russell (1885), Antevs (1943, 1948, 1952), and himself (1961a, 1964) and presents a revised history of the lake without absolute age assignments. He names the lacustrine deposits the Eeta Formation (oldest), Seehoo Formation, and Fallon Formation; these were separated by subaerial deposits of the Wyemaha Formation and Turupah Formation. Morrison and Frye's (1965) history, which has a time scale based on radiocarbon ages, assigns the Seehoo Formation to the period 25,000 to 10,000 yr B.P. Figure 10-3 shows Morrison and Frye's chronology, along with those from four other studies based on ¹⁴C dates on tufa, shells, marl, organic material, dung, and wood; on ²³⁳Th/²³⁴U dates on shells; and on tephrachronology. As with the history of Lake Bonneville, there remain substantial differences among interpretations of Lake Lahontan's history for the period between 25,000 and 10,000 yr B.P.

The first concerted attempt to assign an absolute age scale to Lake Lahontan's (and Lake Bonneville's) history is Broecker and Orr's (1958). They consider the problems caused by the probable initial disequilibrium between the ¹⁴C in the lake and the atmosphere, and they conclude (p. 1012) that a maximum of 500 years should be added to the quoted ages. They consider the postdepositional contamination of carbonate by diffusion with carbon dioxide in the air (pp. 1012-15), but the possibility of contamination by solution and redeposition is only noted. A continuation of that study done by Broecker and Kaufman (1965) includes about 80 new ¹⁴C ages, explores further the problems of contamination, assesses the magnitude of those errors by several methods, and includes 23 new dates based on ²³³Th/²³⁴U ratios in shells and other materials. Their published ¹⁴C ages and lake-level chronology (Figure 10-3A) include a 500-year correction of the laboratory age. As shown by Morrison and Frye (1965: Figure 4), however, a number of ¹⁴C dates from the Lake Lahontan area appear to be in conflict with one another, and a detailed history cannot be easily reconstructed from the radiocarbon data alone.

Broecker and Walton (1959) find that the carbon in most modern lakes in the Great Basin is "old" relative to that of the carbon dioxide in the air and that the size of the discrepancy is related to the lake's area (which represents the potential exchange area) and its accumulated carbonate content (which represents the amount that needs to be exchanged). They believe (p. 33) that only freshwater lakes in drainage areas composed chiefly of siliceous rocks are likely to have dissolved carbon dioxide with radiocarbon values close to those of the atmosphere and that the water in lakes with high levels of dissolved carbonate—caused by carbonate rocks in their drainage area or by evaporative concentration—could have ¹⁴C concentrations as low as 50% of those of the atmosphere and thus have apparent ages up to almost 6000 years. They construct a method of estimating the size of this error in a given lake (pp. 30-32) and conclude that the ¹⁴C ages of samples of lacustrine materials from Lake Bonneville and Lake Lahontan should be reduced by 500 years. Their method for this calculation, however, shows that the magnitude of disequilibrium between the ¹⁴C content of the lake water and the air is inversely pro-

![Figure 10-3. Interpretations of the history of Lake Lahontan during the period 30,000 yr B.P. to present. (A) modified from Broecker and Kaufman, 1965: Figure 4; B modified from Morrison and Frye, 1965: Figure 2; C modified from Benson, 1978: Figure 6; D modified from Benson, 1978: Figure 5; and E modified from Davis, 1978: Figure 3).](image-url)
portional to the volume/area ratio of the lake, among other factors. With this ratio constantly changing as a lake level fluctuates and with no allowance for the possibility of the existence of stratified lakes (making the effective volume to be exchanged smaller), application of a single correction factor to all dates from one lake, regardless of the lake's level and size, leads to systematically distorted versions of the lake's history.

Benson (1978) shows that with careful attention to petrographic details a case can be made for the use of several types of dense tufa for reliable radiocarbon dating. He discusses the problem of the "age" of the sample when deposited, caused by the initial non-equilibration of $^{14}C$ between the atmosphere and the lake water, and problems of later contamination. He also notes that $^{14}C$ is preferentially incorporated in tufa and that $^{14}C$ is also enriched, making the sample "too young," and he notes that several factors tend to offset each other. Benson then reconstructs the history of Lake Lahontan in the Walker Basin (Figure 10-3C) and the Pyramid Basin (Figure 10-3D) on the basis of radiocarbon dates on dense lithoid tufa, and he concludes (1) that one or more high stands uniting all nine of the Lake Lahontan basins occurred between the 25,000 and 21,500 yr B.P. and about 13,600 and 11,000 yr B.P. and (2) that an intermediate-sized lake, which connected Pyramid Lake with other basins but not with the lake in Walker Basin, existed between 21,500 to 15,000 yr B.P. Tufa apparently was not forming between 15,000 and 13,600 yr B.P., but Benson offers no explanation for this finding. Possibly, the lake level fell during this period, as inferred in Seaver's Valley, California. Indirect evidence suggests that Walker Lake desiccated from about 9000 to 5400 yr B.P. (Benson, 1978: 314) and that Pyramid Lake did not. Born's (1972) study of delta growth in the Pyramid Basin (Figure 10-3D) includes data derived from radiocarbon dates on wood that show that the Pyramid Lake arm of Lake Lahontan has not risen above the 1180 m level since 9700 yr B.P.

Using tephrochronology for correlations and radiocarbon dates on wood and muck, Davis (1978) proposes a different time scale for Morrison's (1964) stratigraphic study, upon which Morrison and Frye's (1965) history is based (Figure 10-3E). Davis places the base of the Seeho Formation, which represents the younger of the two most widely preserved series of lake deposits, at an age greater than 36,000 yr B.P. In this respect, his history does not appear compatible with the others. He expresses some reservations concerning the tephrostratigraphic isolation of the units (p. 60) that move the age of the earliest deposits of the Seeho Formation that far back in time, but he has good grounds for his tephrostratigraphic age assignments. His age assignments rest largely on the correct identification of the Seeho Formation and Wyemaha Formation in two sections measured by Morrison (1964: 122) and on Davis's (1978: 80) correlation of them with the Pyramid Island locality, where wood for the two radiocarbon dates was collected. At all three localities, the Churchill Soil, which is the most frequently used basis for identifying the interpluvial Wyemaha Formation, is missing, and so it appears that the stratigraphic correlations need to be reexamined.

Considerable differences remain among the reconstructed Lake Lahontan histories. The problems facing investigators in this area are similar to those of investigators studying Lake Bonneville. Using the same reasoning, therefore, we consider the more recent age-controlled studies in the Lake Lahontan area by Born (1972) and Benson (1978) as well as most of the study by Davis (1978) to be more probably correct because they are based on more stratigraphic and chronometric data than earlier studies.

During the late 19th century, the perennial lakes in the region once covered by Lake Lahontan had a total area of about 1300 km²; the areas of individual lakes were as follows (Harris, 1970: Table 1; Rush, 1970; Figure 4; Russell, 1885: 55-81): Pyramid Lake, 495 km²; Walker Lake, 283 km²; Winnemucca Lake, 236 km²; Honey Lake, 253 km²; Humboldt Lake, 52 km²; and Soda Lake, 1 km². This is about an 18th of the surface area of Lake Lahontan at its highest stage, and it shows that climates during pluvial maxima promoted lakes that were about 18 times larger than the present-day lakes. For the reasons explained in the discussion of Lake Bonneville, however, lower evaporation rates and an increased number of tributaries reduced the volumes of water required from the main tributaries during pluvial periods. The volume of precipitation falling on the enlarged lake, which did not overflow, would have substantially reduced it further. From the same assumptions used in the Lake Bonneville estimate, it appears that maximum volumes of flow in the main rivers leading in pluvial Lake Lahontan were at least 6 times those of the present.

**OWENS RIVER SYSTEM**

The Owens River system consisted of a chain of pluvial lakes occupying a succession of closed basins in southeastern California. The Owens River, which drains the eastern side of the central Sierra Nevada, supplied most of the water to those lakes, although at times Mono Lake became the northernmost lake in the chain when it overflowed and added its surplus water to the headwaters of the Owens

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*Figure 10-4. Diagrammatic cross section of pluvial lakes downstream from the Owens River showing elevations of floors of basin and lake surfaces during high stands. (Modified from Gale, 1914: Plate VII)*
River. Figure 10-4 shows a diagrammatic cross section of the five lakes downstream from the Owens River in that system.

The geology of pluvial Mono Lake was first studied in detail by Russell, who described it in an 1899 report devoted to the present and past lakes in the basin and to the glacial history of the adjacent Sierra Nevada. He noted that at least two episodes of glaciation occurred in the mountains (p. 363) and that shorelines from the latest high stand of the lake were carved on the lowest of the terminal moraines extending into the pluvial lake area (pp. 369-71). From this relation, he deduced that there were two glacial episodes, which he correlated with the two lacustrine units in Lake Lahontan, separated by an interglacial and interlacustral period (pp. 370-71).

Putnam (1950) proposed that the pluvial lake in Mono Basin be named Lake Russell. He correlated the spacing and sizes of the youngest recessional moraines with the spacing and widths of the shorelines cut as the lake last receded, and he concluded that the glaciers and lake receded according to the same timetable. More recently, however, Lajoie (1968) has shown that variations in shoreline widths are more related to variations in the slopes of the surfaces on which they were cut. He (1968; personal communication, 1981) also has made a stratigraphic study of the lake sediments controlled by a tephrochronologic framework and radiocarbon dates. Eighteen radiocarbon dates on tufa and ostracodes from the lacustrine sediments of the last major pluvial cycle exposed in Mono Basin range in age from 34,900 to 12,800 yr B.P.; his 1981 reconstruction of the history of Lake Russell, revised from his 1968 version, is plotted in Figure 10-5A. He also notes that the lack of saline layers in drill holes that extend deeper than the Bishop Tuff show that this lake has not dried during the last 700,000 years.

Lake Russell overflowed at its highest level into Adobe Valley. There, a small lake 25 m deep formed and overflowed eastward into a valley leading southward to the Owens River. Lake Adobe appears to have been shallow at 11,000 yr B.P., but it increased in depth gradually until about 8000 yr B.P., when it shrank to a shallow marsh; about 4400 yr B.P., it again expanded briefly and formed a small lake (Batchelder, 1970).

The Owens River drains an area of about 8500 km², but virtually all of its present flow comes from the 16% of its drainage area that lies on the eastern slope of the Sierra Nevada (Lee, 1912: 9). Until 1912, the Owens River terminated in a saline lake, Owens Lake, which was about 10 m deep and 290 km² in area before agricultural irrigation in the area became extensive. All of the river's water was diverted to Los Angeles in 1912, and the lake desiccated. The lack of buried salines in the upper 280 m of lake fill (Smith and Pratt, 1957: 5-14) suggests that the lake had not desiccated naturally for the last several hundred thousand years. During pluvial periods, however, Lake Owens overflowed and carried its dissolved components downstream. The present quantity of salines in Owens Lake appears to represent only about 2000 years of accumulation, suggesting that it last overflowed about 2000 years ago (Smith, 1976a: p. 99).

The succession of pluvial lakes in Owens Valley and downstream to Panamint Valley was first adequately documented by Gale (1914; see Smith, 1979, pp. 67-68, for a more detailed account of this and earlier work). Gale's study was primarily geomorphic, although he also described the mineralogy, chemistry, and subsurface stratigraphy of Searles Lake, which at that time was being developed into a valuable source of industrial and agricultural chemicals. Except for descriptions of cores (Smith and Pratt, 1957: 5-25), stratigraphic studies of the lacustrine deposits in Owens Lake and China Lake have not been published.

Stratigraphic studies of the subsurface lake deposits in Searles Valley have been published by Smith and Pratt (1957: 25-51), Flint and Gale (1958), Haines (1959), and Smith (1962, 1979: 8-68). Dates on these units, based on the radiocarbon ratios in organic carbon, carbonate minerals, and wood, are listed by Flint and Gale (1958: 704-5), Stuiver (1964: Tables 1-3), and Stuiver and Smith (1979: Table 19, Figure 30). Stratigraphic studies of the extensively exposed lacustrine sediments in Searles Lake have been made, but only preliminary accounts of that work have been published (Smith, 1966: 170-77; 1968: 297-306; 1976b: 591-94; and two informal guidebooks). The pluvial history of Searles Lake, however, has been reconstructed in some detail for the period between 25,000 and 10,000 yr B.P. The subsurface record of interbedded salt layers and mud layers provides clear and well-dated evidence of interpluvial saline or dry lakes that were shallow and pluvial freshwater lakes that were deeper. The correlative outcropping lacustrine sediments document the elevations reached by the enlarged pluvial-lake surfaces; alluvium, fossil soils, and erosional unconformities correlate with the shallow-lake or dry periods. The inferred history of Searles Lake, based on published and unpublished data, is plotted in Figure 10-5B.

Shorelines and scattered outcrops of pluvial-lake sediments in Panamint Valley were noted by Gale (1914: 312-17) but were not studied in detail until the more recent work of R. Smith (1972, 1975, 1978a, 1978b). Smith considers most of the deposits and shoreline to be older than the period of interest in this chapter. The lowest shorelines, considered by both Blackwelder (1954: 37) and Smith (1978a: 19) to represent the most recent lakes, are only 44 m above the lowest point in the valley and 248 m below its spillway. Their well-preserved character and indirect evidence implies a very late Pleistocene age; local runoff is not adequate to sustain a lake of that size without the inclusion of overflow from Searles Lake (Smith, 1978a; 1978b: 19), and that last occurred during very late Pleistocene time (Figure 10-5B). The northern part of Panamint Valley also contained a small

![Figure 10-5](image-url)

*Figure 10-5.* Interpretations of the histories of (A) Lake Russell in Mono Basin (from K. Lajoie, personal communication, 1981), and (B) Searles Lake (from G. I. Smith, unpublished data) during the period 30,000 yr B.P. to present.
PLUVIAL LAKES OF THE WESTERN UNITED STATES

(18-40 km²) lake that radiocarbon dates show existed at least up to about 10,000 yr B.P. (Peterson, 1980: 176-78). Drill-hole data (Smith and Pratt, 1957: 51-62) do not resolve any details of the late-Pleistocene pluvial period in Panamint Valley, but they confirm the existence of a succession of perennial lakes during that and earlier parts of the valley’s history.

The existence of a pluvial lake in Death Valley was presumed long before shoreline evidence was discovered. However, this is not surprising since the sediment and erosion record is meager, and, as noted by Hunt and Mabey (1966: A71), “Whatever the cause, this California lake left one of the least distinct and most incomplete records of any Pleistocene lake in the Great Basin—another California superlative!” The evidence that does exist consists of erosional shorelines, scattered lakeshore gravels, tufa, and core samples of lake deposits recovered from the deeper valley fill (Blackwelder, 1935; Hunt and Mabey, 1966: A69-A76; Hooke, 1971; Hooke, 1972: 2086, 2092). Hooke (1972: 2086-87) proposes a lake 90 m deep in Death Valley between 11,000 and 10,000 yr B.P. on the basis of radiocarbon dates and a hypothesis involving tectonic tilting and alluvial-fan growth. The radiocarbon dates on lacustrine silt and clay, which range from about 11,900 to 21,500 yr B.P., confirm the existence of a lake in the valley during that period, but the paucity of outcropping lacustrine sediments and fresh shoreline features, which are almost uniformly found in other nearby basins with lakes of the same general age, suggests that the lakes of these ages in Death Valley were small and substantially less than 90 m deep.

The present and past balances between evaporation and inflow require each basin in the Owens River chain to be considered separately. The calculations presented in Table 10-2 assume that reductions in air temperatures of 5° and 10°C led to reductions in lake-water temperatures by the same amount and that evaporation rates were reduced to 72% and 50% of their present values (the approximate change in vapor pressures over these temperature ranges). They also assume that precipitation falling on the lakes was not significantly greater than it is at present, because more than 60% of the cumulative lake area lies in substantially more arid basins than represented by the main tributary region. This is a highly simplified approach to a complex problem, but the lake areas are probably within 10% and the inferred evaporation rates within 40% of being correct, meaning that the calculated inflow/evaporation volumes should not differ by more than 50% from the correct value, whereas the calculated changes in required inflow volumes are several hundred percent of present volumes. Inflow from areas downstream from the Owens River is not considered, and it may have compensated for part of the recharge requirement attributed in Table 10-2 to the Owens River, but present runoff from most of the downstream drainage is about 1% of that coming from the eastern Sierra Nevada to the Owens River itself (Langbein et al., 1949: Plate 1). This fact suggests that pluvial runoff volumes to downstream drainages were also relatively small. It appears likely, therefore, that during the period 25,000 to 10,000 yr B.P., when Owens, China, and Searles Lakes were full and Panamint Lake was small (Table 10-1), the flow in the Owens River and its tributaries was at least 3.5 times its present flow. This method of estimating water volumes (Table 10-2) suggests that when all of the valleys in the system were full and a maximum-sized lake existed in Death Valley—the “maximum pluvial” episode in this area—the Owens River’s flow was nearly 6 times that of the present river if pluvial lake temperatures were about 10°C cooler, and was 8 times that of the present if they were 5°C cooler, and was 11 times greater if little or no cooling accompanied those episodes.

OTHER PLUVIAL LAKES IN THE WESTERN UNITED STATES

As is evident from Figure 10-1, many other closed basins in the western United States contained pluvial lakes. Their geologic record is of uneven quality, and about two-thirds of them have not received careful study. Table 10-3 lists 31 dated lacustrine deposits showing pluvial expansions during the period 25,000 to 10,000 yr B.P. The locations of the deposits are indicated on Figure 10-1; lakes not numbered on the figure can be identified on the previously cited larger-scale maps of pluvial lakes. Discussions of selected lake histories in these basins follow. Those selected have the best records, the most dates, and/or the greatest significance.

Arizona

Only one major pluvial lake existed within Arizona, Lake Cochise in the valley now containing Willcox Playa. Beach ridges lie 26 m above the present playa, and lacustrine sediments below the playa’s surface are dark colored and have negative Eh values, documenting the former existence of pluvial lakes (Martin, 1963: 439, Figure 3; Schreiber et al., 1972: 133, 176). Radiocarbon dates on exposed and subsurface carbon and carbonate materials suggest that high-level pluvial lakes occupied the basin during much of the period from before 30,000 until 13,000 yr B.P. and that lower-level lakes existed between 11,500 and 10,500 yr B.P. (Schreiber et al., 1972: 134, 176).

The almost complete lack of pollen in the top 2.1 m of lake sediment, which have a radiocarbon age of about 25,200 yr B.P. at their base, precludes the reconstruction of the surrounding vegetation during most of the pluvial episodes of the last 25,000 years (Martin, 1963: 440-44). Pine pollen in the underlying 1.6 m of sediments below the dated material accounts for 95% to 99% of the total count; this fact reflects a pluvial interval whose intensity is not exceeded in any lower zone in the 43-m core and indicates that only one maximum-intensity pluvial episode occurred in this area during the period represented by the core. Martin and Mosimann (1965: 352-56) note, however, that there appears to have been marked climatic metastability in this area, as indicated by detailed study of the pollen from the Wisconsin-age sediments in this core.

California

Several closed basins in California contained pluvial lakes besides those that were part of the Lake Lahontan and Owens River systems. However, dates are available for only five of them, and stratigraphic control on those dates is available for only two. The existence of pluvial Lake Tulare in the southern San Joaquin Valley is based on a complex subsurface stratigraphy of lacustrine and nonlacustrine deposits that have yielded five radiocarbon dates ranging from about 26,800 to 9000 yr B.P. (Croft, 1968). The deposits underlie an area now characterized by closed drainage caused by tectonic warping, and the past history of lacustrine deposition may have been as dependent on tectonic processes as on pluvial climates. Nevertheless, perennial lakes appear to have occupied the area intermittently during a period from about 24,000 to 8000 yr B.P. Croft (1968: Figure 4) infers a period of nonlacustrine deposition that ended about 12,500 yr B.P., although he notes that the data allow this nonlacustrine interval to have begun as early as 17,000 yr B.P. or as late as 14,000 yr B.P. and
Table 10-2
Relative Inflow Volumes Required to Balance Evaporation at Varying Temperatures
and for Various Lengths of the Owens River System of Pluvial Lakes

<table>
<thead>
<tr>
<th>Size of Last Lake in Chain</th>
<th>Elevation of Water Surface (m)</th>
<th>Added (and Cumulative) Area (km²)</th>
<th>Evaporation (m/yr)</th>
<th>Cumulative Volume (10⁶ m³)</th>
<th>Relative Discharge of Owens River and Its Tributaries (m³/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>T = 0°C</td>
<td>T = 5°C</td>
<td>T = 10°C</td>
</tr>
<tr>
<td>Owens Lake, historic</td>
<td>1095</td>
<td>290</td>
<td>1.27⁰e</td>
<td>0.41f</td>
<td>1.0</td>
</tr>
<tr>
<td>Owens Lake, abnormal³</td>
<td>1085</td>
<td>—</td>
<td>0.85</td>
<td>0.74b</td>
<td>1.8</td>
</tr>
<tr>
<td>Lake Owens, full</td>
<td>1145</td>
<td>694</td>
<td>0.89</td>
<td>0.62</td>
<td>2.1</td>
</tr>
<tr>
<td>China Lake, full</td>
<td>665</td>
<td>155</td>
<td>0.70</td>
<td>0.78</td>
<td>2.6</td>
</tr>
<tr>
<td>Searles, Lake full</td>
<td>600</td>
<td>839</td>
<td>0.82</td>
<td>1.78</td>
<td>6.0</td>
</tr>
<tr>
<td>Lake Panamint, small³</td>
<td>355d</td>
<td>175</td>
<td>0.90</td>
<td>2.01</td>
<td>6.7</td>
</tr>
<tr>
<td>Lake Panamint, full</td>
<td>602</td>
<td>707</td>
<td>1.19</td>
<td>3.62</td>
<td>8.8</td>
</tr>
<tr>
<td>Lake Manly, half³</td>
<td>87</td>
<td>535</td>
<td>1.97</td>
<td>3.38</td>
<td>11.4</td>
</tr>
</tbody>
</table>

a Surface of last lake in chain.
b Net evaporation rate for last lake in chain; except for Owens Lake, rates for present evaporation (T = 0°C) adapted from Meyers’ data on present gross annual rates (1962: Plate 3), which indicate approximately 1.52 m for China, 1.78 m for Searles, 1.93 m for Panamint, and 2.13 m for Death Valley, reduced to net rates by assuming 10 cm annual precipitation on the lakes in China, Searles, and Panamint Valleys, 5 cm on lake in Death Valley; effects of reducing lakes’ water temperatures by 5°C and 10°C (T = 5°C and T = 10°C) calculated by use of factors 0.72 and 0.50, the approximate reduction in vapor pressure of water with 5°C and 10°C reductions in the range of 0°C to 30°C; corrected for changes in pluvial lake surface elevations above present valley floors using lapse rate of 6.5°C/1000 m (0.64 m/ym/1000 m).

c Except as noted, figures are sums of losses from each lake in chain, the product of (evaporation rate for that lake) × (area of that lake), and values represent volumes of both total evaporation loss and offsetting total Owens River inflow.

d Relative to calculated volume of present Owens River, 0.41 × 10⁶ m³/yr.
e Average of net rates observed in May 1939 through April 1940 (Dub, 1947: Table 3) and September 1969 through August 1970 (Friedman et al., 1976: Figure 1), both corrected for salinity effect by assuming coefficient = 0.9.
 ff Calculated conditions in 1872, prior to irrigation, from data of Gale (1914: 254, 255, 261), calculated by using ratio of lake size to river flow in 1909-1912 and determining flow necessary to balance lake size in 1872.
 g 1968-1969, record wet season (Friedman et al., 1976: Figure 1).
h Flow in Owens River observed in 1969 (Friedman et al., 1976: 505) plus volume diverted into Owens Valley Aqueduct during same period.
i Latest Pleistocene lake, 44 m deep (Smith, 1978a).
 j This "maximum pluvial" condition apparently did not occur during the period 25,000 to 10,000 yr B.P. Calculation uses Death Valley evaporation rate for highest shoreline level but assumes that the volume of water required from the Owens River system was only one-third of total because the Mojave River and Amargosa River systems also contributed their flow to the lake in the valley.

A stratigraphically controlled history of lake-level fluctuations in the Silver Lake area, a shallow basin near the northern end of the larger pluvial Lake Mojave, includes radiocarbon dates on shells and tufa recording three separate pluvial stages in the late Pleistocene. A major pluvial episode ended about 14,500 yr B.P., and short pluvial episodes caused overflow into Death Valley between 13,750 and 12,000 and between 11,000 and about 9000 yr B.P. (Ore and Warren, 1971: Figure 7). An episode of partial filling can be inferred from a radiocarbon date on tufa between 8500 and 7500 yr B.P., and the consistency between the other dates on tufa and the dates on shells supports the acceptance of this date and the existence of the episode. However, the basin has been partially filled to depths of 2 to 3 m three times during the 20th century, so the climatic significance of the last stratigraphically dated flooding may not be major.

The five smaller pluvial lakes shown in Figure 10-1 lying southeast of the boundary of the Great Basin (which is drawn on the boundary between the Lahontan Hydrologic Basin and the Colorado Desert Hydrologic Basin as defined by the California Department of Water Resources) are conjectural, as are the three lakes near the northern end of the state. The three largest of the southern lakes—Bristol, Cadiz, and Danby—have extensive subsurface deposits of salts, which must have required large volumes of water, but only Danby Lake shows any reported geomorphic or stratigraphic evidence of the existence of an enlarged lake; a small terrace composed of lacustrine sediments was tentatively reported by Thompson (1929: 708), but it has not been subsequently verified. These basins may eventually be
Table 10-3.
Pluvial Lakes in the Western United States with Radiocarbon-Dated Chronologies in the Range of 25,000 to 10,000 yr B.P.

<table>
<thead>
<tr>
<th>State/ Lake Number</th>
<th>Pluvial Lake (and Name of Modern Lake, Playa, or Valley, if Different)</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Finite ¹⁴C Dates (yr B.P.)</th>
<th>Maximum Area (km²)</th>
<th>References* (Most Useful References in Italics)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arizona</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Cochise (Wilcox)</td>
<td>32°08’</td>
<td>109°51’</td>
<td>33</td>
<td>27,600 ± 900 ( &gt; 30,000)</td>
<td>Schreiber et al., 1972; Long, 1966</td>
</tr>
<tr>
<td>2</td>
<td>Adobe (Black Lake)</td>
<td>37°55’</td>
<td>118°36’</td>
<td>7</td>
<td>11,350 ± 350</td>
<td>Bauvelt, 1970, personal communication, 1978; Snyder et al., 1964; Hubbs and Miller, 1948</td>
</tr>
<tr>
<td>3</td>
<td>Deep Spring</td>
<td>37°17’</td>
<td>118°02’</td>
<td>47b</td>
<td>10,000 ± 1000</td>
<td>ca. 44 unknown*</td>
</tr>
<tr>
<td>4</td>
<td>Le Conte (Salton Sea)</td>
<td>33°20’</td>
<td>115°00’</td>
<td>48</td>
<td>&gt; 50,000</td>
<td>ca. 4600 &gt; 85d Hubbs and Miller, 1948; Hubbs et al., 1960 (R); Hubbs et al., 1963 (R); Hubbs and Bien, 1967 (R); Crane and Griffin, 1958 (R); Ferguson and Libby, 1962, 1963 (R); Bien and Pandolfi, 1972 (R); van de Kamp, 1973; Spiker et al., 1977 (R); Stanley, 1966; Wilson and Wood, 1980</td>
</tr>
<tr>
<td>5</td>
<td>Manix (Troy Lake, Coyote Lake)</td>
<td>35°03’</td>
<td>116°42’</td>
<td>5</td>
<td>30,050 ± 100</td>
<td>407 116 Hubbs and Miller, 1948; Snyder et al., 1964; Hubbs et al., 1962 (R); Ferguson and Libby, 1962 (R); Hubbs et al., 1965 (R); Berger and Libby, 1967 (R); Blackwelder, 1933</td>
</tr>
<tr>
<td>6</td>
<td>Fly Ranch (Dent Valley Salt Pan, Badwater)</td>
<td>36°00’</td>
<td>116°48’</td>
<td>4</td>
<td>21,500 ± 700</td>
<td>1600 183 Hooke, 1972; Snyder et al., 1964</td>
</tr>
<tr>
<td>7</td>
<td>Mojave (Soda Lake, Silver Lake)</td>
<td>35°22’</td>
<td>115°08’</td>
<td>24</td>
<td>15,350 ± 240</td>
<td>ca. 200 &gt; 12 Ore and Warren, 1971; Snyder et al., 1964; Stuiver, 1969 (R)</td>
</tr>
<tr>
<td>8</td>
<td>Panamint</td>
<td>36°18’</td>
<td>117°18’</td>
<td>10</td>
<td>32,900 ± 1700</td>
<td>722 283 Snyder et al., 1964; Hubbs et al., 1963 (R); Berger and Libby, 1966 (R); Smith, 1978; Petersen, 1980</td>
</tr>
<tr>
<td>9</td>
<td>Russell (Mono Lake)</td>
<td>38°03’</td>
<td>118°46’</td>
<td>19f</td>
<td>36,280 ± 600*</td>
<td>692 238 Lajoie 1968, personal communication, 1981; Hubbs et al., 1965 (R); Ferguson and Libby, 1962 (R); Snyder et al., 1964</td>
</tr>
<tr>
<td>10</td>
<td>Searles (China Lake, Searles Lake)</td>
<td>35°36’</td>
<td>117°42’</td>
<td>110</td>
<td>46,350 ± 1500*</td>
<td>994 196 Smith, 1968; Stuiver, 1964; Flint and Gate, 1958; Rubin and Berthold, 1961 (R); Ives et al., 1964 (R); Levin et al., 1965 (R); Ives et al., 1967 (R); Robinson, 1977 (R); Marsters et al., 1969 (R); Feng et al., 1978; Damon et al., 1964 (R)</td>
</tr>
<tr>
<td>11</td>
<td>Tulare (Kern, Buena Vista, and Tulare Lakes)</td>
<td>36°00’</td>
<td>119°40’</td>
<td>8</td>
<td>26,780 ± 600*</td>
<td>ca. 4100 j Dalrymple et al., 1969; Hubbs and Bien, 1967 (R); Buckley et al., 1968 (R)</td>
</tr>
<tr>
<td>Nevada</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Dixie (Humboldt Salt Marsh)</td>
<td>39°55’</td>
<td>117°60’</td>
<td>2</td>
<td>11,700 ± 180</td>
<td>1088 72 Buckley and Willits, 1970 (R) Snyder et al., 1964; Hubbs and Miller, 1948</td>
</tr>
<tr>
<td>13</td>
<td>Lahontan (Pyramid, Walker and Honey Lakes; Carson Sink;</td>
<td>40°00’</td>
<td>119°30’</td>
<td>169</td>
<td>40,000k</td>
<td>22,440 163 Snyder et al., 1964; Hubbs and Miller, 1948; Benson, 1978; 1981; Muffin and Wheat, 1979; Broecker and Orr, 1958; Olson and Broecker, 1961 (R); Broecker</td>
</tr>
<tr>
<td>State</td>
<td>Pluvial Lake (and Name of Modern Lake, Playa or Valley, if Different)</td>
<td>Latitude (°N)</td>
<td>Longitude (°W)</td>
<td>Finite 14C dates (yr B.P.)</td>
<td>Maximum Increase in Depth Relative to Present (m²)</td>
<td>References* <em>(Most Useful Reference in Italics)</em></td>
</tr>
<tr>
<td>---------</td>
<td>---------------------------------------------------------------------</td>
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<td>-----------------------------------------------</td>
</tr>
<tr>
<td>Nevada</td>
<td>Winnemucca, Smoke Creek, Black Rock, Desert Valley, and Buena Vista Basins)</td>
<td>36°19'</td>
<td>115°11'</td>
<td>3 31,300 ± 2500 unknown</td>
<td>unknown</td>
<td>Haynes, 1967; Mifflin and Wheat, 1979</td>
</tr>
<tr>
<td>Nevada</td>
<td>Las Vegas</td>
<td>38°12'</td>
<td>118°21'</td>
<td>1 10,760 ± 400 unknown</td>
<td>unknown</td>
<td>Hay, 1966; Crane and Griffith, 1965 (R)</td>
</tr>
<tr>
<td>Nevada</td>
<td>Teel (Teels Marsh)</td>
<td>34°04'</td>
<td>103°07'</td>
<td>5 22,300 ± 700 ca. 130</td>
<td>&gt;12</td>
<td>Reeves, 1966b; Glass et al., 1973; Leonard and Frye, 1975; Hester, 1975; Porsild and Broecker, 1961 (R)</td>
</tr>
<tr>
<td>Nevada</td>
<td>Arch (Big Salt Lake, Laguna Salada)</td>
<td>34°15'</td>
<td>103°20'</td>
<td>17 15,770 ± 440 unknown</td>
<td>unknown</td>
<td>Haynes, 1975</td>
</tr>
<tr>
<td>Nevada</td>
<td>Blackwater Draw</td>
<td>34°45'</td>
<td>106°00'</td>
<td>4 &gt; 33,000 2860³</td>
<td>90</td>
<td>Bachhuber, 1971; Bachhuber and McEwen, 1977</td>
</tr>
<tr>
<td>Nevada</td>
<td>Estancia (Laguna del Perro, Salina Lake, etc.)</td>
<td>35°27'</td>
<td>103°09'</td>
<td>2 16,010 ± 180 uncertain</td>
<td>uncertain</td>
<td>Leonard and Frye, 1975; Coleman, 1974 (R); Glass et al., 1973</td>
</tr>
<tr>
<td>Nevada</td>
<td>Lea County</td>
<td>34°26'</td>
<td>103°49'</td>
<td>1 15,280 ± 210 unknown</td>
<td>unknown</td>
<td>Coleman, 1974 (R); Glass et al., 1973; Leonard and Frye, 1975</td>
</tr>
<tr>
<td>Nevada</td>
<td>Portales Valley</td>
<td>33°50'</td>
<td>108°10'</td>
<td>16 27,000 ± 5000 660 3200</td>
<td>50³</td>
<td>Powers, 1939; Stearns, 1962; Stuiver and Deevey, 1962; Cisney and Searl, 1956; Damon et al., 1964; Long and Mielke, 1966; Schultz and Smith, 1965; Foreman et al., 1959</td>
</tr>
<tr>
<td>Nevada</td>
<td>San Augustin (San Augustin Playa)</td>
<td>34°27'</td>
<td>108°46'</td>
<td>1 23,000 ± 1500 5</td>
<td>15</td>
<td>Schultz and Smith, 1965; Haynes et al., 1967 (R); Cummings, 1968</td>
</tr>
<tr>
<td>Nevada</td>
<td>Zuni</td>
<td>42°40'</td>
<td>120°30'</td>
<td>6 30,700 ± 2500 1240 1900</td>
<td>115</td>
<td>Allison, 1966; Buckley et al., 1968 (R); Allison, 1954; Phillips and Van Denburgh, 1971; Van Denburgh, 1973; Levin et al., 1965 (R); Ives et al., 1967 (R); Sullivan et al., 1970 (R)</td>
</tr>
<tr>
<td>Nevada</td>
<td>Chewaucan (Abert Lake, Summer Lake)</td>
<td>43°10'</td>
<td>120°45'</td>
<td>4 29,000 ± 2000 3885 1600</td>
<td>49</td>
<td>Bedwell, 1973; Allison, 1966, 1979</td>
</tr>
<tr>
<td>Texas</td>
<td>Guthrie</td>
<td>33°06'</td>
<td>101°48'</td>
<td>4 34,400 ± 3450 unknown</td>
<td>unknown</td>
<td>Reeves and Parry, 1965; Reeves, 1966b</td>
</tr>
<tr>
<td>Texas</td>
<td>Lubbock</td>
<td>33°38'</td>
<td>101°54'</td>
<td>3 12,650 ± 250 unknown</td>
<td>unknown</td>
<td>Green, 1961; Wendorf, 1970; Black, 1974; Hester, 1975; Broecker and Kulp, 1957 (S)</td>
</tr>
<tr>
<td>Texas</td>
<td>Monahans Dunes</td>
<td>31°36'</td>
<td>102°53'</td>
<td>2 19,200 ± 500 unknown</td>
<td>unknown</td>
<td>Haynes, 1975, citing Green, 1961; Broecker and Kulp, 1957 (S); Olson and Broecker, 1961 (R)</td>
</tr>
<tr>
<td>Texas</td>
<td>Mound</td>
<td>33°05'</td>
<td>102°05'</td>
<td>8 &gt;37,000 &lt;20 &lt;15</td>
<td>Reeves and Parry, 1965; Reeves, 1966b; Bates et al., 1970; Hester, 1975; Harbour, 1975</td>
<td></td>
</tr>
<tr>
<td>Texas</td>
<td>Rich</td>
<td>33°17'</td>
<td>102°12'</td>
<td>4 32,525 ± 2400 unknown</td>
<td>unknown</td>
<td>Reeves and Parry, 1965; Reeves, 1966b; Hester, 1975; Haynes, 1975; Olson and Broecker, 1961 (R)</td>
</tr>
<tr>
<td>Texas</td>
<td>White</td>
<td>ca. 33°58'</td>
<td>102°44'</td>
<td>1 19,275 ± 560 unknown</td>
<td>unknown</td>
<td>Reeves and Parry, 1965; Hester, 1975; Harbour, 1975</td>
</tr>
</tbody>
</table>
Table 10-3. Pluvial Lakes in the Western United States with Radiocarbon-Dated Chronologies in the Range of 25,000 to 10,000 yr B.P.

<table>
<thead>
<tr>
<th>State/Lake Number</th>
<th>Pluvial Lake (and Name of Modern Lake, Playa, or Valley, if Different)</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Finite ¹⁴C Dates (yr B.P.)</th>
<th>Maximum Area (km²)</th>
<th>Maximum Depth Relative to Present (m)</th>
<th>References (Most Useful References in Italic)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Utah</td>
<td>Bonneville (Great Salt, Utah, and Sevier Lakes; Great Salt Lake and Escalante Deserts; Cache, Sevier, White, and Rush Valleys)</td>
<td>40°30'</td>
<td>113°00'</td>
<td>114</td>
<td>&gt;37,000⁹</td>
<td>51,640</td>
<td>ca. 335</td>
</tr>
</tbody>
</table>

⁹References to radiocarbon date lists published in Radiocarbon are indicated by (R), those in lists in Science are indicated by (S), and unless cited in text they are not included in the references at the end of this chapter.

¹With one exception, these ages were measured on diagenetic dolomite in lacustrine muds and, therefore, should not be regarded as accurate measures of the time of sedimentation. Peterson et al., (1963) suggests that the dolomite began to nucleate after the close of the last pluvial period, about 10,000 yr B.P.

²The former lake depth imposible to reconstruct because of poor lacustrine record.

³The lake basin has been strongly affected by faulting and tiling, so the late-Pleistocene configuration is difficult to reconstruct accurately.

⁴Highest shoreline remnants dated suggest a maximum depth of 208 m.

⁵Dates are unpublished.

⁶Lacustrine deposits recovered by a core middle of basin indicate that it has been continuously occupied by a lake since a time greater than 730,000 yr B.P., the age of the interbedded Bishop Tuff (Lajoe, 1968).

⁷Major fluctuations in lake level have occurred throughout the last 3.2 million years (Lidicocat et al., 1980).

⁸Lacustrine clay units within the Quaternary sediments of the San Joaquin Valley indicate at least nine major lake expansions beginning well before 600,000 yr B.P.

⁹Depths of large late-Pleistocene and early-Holocene lakes are uncertain because of structural downwarping and changes in the height of the alluvial-fan barriers separating the present lake basins, but they probably were shallow.

¹⁰Oldest ³⁹⁹Th age is 250,000 yr B.P.

¹¹Measured from natural (preirrigation) level of Pyramid Lake (1180 m).

¹²Three dates from lacustrine beds, numerous others from nonlacustrine beds.

¹³Maximum area includes satellite Pinus Wells and Encino Basins.

¹⁴Highest shoreline remnants dated suggest a maximum depth of 69 m.

¹⁵Oldest ³⁹⁹Th age is 105,000 yr B.P. Major fluctuations of the lakes in the Bonneville Basin began before deposition of Bishop Ash, potassium-argon dated at 730,000 yr B.P. (Eardley et al., 1973).
in evaporation (from 112 to 79 cm). These conditions could have resulted from a 550-m depression of the present elevation-precipitation relation and a decrease in summer temperatures that lowered snow lines by 1200 m.

The lacustrine sediments in Las Vegas Valley, inferred by Haynes (1967: 78) to be remnants of pluvial Lake Las Vegas, are considered by Mifflin and Wheat (1979: 27) to be paludal or playa deposits. The lack of other large pluvial lakes in this area, the absence of a geologic mechanism for restricting a smaller water body to the northern part of the valley, the relatively young geologic ages of the fine-grained deposits, and the almost complete absence of other exposures of lake sediments in the entire valley combine to support the more recent view. However, the thickness of mudstones and carbonates (4.5 m) and the details of their lithologies support the earlier conclusion. If the “lake deposits” in Las Vegas Valley are eventually shown to be paludal deposits, the correspondence between their age and the ages of pluvial-lake deposits elsewhere could simply reflect a rise in water tables and an increase in spring activity that accompanied those climatic shifts.

**New Mexico**

Pluvial Lake Estancia in the central part of the state consisted of two freshwater lakes as much as 90 m deep during the interval that began about 18,000 and ended 10,500 yr B.P. The lakes were separated by one low stand near the middle of this period (Buchhuber and McClellan, 1977: 254, Figure 2). Another smaller lake, Lake Willard, formed during the period between about 8500 and 6000 yr B.P. Saline stages at the beginning of the first freshwater stage and at the beginning of Lake Willard supported marine-type foraminiferal assemblages. Their faunal composition suggests that summer temperatures during the initial stages of lake development were nearly 10°C lower than present-day summer temperatures. Leopold (1951), using meteorologic data and methods, suggests a 9°C reduction in summer temperature for this area and estimates a decrease in mean annual temperature during pluvial periods of about 8°C, a decrease in evaporation to about 70% of the present rate, and an increase in precipitation of about 50%.

The San Augustine Plains, a tectonic basin in west-central New Mexico, also contained a pluvial lake, which covered 660 km² to a maximum depth of 50 m (Foreman et al., 1959: 117; Stearns, 1962: 29). Shorelines, beaches, and bars record its high levels; two cores that sample the upper 610 m of valley fill confirm lacustrine deposition for most of the last half of the time represented. The pollen in the upper 197 m of the core suggests that subalpine climates characterized the area only during the latter part of the interval sampled and that warmer and drier climates characterized the earlier times (Clishy and Sears, 1956). A series of radiocarbon dates on carbonate and organic carbon in a nearby 6-m core shows that climatic conditions during almost all of the period from 25,000 to 10,000 yr B.P. produced lakes in this depression (Stuiver and Deevey, 1962). Dates on tufa from the high terraces, though less reliable, suggest high stands in the valley during the last 5000 years of this period (Long and Mielke, 1966).

Other pluvial lakes in New Mexico also reached high levels during this interval. Very small tributary areas characterize some basins, and they, like the lakes in southern California, may reflect regional increases in the pluvial period’s water-table levels rather than increased surface flow. The pluvial-lake water surface in Salt Lake, which lies in a subsidence depression in the east-central High Plains area, stood at a level about 12 m above the present lake’s surface during a period that ended about 14,000 yr B.P. (Glass et al., 1973: 9). Pluvial Lake Zuni, which lies in the center of a small, steep-sided volcanic crater, left marl deposits, radiocarbon dated at about 22,000 yr B.P., as much as 15 m above the present salt-lake floor (Damon et al., 1964).

**Oregon**

Pluvial lakes occupied all of the closed basins in the Oregon segment of the Great Basin, and many of these contain lakes or marshy areas today. The largest, Lake Modoc, lay in the Basin and Range province, west of the northern Great Basin and astride the California-Oregon border. It covered 2800 km² and had a maximum depth of 64 m (Dicken, 1980: 183). Age control is unavailable, but the hydrology of the basin would make translation of a lake history into climatic terms difficult because of the high porosity of the enclosing volcanic rock and because the present outlet, the Klamath River, must have been blocked by ice, landslides, or some other mechanism.

Pluvial Lake Chewaucan, the lake that occupied the basins now containing Abert Lake and Summer Lake in south-central Oregon, covered an area of about 1240 km² to a maximum depth of about 115 m (Allison, 1954; Phillips and Van Denburgh, 1971: B12). As noted by Van Denburgh (1975: C29), the elevations of the shorelines around pluvial Fort Rock Lake to the north, which covered the present Fort Rock Lake and Christmas Lake valleys, are nearly the same; this fact suggests a subsurface hydrologic connection, and any climatic reconstruction should probably combine data from both basins. The history of Fort Rock Lake is partially known on the basis of discontinuous outcroppings of lacustrine sediments that indicate two lake stands separated by a low stand that produced an unconformity (Bedwell, 1973; Allison, 1979: 44). The lower section, 60 to 200 m thick, is overlain unconformably by a few meters of lacustrine deposits that have radiocarbon dates on two samples near their base and about a meter apart averaging about 30,000 yr B.P. The upper unit locally has conformities within it, suggesting that minor lake fluctuations occurred throughout the period. All lake deposits are overlain by the Mazama Ash (6600 yr B.P.).

The combined area of Lake Chewaucan and Fort Rock Lake is 5125 km² (Table 10-3); the present area of the only perennial lakes in the two basins, Abert and Summer, which probably receive substantial amounts of water from the Fort Rock Basin, is 350 km² (Phillips and Van Denburgh, 1971: B5, B24). The pluvial-lake area, therefore, is about 15 times that of the present area, and calculations like those applied to Lake Bonneville. Lake Lahontan, and the Owens River system suggest pluvial inflow volumes at least 5 times those of the present.

A pluvial lake in the Warner Valley covered 1310 km² and had a maximum depth of 98 m. Weide (1976) has applied the climatic reconstruction methods used by Leopold (1951) and Snyder and Langbein (1962) to this area, although it is climatically and geographically quite different, and has concluded that the Snyder and Langbein model most satisfactorily accounts for both the present hydrologic regime (which supports several small lakes) and the pluvial lake’s regime. The present lakes have a 22nd the area of the pluvial lake in that valley, and Weide’s calculated pluvial runoff ranges from 2.4 to 3.0 times the calculated present runoff. Pluvial runoff was the result of an estimated 17% to 37% increase in precipitation and a
16% to 23% decrease in lake evaporation caused by a 4° to 6°C decrease in mean annual temperature (Weide, 1976: Table 1).

Texas

Many small pluvial lakes existed in the southern High Plains of northwestern Texas. Most left geomorphic, stratigraphic, and paleontologic records of their existence, and there have been numerous studies of them, especially during the last two decades. The smallest lakes formed mostly in basins created at least in part by deflation; larger lakes occurred in basins attributed to the blocking of older drainage channels followed by erosional enlargement (Reeves, 1966a: 271-81). Since the last pluvial period, many of the lacustrine deposits have themselves been dissected by wind and water. The late Wisconsin lacustrine deposits are composed of gravel, sand, black to blue-gray clay (mostly silt), carbonates (mostly dolomite), and gypsum (Reeves, 1976: 220-23).

Most of the lake deposits are assigned by Reeves (1968, 1972, 1976: 220) to the Tahokan pluvial, inferred to have lasted from about 22,000 to 12,500 yr B.P. Radiocarbon dates of these sediments are not numerous, and many are on carbonates and, therefore, are questionable. Carbonates near the top and base of the lower of the two units of lacustrine deposits in Rich Lake have radiocarbon dates of about 26,500 to 17,400 yr B.P., and the upper unit is estimated to represent the period between 16,000 and 12,000 yr B.P. Shells in correlative lacustrine deposits in the Monahans dune area have dates of about 19,200 and 13,400 yr B.P. (Haynes, 1975: 80-82). Between 13,000 and 11,000 yr B.P., however, desiccation, lower water tables, and deflation characterized the region (Haynes, 1975: 83).

Pluvial lakes in northwestern Texas are considered by Reeves (1976: 214) to have been the product of a 100% increase in precipitation. Some of the basins, however, contained lakes that occupied as much as 85% of their drainage basins (Reeves, 1966a: 643), and these lake-to-drainage-area ratios are much greater than those found in other pluvial-lake basins. This ratio, designated by Snyder and Langebein (1962: 2389, Table 3) as Z, was calculated by them for eight typical pluvial lakes, and they found the values of Z to lie in the range of 0.22 to 1.12, with the highest values characteristic of lakes adjacent to high mountain ranges, as theory would suggest. The Z value for some of the Texas pluvial lakes would exceed 4.0, and the adjacent regions are definitely not characterized by high mountain ranges. It seems likely, therefore, that ground water contributions were required. Because of the impervious nature of the regional soils, Reeves (1973: 701-2) considers subsurface flow to the pluvial lakes in the southern High Plains to be a minor part of their hydrologic budget. However, we feel that the anomalous value of Z and the disproportionate size of even a few lakes in this area relative to their basins argue more strongly for their nourishment by an elevated pluvial-period groundwater table and for a major contribution to all the lakes from groundwater. Many or all of these lakes, therefore, in combination with some of the pluvial lakes in southern California and New Mexico, may document increases in the level of the pluvial period's groundwater table over a large segment of the southwestern United States south of about latitude 35° and from Texas westward. This regional transition in hydrologic regimes may be a result of the southward increase in summer precipitation in this area along a steep gradient at this latitude, as is indicated by paleobotanical evidence (Wells, 1979: 322-24). Lake Cochise in Arizona may also have had this origin, but the evidence is inconclusive.

Areal and Temporal Variations in Pluvial-Lake Levels

Evidence that can be used to reconstruct partial histories of former lake levels is available in 31 closed basins of the West. In Figures 10-2, 10-3, and 10-5, curves representing the histories of four basins are plotted: Lake Bonneville, Lake Lahontan, and two lakes in the Owens River system. Although disagreements remain among interpretations of the details of these histories, there appear to be broad areas of agreement. There is agreement, for example, on the fact that rapid and large-amplitude fluctuations occurred in these late-Quaternary pluvial lakes. It is also generally accepted that there were one or more high lake stands during the first two-thirds of the period between 25,000 and 10,000 yr B.P. and that there were one or more very brief periods of expansion about 12,000 yr B.P. In addition, most of the evidence shows that there were significant fluctuations in the sizes of these lakes during the last 10,000 years.

Because it is possible to make reliable reconstructions of past water levels during at least part of the history of many other lakes, it is feasible to extend the spatial coverage of the dated pluvial-lake histories beyond the areas represented in Figures 10-2, 10-3, and 10-5 by pooling all the available dated lake-level evidence, assessing its validity, and assuming that regional climates affected large areas in a similar manner. To do this, we made a careful search of the literature relevant to the pluvial lakes of the western United States, following the procedures established by Street and Grove (1976, 1979) and concentrating on closed basins that have yielded radiocarbon dates. The present survey revealed 31 dated basins in the United States (including the Salton Sea), and the entire compilation has been checked and revised. The results are summarized in Figure 10-6, and the basins for which we found data are listed in Table 10-3.

A standard procedure is used for expressing the evidence in a semiquantitative form (Street and Grove, 1979: 84-87). Each basin is considered in relation to its own internal range of variation. This procedure allows for minor downcutting of outlets or uncertainties in the elevations of the shorelines or lacustrine sediments. Each radiocarbon date and the relative elevation of the indicated lake level is assigned to 1000-year time intervals covering the period between 30,000 years ago and the present. Where dates within a 1000-year interval indicate different levels, the level chosen for mapping is defined as follows: High: lake level was high (70% to 100% of the maximum lake level) during all or part of this time period; intermediate: lake level was intermediate (15% to 70%) or low, but not high, during all or part of this time period; and low: lake level was low (0% to 15%) during all of this time period. This method underemphasizes arid periods shorter than 1000 years, and it may overemphasize single radiocarbon dates from a given basin, which mostly came from lacustrine rather than nonlacustrine deposits.

Where radiocarbon dates conflict, the interpretation used is based on the following inferred order of the dates' decreasing reliability: (1) charcoal and wood in basin sediments or cave sequences; (2) peat or organic carbon in lake sediments; (3) disseminated inorganic carbonate, calcareous algae, ostracods, and mollusks in lake sediments; (4) tufa; and (5) carbonate sediments. Dates on bone and on soil humus or carbonate have been treated with caution. No adjustment has been made for the initial 14C/12C ratio of the lake waters.

The information of lake levels is summarized in two ways. In Figure 10-6, the history of lake levels over the entire West is shown in the form of a histogram of lake levels against time. Figure 10-7
Figure 10-6. Relative percentages of basins containing dated materials at elevations indicative of low, intermediate, or high stages versus their radiocarbon ages.

shows the same information on six maps for 1000-year time periods starting at 24,000, 18,000, 16,000, 13,000, 12,000, and 10,000 yr B.P. The maps indicate the areal distribution of each lake level during that block of time and presumably reveal regional variations in the intensity of pluvial conditions. It should be noted that the symbols representing Lake Bonneville and Lake Lahontan indicate conditions over areas that cover a quarter to a third of their respective states. All of the 1000-year intervals were plotted; the time periods illustrated represent the clearest or most significant patterns.

During the period 24,000 to 14,000 yr B.P. (Figure 10-6), all of the lakes for which information is available were either high or intermediate. High levels were most widespread from 24,000 to 21,000 yr B.P., when only Lake Bonneville appears to have remained intermediate. Between 21,000 and 14,000 yr B.P., only about 70% of the data points record high levels. The intermediate-depth lakes of these ages clustered in the northwestern part of the pluvial-lake region (Figures 10-7B and 10-7C), particularly in areas adjacent to the Sierra Nevada, although there was considerable fluctuation. No information is available from Oregon to determine whether the region of intermediate levels extended that far northwest. A brief episode of partial desiccation in the southern High Plains appears to have been recorded by the widespread but thin Vigo Park dolomite identified by Reeves and Parry (1965; Reeves, 1976), although the radiocarbon dates from this lithologic unit are scattered widely and are probably not very reliable.

The period between 14,000 and 10,000 yr B.P. was characterized by rapid, large-amplitude fluctuations that may or may not have been synchronous across the region. The apparent lack of synchrony in the raw data could be the result of errors inherent in the radiocarbon dating method when applied to the record of events that were apparently short. The data may, however, reflect genuine spatial variations in lake behavior. Many lakes experienced a drop in levels centered on the period 14,000 to 13,000 yr B.P. In Arizona, New Mexico, and Texas, 13,000 yr B.P. has been considered the end of the main lacustrine phase (Haynes, 1967), although apparent high stands of a few lake levels in these areas (Figures 10-7D and 10-7E) partially conflict with this view. The pattern of fluctuations during the period 12,000 to 10,000 yr B.P. is particularly complex. A major expansion of Lakes Mojave, Searsar, Mono, and Tulare, located in southern California near the Sierra Nevada, culminated between about 13,500 and 11,000 yr B.P. Broecker and Orr (1958: Figure 9) and Broecker and Kaufman (1965: Figure 3) also infer that the first of two major expansions of Lake Bonneville and Lake Lahontan occurred within this period, but Bright's (1966) radiocarbon data from peaty lake sediments appear to preclude any overflow of Lake Bonneville after about 13,000 yr B.P. Morrison and Frye (1965: Figure 2), however, indicate several expansions of Lake Bonneville and Lake Lahontan starting about 10,000 yr B.P. on the basis of stratigraphic evidence, and we suspect that one or more modest reexpansions did occur. Distinct expansions occurred in Searsar Lake and Lake Mojave between about 11,000 and 13,000 yr B.P., and a brief lacustrine recovery occurred in the pluvial lakes of Arizona, New Mexico, and Texas during this time (Wendorf, 1961).

Between 10,000 and 5000 yr B.P. (Figure 10-6), many areas were characterized by drought, which culminated between 6000 and 5000 yr B.P., when not a single lake is known to have been high. In some basins, the fall in water levels during the first part of this period was apparently rapid and unreversed, but in others there seems to have been some lake level fluctuation.

Between 5000 yr B.P. and the present, there have been significant reexpansions of some lakes, notably those on the western margins of the Great Basin and in California. The vertical amplitude of these fluctuations is surprisingly large: up to 90 m in the Walker Lake area, 85 m around the Salton Sea, and 65 m in Searsar Lake. During historic time, lake levels were higher before the start of irrigation agriculture in the late 19th century, and present lake levels in many areas are unrepresentative of the average for the last few centuries or millennia (Harding, 1965).
Paleoecological Estimates Derived from Lake-Level Data

Lake-level fluctuations are potentially the best source of quantitative estimates of paleoprecipitation and one of the best sources for generalized paleoclimatic data in desert areas where pollen data or studies of pack-rat middens are sparse. However, attempts to derive paleoprecipitation estimates from lake-level information have limitations, largely because of the difficulties involved in estimating past rates of evaporation and percentages of precipitation that became runoff. The problems encountered, described in numerous papers, have been recently summarized by Brakenridge (1978), Mifflin and Wheat (1979), and Benson (1981).

The surface area of a topographically closed lake fluctuates in order to balance the lake’s water budget, as expressed in the following equation:

\[ \text{Lake Level} = \text{Evaporation} - \text{Precipitation} - \text{Runoff} \]

Figure 10.7. Locations of lakes having low, intermediate, or high stands during 1000-year periods that began (A) 24,000, (B) 18,000, (C) 16,000, (D) 13,000, (E) 12,000, and (F) 10,000 yr B.P.
\[ A_1 P_L + A_2 P_T k + G_t = A_1 E + G_O, \]

where:

- \( A_1 \) = lake area,
- \( P_L \) = precipitation falling on the lake,
- \( A_2 \) = tributary catchment area,
- \( P_T \) = mean annual precipitation falling in the tributary area,
- \( k \) = runoff coefficient (the proportion of the rain or snow that eventually becomes runoff and reaches the lake),
- \( G_t \) = ground water inflow to the basin,
- \( E \) = mean annual evaporation from the lake, and
- \( G_O \) = groundwater outflow from the basin.

Theoretically, this equation can be used to calculate past values of \( P_L \) and \( P_T \) provided that: (1) the basin has been topographically and hydrographically stable, (2) \( A_1 \) and \( A_2 \) can be determined from geologic information, (3) \( G_t \) and \( G_O \) are negligible or can be determined from hydrologic information, and (4) \( k \) and \( E \) can be reliably estimated.

In most late-Pleistocene pluvial basins where a good record exists, pluvial values for \( A_1 \) and \( A_2 \) can be measured quite accurately, but the remaining values are less readily determined. Present values for \( P_L \) can be measured, but, as emphasized by Weide (1976), even present values for \( P_T \) are not easily determined in areas where topographic relief is great. Present values of \( k \) are imperfectly known, and estimates of pluvial values of \( P_L \) and \( k \) are, therefore, even less reliable. Similarly, present values (or even the sign of) \( G_t \) and \( G_O \) are commonly not well known, and their values and signs may not have been the same during pluvial periods in the past. The fact that we know far less than we could of the present relationships among these climatic and hydrologic factors in desert areas suggests this as a topic needing more study. Especially needed is a better basis for developing estimates of pluvial runoff from areas now contributing no runoff, which constitute the largest areal percentage of many basins.

The value for \( E \) depends only in part on the temperature of the lake water. However, because the other relevant factors are difficult to reconstruct from geologic data, most studies—including the present one—assume that pluvial values for \( E \) can be satisfactorily estimated from empirical relationships between modern data on evaporation and air temperature. Significant differences of opinion remain, however, about the change in air-temperature values during Pleistocene pluvial periods (Table 10-2). Temperature estimates used in reconstructing lake histories are derived from a host of geologic and hydrologic criteria, all of which are subject to uncertainties, and some are derived from evidence in areas far removed from the site of the paleolake. Possible errors in paleotemperatures are then compounded by becoming the basis of only partially constrained assumptions concerning past atmospheric lapse rates and relative humidity, which are combined with nearly unconstrained estimates of percentages of cloud cover, velocities of winds, and the seasonal distributions of all these variables. The conclusions reached by various authors concerning pluvial values for \( E \), \( P_L \), and \( P_T \) have, therefore, differed greatly, both because of difference in the paleotemperature values adopted and because of their assumptions concerning the other variables.

The problems stemming from uncertainties about pluvial paleotemperatures are not avoided by the method used in this chapter, which compares present and pluvial lake surface areas and assumes that pluvial-period air temperatures bore about the same (although unspecified) relation to pluvial-lake surface-water temperatures as they do today (obviously, this method can only be used in basins that still contain lakes). It also assumes that other factors tended to cancel each other. During pluvial periods, increased evaporation was probably promoted by higher wind velocities (indicated by the clast sizes found in pluvial-lake gravel bars constructed by wind-driven waves and currents) and by decreased salinities. Decreased evaporation was promoted by higher relative humidities and lower solar radiation values (which are suggested by several meteorologic considerations). It also assumes that the decreased net evaporation caused by greater amounts of precipitation on the lake surface can be arbitrarily placed at 35%. Many important factors are not reflected in the results of this simplified approach, but for some purposes estimates of the volume of pluvial runoff—the values derived here—are of more interest than estimates of the precipitation and/or temperature that caused them, for they lead more directly to estimates of pluvial-period stream dimensions, flood magnitudes, and sedimentation and erosion rates.

**Conclusions**

During pluvial maxima, some (probably most) pluvial-period ground-water tables in the American Southwest rose more than 10 m. Most streams in the remaining parts of the western United States carried substantially more water during pluvial periods than they do now, and precipitation amounts probably also increased markedly. The volume of stream flow during "maximum pluvial" periods in this region possibly approached an order of magnitude greater than today's.

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