# Chapter 25

# Evidence for Climate Change From Desert Basin Palaeolakes

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## Introduction

Lakes have long been recognized as being rich storehouses of environmental information. A lake basin collects water, but also sediment, much of which has been weathered and transported via fluvial processes from the near and far reaches of its drainage basin. The amount of water held in a lake is recorded on the landscape in coastal erosional and depositional landforms created at the water's edge. The sediments deposited on the bottom of the lake can be clastic, geochemical, or biogenic, and include materials derived within the standing water body itself, such as through coastal erosion, chemical precipitation, or biogenic concentration, as well as those delivered to the lake from the surrounding drainage basin. In most cases only a small percentage of a lake's sediment load is delivered from outside of the drainage basin as aeolian fallout. Because, under natural conditions, climate is the main determinant of the amount of water in a lake and because it influences some important characteristics of the lacustrine sediments and biota, changing climatic conditions are represented in the suites of abandoned shorelines and accumulations of sediments left by the lake over time (Fig. 25.1). This archival property makes the geomorphic and sedimentologic evidence of present and past lakes valuable as environmental and palaeoenvironmental indicators. Such evidence from late Quaternary palaeolakes, in fact, ranks among of the most complete and accessible sources of palaeoclimatic proxy

**Fig. 25.1** An abandoned gravel shoreline in the Great Basin, USA, partially overlain by pelagic lacustrine deposits of calcium carbonate (marl)

data currently available for the late Pleistocene and Holocene.

Earth scientists have conducted comprehensive studies of the geomorphic and sediment evidence of late Pleistocene and Holocene palaeolakes, and have made palaeoclimatic interpretations from them, since the late 19th century (Russell 1885, Gilbert 1890). Limited by poor age control, and to some extent by interest in other topics when the Davisian cycle of erosion paradigm was popular (Davis 1899), the number of palaeolake studies waned during the first half of the 20th century. About mid-century, palaeolake research began a slow but steady growth under the process geomorphology paradigm and as numerical dating techniques became established and increasingly refined. Eventually the growth in palaeolake research began to accelerate, along with interest in earth-system science, starting about 1980. Since approximately the mid-1990s, the number of palaeolake researchers and publications has grown dramatically reflecting

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increasing social and scientific concern with human impacts on the environment and global climate change.

Today, palaeolake investigations contribute to climate studies in many ways. Researchers work on accurately reconstructing details of the timing and extent of palaeolake-level fluctuations (e.g. Fornari et al. 2001, Godsey et al. 2005), estimating the local and regional values of climatic variables and circulation attributes that could have led to those fluctuations (e.g. Benson 1993, Bookhagen et al. 2001, Stone 2006, Dühnforth et al. 2006), searching for spatial and temporal similarities and differences in the behavior of multiple palaeolakes (e.g. Benson et al. 1995, Krider 1998, Mensing 2001, Zhang et al. 2004, Balch et al. 2005), and comparing the palaeoclimatic signal determined from palaeolakes with climate signals derived from other sources (e.g. Benson et al. 1998, Broecker et al. 1998, Lin et al. 1998, Stager et al. 2002, Balch et al. 2005). These efforts provide information on the amount and rate of natural climate variability experienced during the late Quaternary, and therefore on what might be possible in the future. They supply a record of past climatic conditions that emerging models of global climate change should be able to successfully hindcast. Furthermore, comparing fluctuations in various palaeolakes around the globe with oscillations present in such data sources as the marine oxygen-isotope record, the Greenland ice cores, and the earth's orbital parameters helps scientists understand the mechanisms, sensitivities, and teleconnections of the natural climate system. Clearly, reconstructing the timing and extent of palaeolake fluctuations is the scientific foundation that makes these applications possible.

## Desert Basin Palaeolakes

Lakes form wherever there is an adequate basin of containment and enough surplus water to accumulate in it. Topographic depressions that function as lake basins may be derived from a wide variety of sources. They originate through tectonic, volcanic, fluvial, aeolian, mass wasting, glacial, meteoritic, or other processes (Hutchinson 1957). Most lakes in humid climates receive so much inflow that the level of the standing water body permanently attains, and continually spills out over, the lowest point along the boundary of the containment basin. This low point is called the sill or threshold, and in humid regions the overflowing stream is typically part of an integrated, throughflowing, fluvial drainage system. Such open-basin, or externally drained, lakes have the elevation of their water level controlled by the elevation of the threshold. An increase of flow into an open-basin lake is handled by an increase in discharge out of the lake. Although the cross-sectional depth of the stream flowing out over the threshold will vary to some extent with discharge, much of the variation in volume of water is accounted for instead by the other two fluvial discharge variables, cross-sectional width and velocity of flow. As a result, the water level of open-basin lakes tends to be maintained very near the elevation of the threshold. Although this can lead to strongly developed coastal landforms within that narrow vertical zone, threshold control largely prohibits changes in the amount of water delivered to the lake from being sensitively recorded in distinct, multiple shorelines. A detailed record of changing conditions of effective moisture is thus lost. Alternatively, successively lower shorelines sometimes form in open lake basins as a result of fluvial erosion of the threshold and irrespective of vacillations in the regional effective moisture.

In many arid regions, topographic basins are commonly not connected with each other by throughflowing surface drainage, and this is primarily due to climatic factors (Langbein 1961). As in other climatic regions, topographic basins that may pond water can be formed by a variety of processes. In desert environments, however, once a large basin exists it is unlikely that sufficient surface water will be generated to completely fill the containment basin, spill over, and contribute to an integrated surface drainage system that reaches ultimate base level. Some desert basins contain perennial lakes while many others currently support only ephemeral or intermittent lakes (playas or playa lakes) (Mifflin and Wheat 1979, Smith and Street-Perrott 1983, Williams and Bedinger 1984). Perennial lakes in desert basins tend to be closed-basin, or subthreshold, lakes rather than open-basin lakes. As a result, they may exist for long periods of time because they are not destroying their own basin closure by fluvial erosion at the threshold. More importantly, by not being threshold controlled, the lakes are free to fluctuate in level in response to changes in effective moisture leaving telltale coastal landforms at a variety of water stillstand levels. The largest desert lakes in existence today are those like the Salton Sea in California that are supplied by exotic streams which originate in distant regions, those like Lake Eyre in Australia with very large drainage basins, lakes that lie in drainage basins which have some terrain outside of the arid climatic regime, such as the Dead Sea in Israel, and hemiarid lakes like Pyramid Lake in Nevada, which are those fed by adjacent nonarid highlands (Fig. 25.2) (Currey 1994, Wilkins and Currey 1997). Note that these categories are not all mutually exclusive.

When an arid region undergoes a climate change to circumstances of greater available moisture, existing perennial lakes expand while new closed-basin lakes become established in basins that previously held only playas. Because of late Pleistocene and Holocene climate fluctuations, many now-desert basins of the middle and subtropical latitudes display considerable geomorphic and sedimentological evidence of having contained larger lakes during various times of greater effective wetness in the late Quaternary. These are sometimes referred to as pluvial lakes, but that term is discouraged since it implies that the climate responsible for them was only rainier than present with no change in temperature or other influential variables. Regardless, the return to arid conditions, with its concomitant sparse vegetation and limited weathering rates, has left much of the palaeolake evidence well preserved, visible, exposed, and accessible to scientific study (Fig. 25.3). Researchers have enumerated about 100 late Quaternary palaeolakes in the Basin and Range province of the western US alone (Williams and Bedinger 1984), with Lakes Bonneville and Lahontan being the largest (Fig. 25.2). Considerable scientific attention has also been directed toward palaeolakes on the Altiplano of Bolivia, Peru, Chile, and Argentina (Valero-Garcés et al. 1999, Fornari et al. 2001, D'Agostino et al. 2002); Megalake Chad in North Africa (Leblanc et al. 2006a); Lake Lisan and others in the Jordan-Dead Sea Rift Valley (Stein 2001, Bartov et al. 2002, Hazan et al. 2005, Migowski

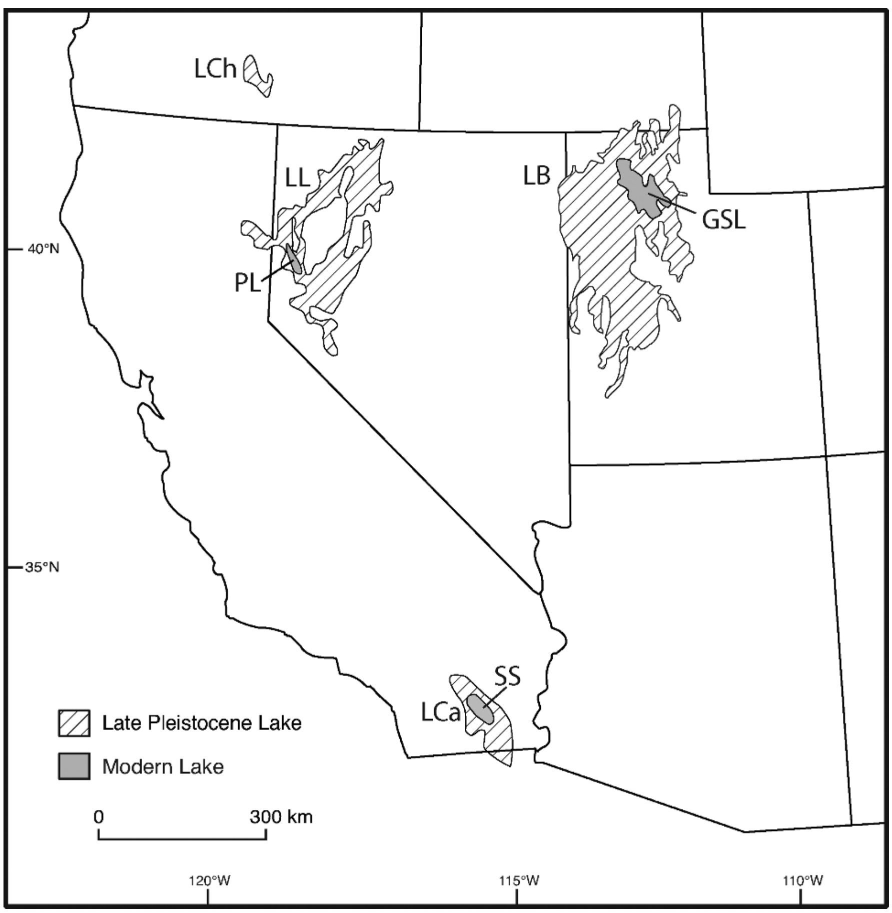


**Fig. 25.2** Location of selected late Pleistocene and modern lakes of western North America (those discussed in the text). SS = Salton Sea, GSL = Great Salt Lake, LB = Lake Bonneville, PL = Pyramid Lake, LL = Lake Lahontan, LCh = Lake Chewaucan. The Salton Sea is shown here in relation to Lake Cahuilla (LCa), which occupied that basin in the late Pleistocene

**Fig. 25.3** An impressive set of relict shorelines created by late Pleistocene Lake Bonneville, Utah

et al. 2006); predecessors of Lake Eyre in Australia (Croke et al. 1998, Nanson et al. 1998, Stone 2006); and Megalake Tengger and others in northwestern China and Mongolia (Qin and Huang 1998, Peck et al. 2002, Zhang et al. 2004, Gao et al. 2006, Jiang et al. 2006).

## Geomorphology of Desert Basin Palaeolakes

Relict coastal landforms still visible in the arid, subaerial landscape typically constitute the most obvious and compelling evidence that a sizeable lake once existed in a now-desert basin (Avouac et al. 1996). It is only through the identification, correlation, and mapping of preserved shoreline segments that the spatial extent of the water body can be accurately reconstructed and its elevation and surface area determined (Migowski et al. 2006). Palaeolake surface area, as discussed later in this chapter, is a fundamental variable for assessing palaeoclimatic conditions. In addition to the highest water level attained by the lake, it is often desirable to delineate the extent of prominent lower shorelines, which may also mark important climatically induced stillstands or oscillations of the water plane. In some cases these lower shorelines are only visible in stratigraphic exposures because of burial by later lacustrine or subaerial deposits.

Identifying, correlating, and mapping segments of a given shoreline, even a prominent one, can be challenging. Although when it was formed the shoreline demarcated the complete perimeter of the lake, reworking or burial due to subsequent lacustrine processes and postlacustrine attack by especially fluvial, alluvial fan, and aeolian processes obliterate geomorphic evidence of numerous shoreline segments (Fig. 25.4) (Sack 1995). Simple contour tracing is rarely an option for shoreline mapping because of local postlake geomorphic rearrangement of the landscape. In addition, palaeolake shorelines, which were horizontal when created, can be warped by hydroisostatic rebound and offset in places by tectonism (Lambert et al. 1998, Adams and Wesnousky 1999). Hydroisostatic rebound is caused by a reduction in water load and elevates a shoreline from its original position by a distance that depends on the amount of unloading. The shallowest water, and therefore the smallest amount of unloading and rebound, will occur near the margin of a lake basin. Both isostatic and tectonic processes have impacted the impressive relict shorelines of late Pleistocene Lake Bonneville, for example, with maximum differential rebound of 74 m for the highest shoreline (Figs. 25.5 and 25.6) (Currey 1982). Another problem in shoreline mapping stems from the fact that desert piedmont

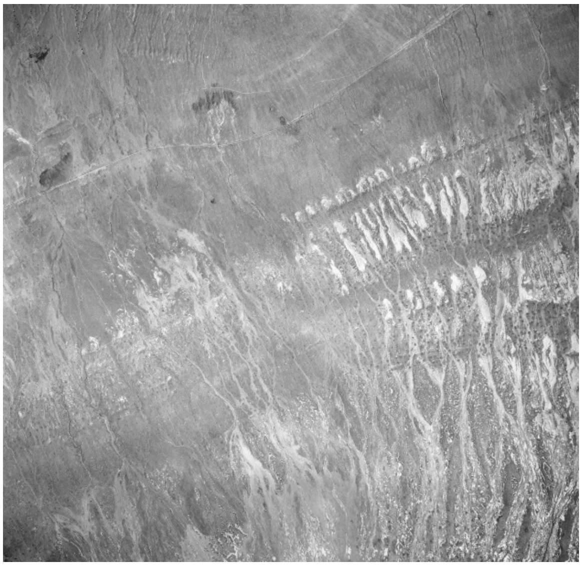

**Fig. 25.4** Portion of a 1:20,000-scale vertical aerial photograph (AAH-14W-99) showing preserved and eroded shorelines and sediments of Lake Bonneville

escarpments can be made by tectonic, mass wasting, fluvial, aeolian, coastal, or other origins, and it is sometimes difficult to identify a bluff with certainty as a relict coastal feature (Gilbert 1890, Knott et al. 2002, Hooke 2004).

Geomorphology is also key to distinguishing those aspects of a palaeolake chronology that reflect climatic conditions from those that result from hydrographic conditions. Climatically induced changes in effective moisture cause fluctuations in lake level but so do basin geomorphic factors. It is critical, for example, to identify any periods of threshold control that might have occurred during the lake's existence. The stabilizing effect of external drainage on the level of a lake has already been noted. If a period of exterior drainage goes unrecognized, the threshold-controlled shoreline will be attributed to a prolonged stability in effective moisture, which probably did not occur. Conversely, during an open-basin phase of a lake, positive or negative changes in the elevation of the threshold due to volcanism, tectonism, fluvial erosion, or mass movement can cause the lake level to rise or fall without a climate change. The geomorphic event of threshold failure, and not a climatic event, caused Lake Bonneville to drop 104 m from the Bonneville to the Provo shoreline in less than a year (Fig. 25.7) (Gilbert 1890, Jarrett and Malde 1987, Burr and Currey 1988).

Geomorphic effects of isostasy produce changes in shoreline position that could be misinterpreted as climatic responses. The highest shoreline of Lake Bonneville was formed as a result of an extended period of threshold control. While at that level, hydroisostatic loading caused the central portion of this large lake basin to subside relative to the outlet (Currey 1980, Currey et al. 1983, Burr and Currey 1988). As a result, in basin interior locations the lake left a record of apparent transgressions as the water plane impinged on terrain that was subsiding from a subaerial to a subaqueous position.

Subbasins dynamics is yet another geomorphic element that can play an important role in controlling the level of a palaeolake (Fornari et al. 2001, Sack 2002, Brown et al. 2003). Many palaeolakes, including for example Lakes Bonneville, Lahontan, and Chewaucan in western North America (Fig. 25.2) (Eardley et al. 1957, Allison 1982, Benson and Thompson 1987, Sack 2002), Lake Chillingollah in Australia (Stone 2006), and Lake Lisan in Israel (Bartov et al. 2002), consisted of a collection of Subbasins separated from each other by interior thresholds (Fig. 25.8). Each Subbasins had a unique interval of integration with the main palaeolake determined by its local hydrologic balance and the elevation of the dividing interior threshold. Some Subbasins contained isolated, independent palaeolakes before and after their integration period with the larger water body (Allison 1982, Sack 2002). During the transgressive phase of a lake cycle, the water level will naturally rise at different rates in different Subbasins. When it reaches the elevation of the lowest interior threshold of the primary Subbasins it will flow over that divide into an adjoining closed Subbasins. Unless there is significant erosion or slope failure at the interior threshold, the water level in the main basin will be held approximately constant while the water body in the Subbasins undergoing integration rises to equilibrate with it. A shoreline will form in the main basin as the result of the stillstand, whereas the water level can rise too quickly to leave shoreline evidence in the adjacent, filling Subbasins (Sack 1990). The Subbasins nature of a palaeolake must be thoroughly investigated so that stillstands and rapid rises in lake level caused by Subbasins integrations and isolations are not given climatic interpretations. Most large palaeolakes consisted of Subbasins and underwent complex Subbasins dynamics.

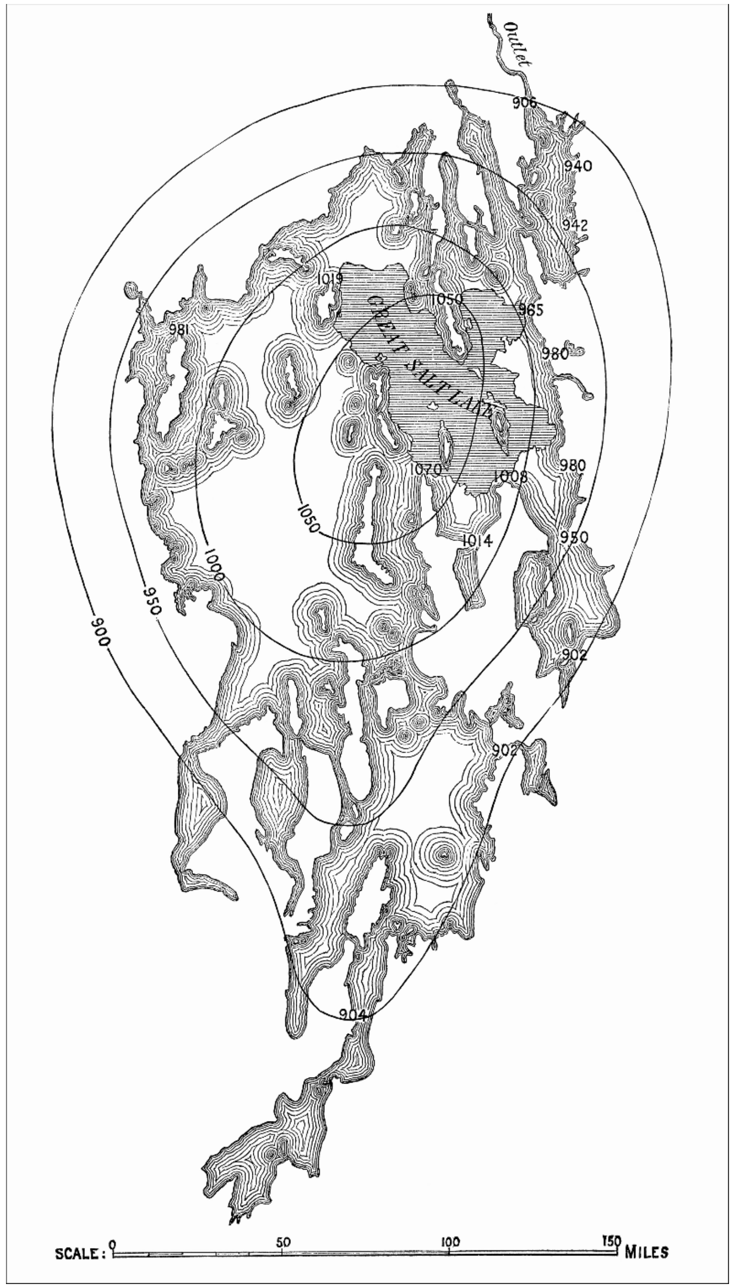


**Fig. 25.5** Theoretical curves of post-Bonneville deformation (Gilbert 1890, Plate L). The map depicts the general pattern of isostatic rebound of the highest shoreline of Lake Bonneville, called the Bonneville shoreline. Units are in feet above Great Salt Lake. The location of Lake Bonneville appears in Fig. 2

Drainage basin dynamics also lead to geomorphically induced fluctuations in the level of a terminal lake. An increase or decrease in the drainage basin area resulting from tectonism, volcanism, mass movement, or stream capture alters stream flow, which causes closed-basin lake level fluctuations without a change in climate. The Holocene successor to Lake Bonneville, Utah's Great Salt Lake (Fig. 25.2), receives its greatest inflow from the Bear River. This large river may have been diverted into the Bonneville basin between the last two major lake cycles (Bouchard et al. 1998, Hart et al. 2004), which occurred during marine oxygen-isotope stages (MIS) 2 and 6. The addition of the Bear River as a tributary could have contributed to the MIS 2 Bonneville basin palaeolake rising approximately 42 m higher than the MIS 6 palaeolake and attaining open-basin status (Currey 1982, McCoy 1987).

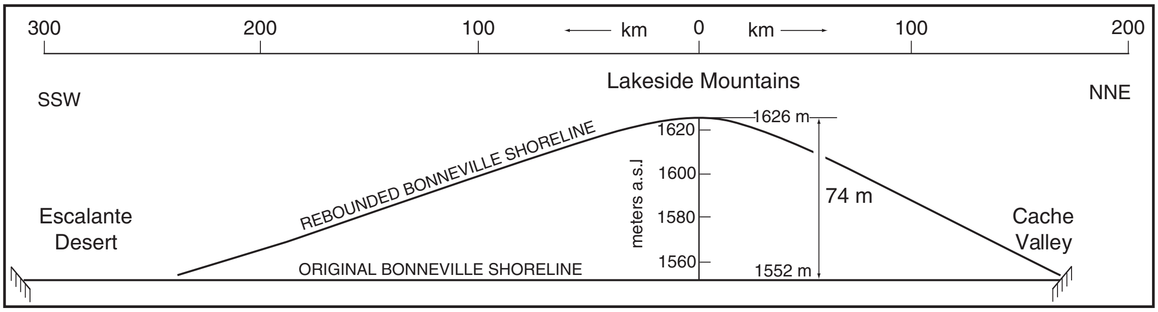


**Fig. 25.6** Transect from south-southwest to north-northeast across the Bonneville basin showing the modern (rebounded) elevation of the Bonneville shoreline in relation to the elevation at which it was created (after Currey 1990, p. 203). This shoreline marks the maximum extent of Lake Bonneville. The deepest part of the lake basin lies near the center of the basin adjacent to the Lakeside Mountains, therefore preserved shoreline remnants located there display the greatest amount of hydroisostatic rebound

## Geomorphic Techniques

Geomorphologists study desert-basin palaeolakes with a variety of field and laboratory techniques. Field geomorphic, and related sedimentary, observations and measurements remain fundamental to palaeolake studies, as does morphostratigraphy, which uses the form of sediment packages in stratigraphic exposure to interpret landforms subsequently buried by other sediments (Fig. 25.9). Stereoscopic interpretation of aerial photographs remains a valuable asset for shoreline mapping (Nanson et al. 1998). Air photo mapping requires close inspection of shoreline landforms on large- and intermediate-scale air photos. This process also aids in the identification of stage-specific geomorphic signatures, which may reflect important aspects of the palaeolake history, and in the identification of well-developed or well-exposed sites for detailed field investigation. Digital elevation models (DEMs) help researchers reconstruct shorelines and contend with the

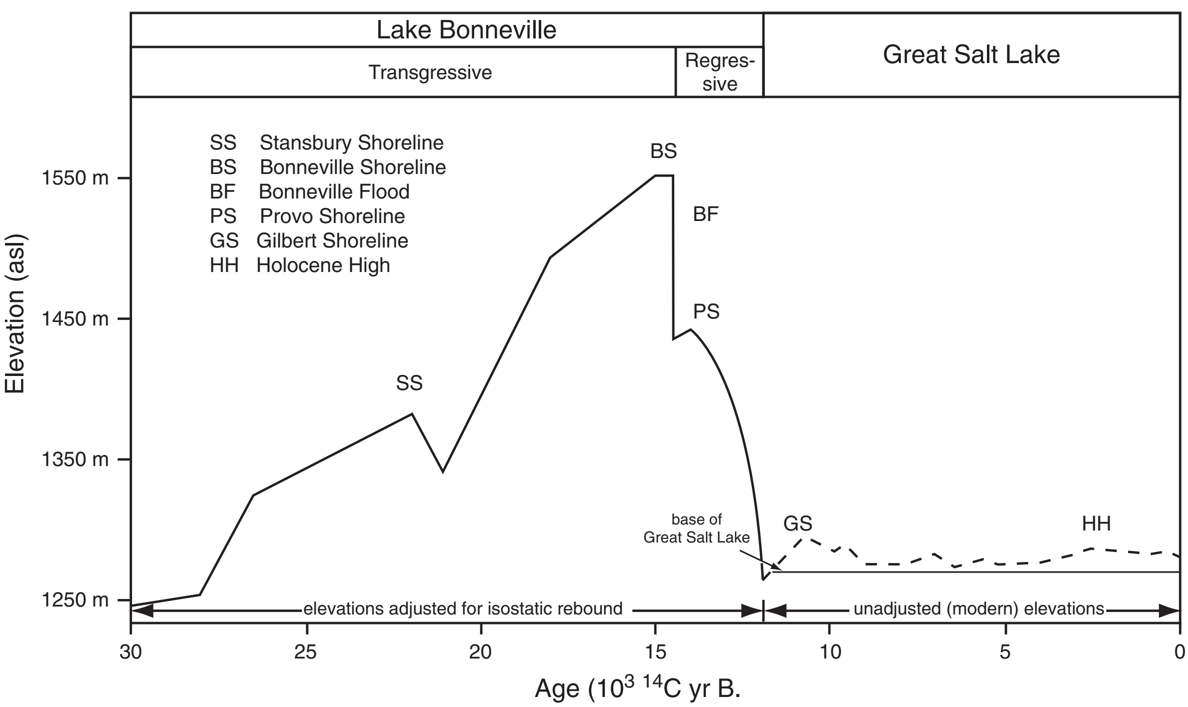


**Fig. 25.7** Generalized hydrograph of Lake Bonneville and its Holocene successor, Great Salt Lake

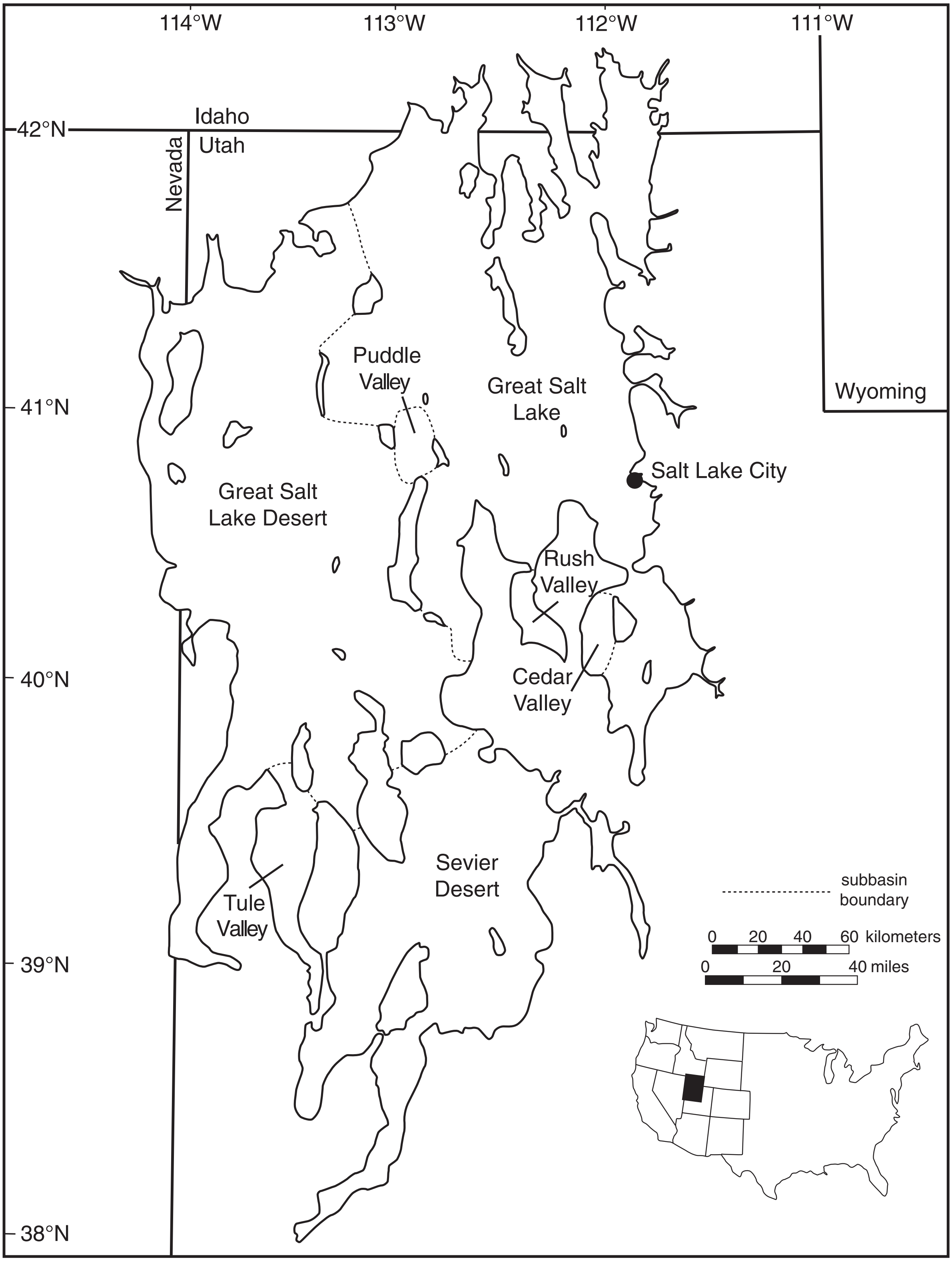


**Fig. 25.8** Major subbasins of Lake Bonneville

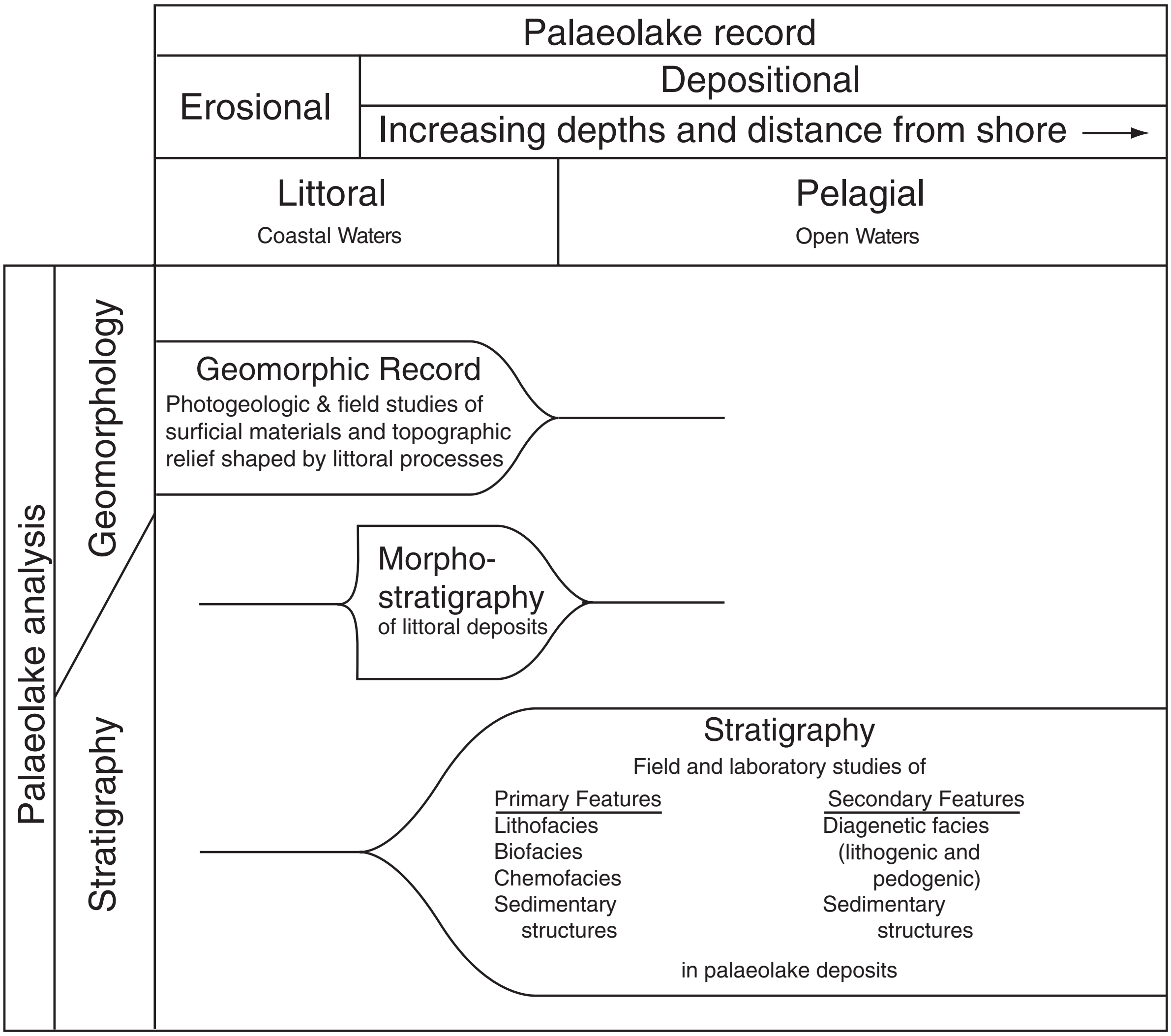


**Fig. 25.9** Morphostratigraphy and its relationship to geomorphology and stratigraphy in palaeolake studies (after Currey 1990, p. 200). Sedimentology is a fundamental constituent of all three forms of analysis, being largely inapplicable only for the geomorphic study of erosional landforms

problem of discontinuous preservation of the shoreline perimeter (DeVogel et al. 2004). Palaeolake investigators use DEMs, radar topographic data, and various sources of satellite imagery to identify shorelines and thresholds particularly in regions with poor accessibility (Komatsu et al. 2001, Schuster et al. 2003, Leblanc et al. 2006a,b, Ghoneim and El Baz 2007). Palaeolake landforms submerged under present water bodies have been identified with depth-profiling techniques (Ricketts et al. 2001), whereas data handling, visualization, and virtual filling of palaeolakes are accomplished with GIS and computer modelling (DeVogel et al. 2004).

Obtaining accurate measurements of shoreline elevation and reliable numeric age determinations are necessary for shoreline segment correlation, for assessing amounts and rates of hydroisostatic and neotectonic offsets, and for constructing detailed palaeolake time vs. water level graphs. At present, the most accurate elevations are acquired in the field with electronic total stations, provided sufficient bench marks are available. Differential global positioning systems are also quite useful for field measurements of shoreline elevation (Hoelzmann et al. 2001). Algorithms are constructed to determine the original elevation of a shoreline that has been subjected to hydroisostatic rebound (Currey and Oviatt 1985). In most cases this rebound-free, or derebounded, elevation is what should be plotted on hydrographs of large palaeolakes (Fig. 25.7).

Portraying the time factor on a palaeolake hydrograph depends on finding datable material in unequivocal stratigraphic context with respect to a known water level. This is often quite challenging in desert palaeolake basins (Geyh et al. 1999). Accelerator mass spectrometer (AMS) radiocarbon dating, uranium-series (U-series) dating, amino acid dating, and occasionally optically stimulated luminescence (OSL) and tephrochronology, offer the most precise age determinations. Numeric ages based on radiocarbon, however, which are the most commonly employed, suffer from some unknowns, depending on the type of material dated. Carbonate samples, as from shells and tufa, might be contaminated with younger carbon or with older carbon derived from fossil water, which is called the reservoir effect (Benson 1993). Some organisms, moreover, tend to use $^{14}$C-deficient carbon in making their shells so that they do not reflect the $^{14}$C balance of the water body (Pigati et al. 2004). Relative age estimates and correlations between shoreline segments have been determined with such techniques as degree of shoreline development or degradation (Wilkins and Currey 1997, Hooke 1999), rock varnish accumulation (Liu et al. 2000), soil development (Adams and Wesnousky 1999, Stone 2006), and cross-cutting by features of known age.

Clearly, geomorphology is a major source of information concerning palaeolakes and their fluctuations, but that data source also has its limitations. The geomorphic record is naturally weak for small palaeolakes with limited fetch, consists of negative evidence for rapid changes in lake level, may be reworked by later water-level oscillations or buried by later lacustrine sediments, and becomes increasingly obliterated by subaerial processes with increasing time since exposure (Sack 1995). Fortunately, the record of accumulated palaeolake sediment supplies additional insights into the nature of the palaeolake and its regional environment.

## Sedimentology and Stratigraphy of Desert Basin Palaeolakes

The size of a palaeolake is directly indicated only by the position of its shoreline as marked by erosional and depositional coastal landforms; materials that have been deposited within the lake provide other significant palaeoenvironmental information. Indeed, a continuous record of a palaeolake's sedimentation history will exist in the deepest part of the basin if the lake did not experience complete desiccation during which sediment was lost through deflation (Magee et al. 1995, Nanson et al. 1998). Completeness of the sediment record decreases with proximity to the shoreline.

Palaeolake sediment records are studied primarily from sediment cores and outcrops, but seismic profiles (Valero-Garcés et al. 1996) and ground penetrating radar images have also been used. Researchers have developed an impressive array of approaches for gleaning palaeoenvironmental data from lacustrine sediments. A vertical increase or decrease in the grain size of lithic fragments and sediment density suggests a falling or rising water level, respectively (e.g. Davies 2006). Pollen reflects the climate of the drainage basin (e.g. Mensing 2001, Zhang et al. 2002). Salinity, relative water depth, and/or subsurface versus surface sources of lake water are commonly investigated with diatom and ostracod assemblages, isotope geochemistry, elemental chemistry, carbonate content, amount of organic carbon, total inorganic carbon, concentration of magnetic minerals, and sedimentary structures (e.g. Benson et al. 1998, Ricketts et al. 2001, Balch et al. 2005, Flower et al. 2006). Multiple proxies from the same interval, however, sometimes indicate conflicting climatic signals (Dearing 1997, Grosjean et al. 2003). In addition, none of these analyses reveal exact lake size, although when multiple cores are retrieved from across a palaeolake basin, their correlation discloses some spatial characteristics of the palaeolake. The sediment record from cores, on the other hand, contains a much greater abundance of materials that indicate age than do coastal landforms. AMS radiocarbon, U-series, amino acid, and tephrochronology analyses are the most commonly employed means of determining the age of a fluctuation interpreted from cores. These, plus various thermoluminescence techniques, provide reliable age determinations from outcrop samples.

## Palaeoclimatic Reconstruction

Once a palaeolake chronology has been reliably reconstructed and the influences of geomorphically induced versus climatically induced water-level changes have been identified, it is still difficult to deduce specific palaeoclimatic variables from what is essentially a palaeohydrologic record. Even if only climatic variables were involved, there are multiple climatic scenarios that could have resulted in a rise in lake level. Some of these include increased average annual precipitation, decreased average annual evapotranspiration, increased cool season precipitation, decreased warm season evapotranspiration, or combined changes in precipitation and evapotranspiration on an average annual or seasonal basis. Evapotranspiration itself, moreover, responds to a variety of atmospheric variables, such as temperature, cloudiness, and windiness. Because an increase in precipitation would result in more vegetation, an increase in evapotranspiration should accompany a precipitation increase as well (Mifflin and Wheat 1979, Qin and Huang 1998). The most likely scenario incorporates changes in both precipitation and evapotranspiration.

An important link between the geomorphic evidence of lake size and the specific climatic variables responsible for it consists of the $z$ ratio of lake surface area, $A_L$, to tributary basin area, $A_B$, with the latter consisting of the drainage basin area excluding $A_L$ (Snyder and Langbein 1962, Mifflin and Wheat 1979, Street-Perrott and Harrison 1985):

$$z = A_L/A_B. \tag{25.1}$$

For a closed-basin lake that has insignificant groundwater transfer and annual input of water equal to annual output, the water balance is represented by the equation:

$$R + A_L P_L = A_L E_L, \tag{25.2}$$

where $R$ is annual tributary runoff into the lake, $P_L$ is direct precipitation onto the lake, and $E_L$ is evaporation from the lake. Runoff, however, can be expressed in terms of tributary basin precipitation, $P_B$, and tributary basin evaporation, $E_B$:

$$R = A_B(P_B - E_B). \tag{25.3}$$

By substituting for $R$, Equation (25.2) becomes:

$$A_B(P_B - E_B) + A_L(P_L) = A_L(E_L) \tag{25.4a}$$

$$A_B(P_B - E_B) = A_L(E_L - P_L) \tag{25.4b}$$

$$A_L/A_B = (P_B - E_B)/(E_L - P_L). \tag{25.4c}$$

Therefore,

$$z = A_L/A_B = (P_B - E_B)/(E_L - P_L). \tag{25.5}$$

Through the $z$ ratio, the geomorphic evidence of lake area and basin area is related directly to the palaeoclimatic variables of precipitation and evaporation (Snyder and Langbein 1962, Mifflin and Wheat 1979, Street-Perrott and Harrison 1985). Variables $A_L$ and $A_B$ are reconstructed from relict shoreline evidence of late Pleistocene palaeolakes, but determining values for the evaporation and precipitation variables represents a greater challenge. A common approach consists of using modern relationships among the study region's temperature, precipitation, evapotranspiration, and runoff to estimate values for the palaeoclimatic variables, within limits set by such proxies as pollen, and solving iterations of the formula until the $z$ ratio calculated from the climatic variables approaches the $z$ ratio obtained from the geomorphic data (Mifflin and Wheat 1979, Barber and Finney 2000, Menking et al. 2004). Although a unique solution does not exist, sensitivity tests reveal the most likely ranges of palaeoclimatic variables that would have resulted in a palaeolake of the observed extent (Bookhagen et al. 2001, Jones et al. 2007).

Any discrepancies between geomorphically derived values of the $z$ ratio and values derived by estimating the palaeoclimatic variables may be due to (a) errors in determining $A_L$ and $A_B$, (b) imprecision in the relationships established for modern data (Kotwicki and Allan 1998), or (c) the possibly inaccurate assumption that the relationships among the palaeoclimatic variables can be adequately modelled by the relationships among the modern climatic variables (Mifflin and Wheat 1979, Benson and Thompson 1987). The first type of error is minimized by careful mapping and fieldwork and by employing in calculations only those lake basins that have well preserved geomorphic evidence which can be mapped with a high degree of confidence. In this regard, late Quaternary palaeolake basins in arid regions probably have the greatest potential. The second source of error will no doubt decrease as geomorphologists and climatologists continue to investigate the modern empirical relationships among the relevant climatic and hydrologic variables. Validity of the assumption that the empirical relationships remain uniform over time can be checked by comparing palaeoclimatic inferences drawn from palaeolake data with inferences drawn from other sources of palaeoenvironmental evidence. Potential error in reconstructing palaeoclimatic variables from palaeolake data can be minimized by using information from multiple proxies and/or many basins to investigate regional trends instead of focusing on single data sources, and recent research has moved in these directions. Further accuracy should derive from balancing energy (Bergonzini et al. 1997) and isotopic budgets (Jones et al. 2007) in addition to the hydrologic budgets, and this approach holds considerable promise for palaeoclimatic research based on desert basin palaeolakes.

## Conclusions

Specific values of palaeotemperature and palaeoprecipitation cannot yet be determined with certainty from the palaeohydrologic evidence of desert basin palaeolakes, but approaches continue to become more sophisticated and multivariate. In the meantime, with increasing concern for global climate change and human impacts on the environment, a growing body of palaeolake research focuses on characterizing the relative amplitude, duration, and chronology of past changes in regional effective moisture, explaining these in terms of altered atmospheric circulation patterns, especially shifts in the jet streams and in monsoonal circulation, and correlating palaeolake fluctuations with climatic events represented in the marine oxygen-isotope record and Greenland ice cores. This broader, i.e., more global, perspective remains grounded in the science of reconstructing in detail the fluctuation chronology of individual palaeolakes. Palaeolake researchers accomplish this using all of the tools at their disposal – with fundamental geomorphic, morphostratigraphic, and sedimentologic/stratigraphic methods and technologically evolving techniques.

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