

CHAPTER 3**A Model of Runoff, Evaporation, and Overspill in the Owens River
System of Lakes, Eastern California**

KIRSTEN M. MENKING and ROBERT S. ANDERSON

*Department of Earth Sciences and Institute of Tectonics,
University of California, Santa Cruz, CA 95064***ABSTRACT**

We present a numerical model of water balance for the Owens Lake chain of lakes, southeastern California. These lakes are separated by a series of bedrock sills, have received their water primarily from precipitation in the adjacent Sierra Nevada, and overflowed during extremely wet periods. While the modern climate in the region does not support the existence of lakes in any valley but Owens, shoreline and core evidence suggests that all valleys contained significant lakes during parts of the Pleistocene. Our modeling effort attempts to determine what combinations of runoff and evaporation were necessary to create these large lakes, what the response time of each lake is to perturbations in climatic variables, and whether the lakes are likely to have achieved steady-state with respect to long-term climatic fluctuations.

The model explicitly incorporates the hypsometry of each lake, derived from 3 arc-second DEM data. Runoff is applied and lake volume updated; detailed lake basin hypsometry is employed to calculate new depth and surface area. Evaporative flux of water from each lake is calculated by multiplying lake surface area by an assumed evaporation rate. When lake volume exceeds the maximum volume dictated by the spillway elevation, excess volume is transferred to the next lake in the chain. The $\delta^{18}\text{O}$ of the model Owens Lake is also calculated for comparison with $\delta^{18}\text{O}$ values measured on carbonates from a core of Owens Lake.

Results suggest that all of the lakes achieve steady-state volumes within a matter of decades, indicating that droughts or particularly wet periods of order several

decades to centuries should have been recorded in the sediments of the Owens Lake core, and that all lakes should achieve steady-state altitudes with respect to yet longer-period climate variations. Furthermore, modeled $\delta^{18}\text{O}$ values also reach steady state, with periods of overflow showing depleted values close to Owens River water, and periods of low lake level showing enriched values. "Maximum pluvial" lake stages require an increase in the flux of Owens River water, relative to modern values, of nearly an order of magnitude. Finally, comparison of a modeled lake-level history and oxygen isotopic composition to the lake-level reconstruction of Jannik and others (1991) and to the $\delta^{18}\text{O}$ record for the Owens Lake core (Benson and Bischoff, 1993) shows general agreement.

INTRODUCTION

A chain of Pleistocene lakes, headed by Owens Lake, lies to the south and east of the Sierra Nevada Mountains in southeastern California (Fig. 3.1). Abandoned shorelines found high above valley floors testify to the previous greatness of these lakes which are now all dry playas. The lake basins are separated from each other by a series of bedrock sills (Fig. 3.2) and are thought to have received most of their water from precipitation in the Sierra Nevada (Smith, 1976; Smith and Street-Perrott, 1983). When Owens Lake exceeded its confines, it overflowed into the shallow China Lake basin, which rapidly filled and overspilled into the Searles Lake basin. During exceptionally wet periods, China and Searles Lakes coalesced, filled to a common sill level, and spilled into Panamint Valley and ultimately into Death Valley (Smith and Street-Perrott, 1983); surface area for the chain of lakes exceeded 9x the modern value. In this paper, we present a numerical water balance model used to address a number of questions including: what combinations of inflow, evaporation, and overspill are overspilled into the Searles Lake basin. During exceptionally wet periods, China and Searles Lakes coalesced, filled to a common sill level, and spilled into Panamint Valley and ultimately into Death Valley (Smith and Street-Perrott, 1983); surface area for the chain of lakes exceeded 9x the modern value. In this paper, we present a numerical water balance model used to address a number of questions including: what combinations of inflow, evaporation, and overspill are required to produce lakes with observed shoreline

elevations? What is the characteristic response time of each lake if it is initially dry and then allowed to approach steady state, and what factors control this response time? Are lakes at steady-state with respect to long time-scale climatic fluctuations?

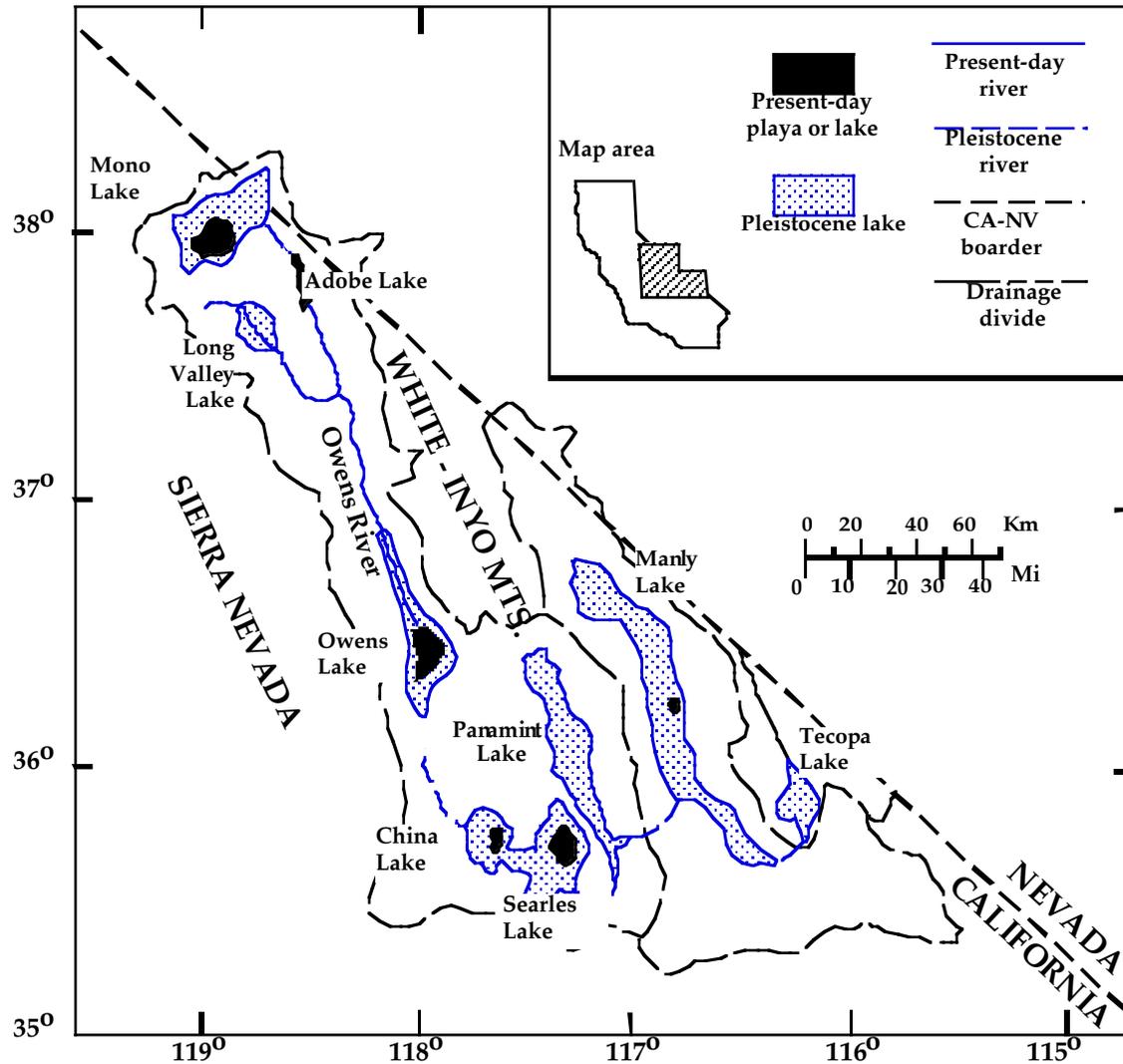


Fig. 3.1: Map of the Owens Lake chain of lakes in eastern California. Black areas represent modern lakes and playas. Grey areas represent the extents of Pleistocene pluvial lakes. Two of the three routes for introduction of water into Death Valley are shown (from overspill from Panamint Lake or from Tecopa Lake).

Lake Hypsometries Used in the Model

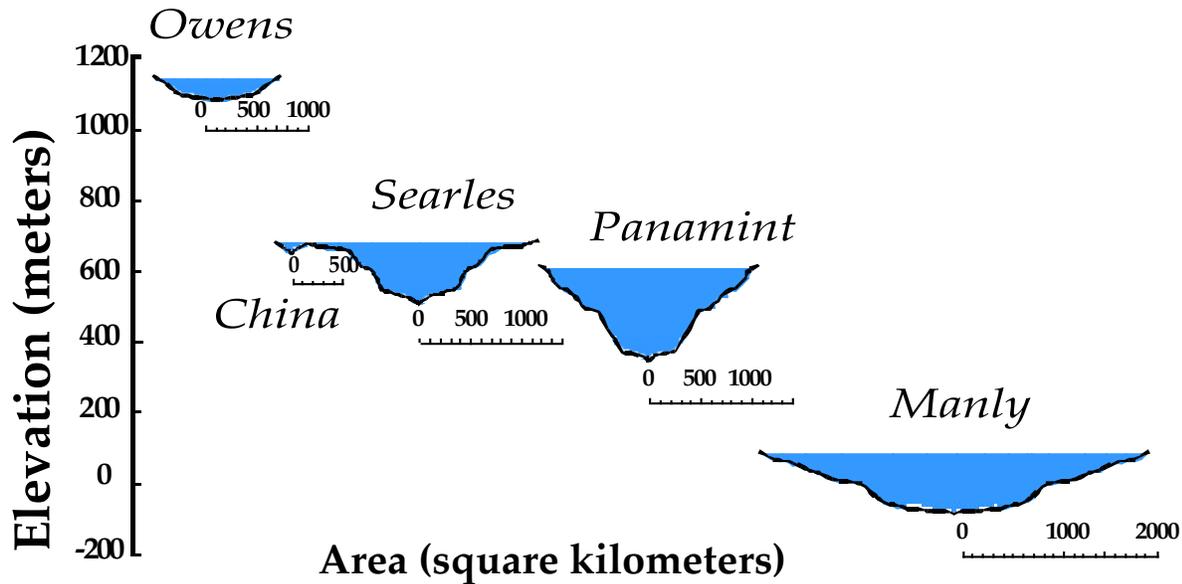
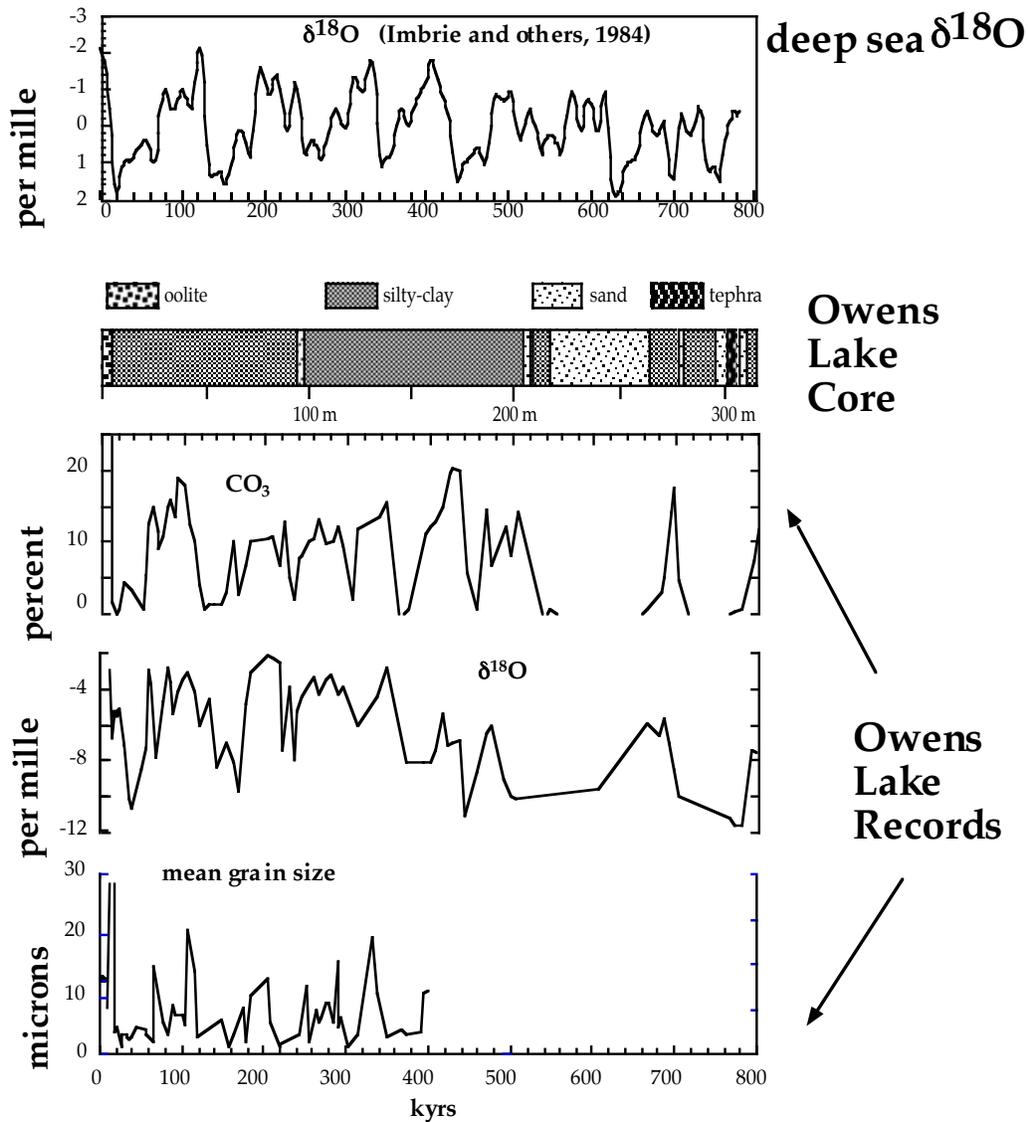


Fig. 3.2: Diagram of each lake's hypsometry and elevational relationship to the other lakes in the chain. Spillway elevations and basal elevations reported in table 1. When China Lake reached an altitude of 665 m, it spilled into the Searles Lake basin. Searles then filled to the China Lake spillway at 665 m, and the two lakes coalesced, filling to an ultimate level of 690 m. Further influx of water prompted spill into the Panamint Lake basin.

In 1992, the U.S. Geological Survey cored Owens Lake to obtain a continuous record of terrestrial paleoclimate in the Sierran region (Smith, 1993). Several paleoclimatic proxies, such as mean grain size (Menking, in press), oxygen isotopic composition (Benson and Bischoff, 1993), and carbonate content (Bischoff and others, in press (a)) bear strong resemblances to the oxygen isotopic record from deep-sea sediments (Fig. 3.3). These proxies suggest that glacial periods saw a fresh, isotopically depleted, overflowing Owens Lake while interglacial periods were marked by a shallow, saline, and isotopically enriched Owens Lake. In this paper, we use the measured carbonate oxygen isotopic composition determined by Benson and Bischoff (1993) as an additional constraint on variations in runoff and evaporation necessary to create large Pleistocene lakes.

To investigate how the Owens chain of paleo-lakes responds to variations in



climate, we perform a series of experiments with the water-balance model. We first

Fig. 3.3: Paleoclimatic proxies from the Owens Lake core compared to the marine oxygen isotopic record from Imbrie and others (1984). General agreement between the marine record and the Owens Lake records suggests that the regional climate in eastern California is being driven by the same processes that influence global climate. All three Owens Lake proxies are thought to reflect fluctuations in lake size. Carbonate content reflects changing salinity of Owens Lake; during glacial periods (such as at 20 ka) carbonate content is very low, suggesting that Owens Lake is fresh and spilling over into the China Lake basin. In contrast, during interglacials (such as present and at 100 ka) carbonate content reaches >15% indicating that the lake is shallow and saline. Oxygen isotopic content of carbonates precipitated in equilibrium with lake water probably also reflects lake level. Benson and Bischoff (1993) have interpreted values more enriched than -4 ‰ as indicative of closed lake conditions. Finally, mean grain size at the core site is finest during glacial periods, and coarser during interglacials. The close correspondence of the grain size and carbonate content records indicate that both are responding to changes in lake size.

conduct steady-state calculations in which we determine the response time of each lake to a constant input of runoff and constant evaporation rate. We next allow runoff and evaporation to oscillate sinusoidally between modern and assumed-maximum-glacial conditions with different periods of oscillation. To simulate maximum-glacial conditions, we use glacial-geologic and paleobotanical evidence of past temperature and precipitation rates for the southwestern Great Basin. These oscillatory climate experiments serve to confirm the response times determined in the steady-state experiments, and to explore whether water balance is controlled primarily by runoff or by evaporation. Finally, we drive runoff and evaporation with a more realistic forcing function, the $\delta^{18}\text{O}$ curve for deep sea sediments (Imbrie and others, 1984). We assess our reconstruction of water-balance history by tracking Owens Lake water oxygen isotopic composition for different model runs. Goodness of fit between modeled isotopic composition and the measured carbonate oxygen isotopic composition determined by Benson and Bischoff (1993) for the Owens Lake core is assessed visually. We also compare our modeled lake-level record to a lake-level history constructed by Jannik and others (1991), who used stratigraphic and Cl-accumulation evidence from cores of Searles and Panamint lakes. The model is run for the time period 200 ka to the present, such that two large glaciations (associated with isotope stages 2 and 6) and one large interglacial (isotope stage 5) are represented. Before proceeding to a description of the model, we first outline the geologic and climatic setting of the study area, and review previously conducted work.

TECTONIC AND CLIMATIC SETTING

The Owens River system of lakes lies to the east and south of the Sierra Nevada Mts. in the Basin and Range tectonic province. Each lake resides in a downdropped fault-block valley bounded by essentially north-south trending ranges. Relief between valley floors and range crests is greater than 3000 m in some locations. All of the valleys are closed topographic basins with internal drainage networks. Extensive warping of previously-existing lake shorelines, visible fault scarps, modern seismic activity, and the presence of young volcanic cones and flows in many valleys are evidence of the continuing tectonic activity in the region (Bierman and others, 1991; Hollett and others, 1991; Smith, 1976; Densmore, 1994). In addition to volcanic deposits, each valley is

filled with alluvial fan sediments derived from the surrounding ranges, fluvial deposits associated with the internal drainage network, and lacustrine deposits associated with the Owens chain of lakes.

Precipitation in the Owens Valley drainage basin is derived chiefly from the Pacific Ocean and falls primarily over the eastern flank of the Sierra Nevada, which receives some 60-80% of the rain and snow falling on the valley (Hollett and others, 1991). Most of the precipitation falls during the winter months (Benson and Thompson, 1987a; Smith, 1976), and average precipitation rates range from 13-15 cm/yr on the valley floor to more than 100 cm/yr at the crest of the Sierra. These mountains create a strong rain shadow, cutting the average annual precipitation in the White and Inyo Mountains to 18 to 35 cm/yr (Hollett and others, 1991; California Water Atlas, 1979). Valleys to the east of Owens Valley are located not only in the rain shadow of the Sierra Nevada but also in the lee of other north-south trending ranges. As a result, regional precipitation tends to decrease with distance from the Sierra Nevada, with the lowest precipitation values, <10 cm/yr, being recorded in Death Valley (California Water Atlas, 1979). Those valleys containing paleo-lakes China, Searles, Panamint and Manly currently are marked by very low values of surface runoff, 0.63 cm or less per year. The western flank of the Owens Valley, by contrast, experiences runoff values between 0.63-100 cm/yr with runoff increasing with rising altitude (Langbein and others, 1949; Hollett and others, 1991).

Average monthly air temperature in the valleys ranges from less than 0°C in the winter to more than 30°C in the summer, with mean annual temperature near Owens Lake ~15°C (Hollett and others, 1991; Smith, 1976; Phillips and others, 1992). Winds blow predominantly up- and down-valley, with speeds of up to 50 km/hr. Relative humidity in the Owens drainage basin varies between less than 30% in the summer to more than 40% in winter, and evapotranspiration rates range from 30 to 100 cm/yr (Hollett and others, 1991). These climatic conditions favor the growth of alkaline scrub plant and meadow plant communities. Juniper and pine woodlands are found at higher elevations. Meyers (1962) and Smith and Street-Perrott (1983) report evaporation rates for lakes and reservoirs in several western states. Evaporation rate increases with distance from the Sierra Nevada, with the lowest average evaporation rates found in Owens Valley (75-110 cm/yr), and the highest found in Death Valley (180-220 cm/yr).

Modern climatic conditions do not support the existence of lakes in any valley except Owens. Prior to the construction of the Los Angeles aqueduct (1912) and the attendant diversion of Sierran runoff into the aqueduct, discharge into the Owens River maintained a 14 m deep, 290 km² area Owens Lake (Smith and Street-Perrott, 1983). As a result of the artificial diversions, this lake has since essentially desiccated. Paleo-shoreline evidence (Gale, 1914; Smith, 1976; Blackwelder, 1933 and 1954; Hunt and Mabey, 1966; Hooke, 1972) indicates that climatic conditions during some periods of the Pleistocene must have differed significantly from modern conditions. During what Smith (1984) terms extreme "pluvial" periods, Owens, China, Searles, and Panamint Lakes filled to overflowing, and Manly Lake, in Death Valley, filled to a depth of 173 m. At present, the combined surface area of all of the lakes (modern surface area being 0 km² in the case of China, Searles, Panamint, and Manly Lakes, and taken as the historical lake surface area [290 km²] for Owens Lake) represents only 1.1% of the total combined drainage area contributing to these lakes. At their maximum high-stand elevations, however, the five lakes covered 9.1% of the total combined drainage area. The climatic conditions required to maintain these large lakes are unclear, and provide the motivations for conducting this study.

PREVIOUS WORK AND CONSTRAINTS ON PLEISTOCENE PALEOCLIMATE IN THE SIERRA NEVADA REGION

Paleoclimatic evidence

Gale (1914), Smith (1976), Blackwelder (1933 and 1954), Hunt and Mabey (1966), Hooke (1972), and Smith (1984) described geomorphic evidence for large lakes in the currently dry Owens, China, Searles, Panamint and Death Valleys. This evidence consists of tufa towers and tufa shorelines, well-rounded beach gravels, deltaic deposits, beach-rock, and coquina shorelines incorporating fresh-water gastropods. Unfortunately, attempts at dating shoreline deposits have frequently been frustrated by a lack of datable material, or by post-depositional alteration of carbonate (Fitzpatrick and Bischoff, 1993; Smith, 1976). As a result, much of the information on timing of relative lake level stands has come from core evidence.

The drilling of several cores into Panamint and Searles Lakes in the 1950s (Smith and Pratt, 1957), and into Lake Manly in the 1990s (Lowenstein and others, 1994; Ku

and others, 1994; Li and others, 1994; Roberts and others, 1994) showed a succession of high and low lake-level stands in these valleys with high lake levels marked by silt and clay deposition and low levels marked by evaporite deposition. Table 3.1 lists the core evidence for lake-level fluctuations. Jannik and others (1991) attributed low accumulation rates of chloride in Searles Lake to periods of overspill of Searles Lake into Panamint Valley, that overspill carrying the dissolved load of chloride into Panamint Lake where it was precipitated. In the 200 ka period relevant to this paper, they showed periods of low chloride accumulation in the Searles Lake core, and therefore overspill into Panamint Valley, occurring from 120-150 ka, and again from 10-24 ka. They did not determine any periods of overflow from Panamint Lake to Death Valley. On the other hand, Lowenstein and others (1994), Ku and others (1994), Li and others (1994) and Roberts and others (1994) found evidence for large lakes in Death Valley over the last 200 ka. These authors did not address the question of what was the source of the water in Death Valley during its large perennial lake cycles, however. There are three routes by which water entered Death Valley: through overspill from Panamint Valley to the west, from the Amargosa River to the south, and from the Tecopa Lake system to the east. At present, the relative importance of these sources is not known (Lowenstein, personal communication). Given the Cl accumulation evidence of Jannik and others (1993), which suggests that Searles Lake spilled over into Panamint Valley from 120-150 ka, but not earlier, it appears as though the large lake that existed in Death Valley at 186 ka may have been produced by runoff from the Tecopa and Amargosa entrances to Death Valley, rather than by overspill from Panamint Valley.

In 1992, the U.S. Geological Survey cored Owens Lake to obtain a continuous record of terrestrial paleoclimate in the Sierran region (Smith, 1993). Unlike Searles Lake (Smith, 1984), which underwent several filling and desiccation cycles captured in sedimentologic variations between deep lacustrine clays and evaporites, Owens Lake appears to have contained water for at least 786 ka. A close examination of the sediments in the 323-m-long core OL-92 (the base of the core is at the Bruhnes/Matuyama magnetic reversal at 786 ka) reveals no evidence for a desiccation event at any time except at approximately 5 ka when a 1-2 m thick section of oolites formed. There are no thick evaporite packages or playa assemblages, and there exists

TABLE 3.1: Core and Shoreline Evidence of Past Lake Levels for Several Owens Chain Lakes

<u>Lake</u>	<u>Reference</u>	<u>Paleoclimatic Information</u>	<u>Evidence</u>
Searles	Smith (1984)	Nine ~400 ka duration paleohydrologic regimes ranging from "wet" to "intermediate" to "dry." Some shorter duration fluctuations observed.	930 m KM3 Searles Lake core (3.2 Ma), sediments vary between deep lacustrine muds and evaporites
Searles and Panamint	Jannik and others (1991)	Overspill from Searles Lake to Panamint Valley from 120-150 ka, and again from 10-24 ka. No overspill to Death Valley from 0-200 ka.	Accumulation of chloride in cores from Searles and Panamint Lakes. Chronology of cores established with ³⁶ Cl.
Manly	Lowenstein and others (1994) Ku and others (1994) Li and others (1994) Roberts and others (1994)	Saline pan from 0-10 ka, shallow saline lake from 10-30 ka, saline pan with two small lake cycles from 30-60 ka, mudflat from 98-120 ka, perennial saline lake from 120-128 ka, large perennial lake from 128 to 186 ka, saline lake alternating with salt pan from 186-193 ka. Lake Manly at its deepest (~174 m) at 185 ka. This stage lasted until 130 ka.	Core stratigraphy varies from black deep-lacustrine muds to halite beds. Fabrics in halite beds distinguish saline pan from saline lake environments. Tufa shoreline ages and core chronology established with U-series methods.

no evidence for dissolution of evaporite minerals. Nevertheless, there exists evidence for substantial lake level variations in such paleoclimatic proxy records as mean grain size (Menking, in press), carbonate content (Bischoff and others, in press (a)), and

oxygen isotopic composition of inorganically precipitated carbonate (Benson and Bischoff, 1993) (Fig. 3.3).

Geochemical calculations performed by Bischoff and others (in press (a)) and by Benson and Bischoff (1993) on OL-92 sediments may constrain periods of closed- and open-lake conditions. Bischoff and others (in press (a)) used calcium carbonate accumulation rates to argue that Owens Lake remained closed for periods lasting up to 50 kyrs, separated by shorter (20-30 kyr) durations of overflow. They noted, however, that discrepancies exist between their calculated timing of overflow events and the stratigraphic evidence of overflow into the Searles Lake basin. In particular, from 50-120 ka, these authors found no evidence for overflow, while Bischoff and others (1985) reported stratigraphic evidence for overflow events during this period from a core from Searles Lake. Furthermore, Bradbury (1993) identified several fresh-water diatom species in the Owens Lake core which would indicate spillover between 80-90 ka. Nevertheless, Bischoff and others' (in press (a)) calculations of overflow from Owens Lake compare favorably to Jannik and others' (1991) estimations of overflow of Searles Lake into Panamint Valley, with the former authors suggesting overflows between ~10 and 50 ka, and again between ~110 and 150 ka. Benson and Bischoff (1993) interpreted oxygen isotopic compositions of lacustrine carbonates more enriched than -4‰ (PDB) as indicative of closed lake conditions. If correct, then Owens Lake spilled into the China and Searles basins from 10 to 55 ka, 61-75 ka, 84-94 ka, and 110-176 ka. As the geochemical analyses conducted by Bischoff and others (in press (a)) and by Benson and Bischoff (1993) were performed on samples with an 8 kyr resolution, caution is warranted in ruling out short episodes of overflow in the intervening periods.

The lake records described above require that climate during some parts of the Pleistocene must have been very different from modern. Many workers have addressed the questions of what temperatures and precipitation rates prevailed during glacial periods in the southwestern Great Basin, Table 3.2. The Sierra Nevada are dotted with evidence of multiple Pleistocene (Phillips, 1990) and Holocene (Birman, 1964; Burbank, 1991) glacial advances, and the location of past glacial equilibrium line altitudes calls for

TABLE 3.2: Paleo-Temperature and Paleo-Precipitation Estimates for the Last Glacial Maximum

<u>Reference</u>	<u>Paleoclimatic Information</u>	<u>Evidence</u>
Porter (1983)	Mean annual temperature (MAT) during glaciations 5 °C colder than modern.	Glacial equilibrium line altitudes 800 m lower than modern in Sierra Nevada.
Dohrenwend (1984)	Either no increase in precipitation and 7 °C lowering of MAT relative to modern, or 370 mm increase in precipitation and 5.5 °C lowering of MAT.	Nivation hollow threshold altitude depressed during glaciations; hollows identified in Sierra Nevada and White Mts., among other ranges.
Spaulding and Graumlich (1986)	≤40% increase in precipitation and ~5 °C decrease in temperature in Mojave Desert during last glacial maximum. No increase in precipitation above 36 °N latitude.	Packrat middens
Stamm (1991)	70% increase in precipitation and 6 °C decrease in MAT at Lake Lahontan, ~350 km north of the Owens chain of lakes.	Solutions to a local climate model constructed for the southwestern U.S. Boundary conditions come from global climate models.

a lowering of mean annual temperature by 5°C relative to modern values (Porter, 1983). Several other workers (Dohrenwend, 1984; Spaulding and Graumlich, 1986; Stamm, 1991) have estimated similar regional decreases in temperature for the last glacial maximum. In addition to temperature variations, precipitation is also thought to have increased by 40% (Spaulding and Graumlich, 1986) to 70% (Stamm, 1991) during parts of the Pleistocene.

Paleoclimatic calculations and models

Various models have been constructed to address the effects of changing climate on the water balance of closed-basin lakes. Smith and Street-Perrott (1983) assessed the amount of Owens River discharge necessary to fill successively each lake in the chain, assuming that none of the four lakes down-river from Owens received any water from its own local drainage basin; with various evaporation values, the flow in the Owens River would have to be increased by 5x to >10x modern in order to fill Lake Manly to its high-stand level. Phillips and others (1992) reconstructed lake-surface-area histories for the Owens lake-chain based upon oxygen-isotopic values of carbonates from a core of Searles Lake and upon a water-balance model which related oxygen-isotopic composition of Searles Lake water to different steady climate forcings. They found an episode of overspill from Searles to Panamint at 18-24 ka and from Panamint to Manly Lake at ~140 ka (p. 135). The model of Orndorff and Craig (1994) determines snowpack, runoff, and lake extent for the Owens chain of lakes from temperature, precipitation, and topographic data. This regional climate model was constructed to compare estimates of past temperature and precipitation rates from paleobotanical and glacial-geologic evidence to known extents of lakes, and allows spatial variability in temperature, precipitation, windfield, and topography. Orndorff (personal communication) has informed us that the model predicts a 3.5-fold increase in runoff in the Owens Lake region during the last glacial maximum. Finally, Hostetler and Benson (1994) constructed a predictive model of oxygen and deuterium isotopes for Pyramid Lake, Nevada, based upon several climatic variables including air temperature, relative humidity, windspeed, solar and thermal radiative fluxes, amount of precipitation falling on the watershed, and isotopic composition of precipitation. Given these variables, they calculated the volume of the lake at any time step, the amount of water

evaporating from the lake's surface, and the isotopic composition both of the evaporated water and of the remaining lake water.

The model described in this paper is different from that of Smith and Street-Perrott (1983) and Orndorff and Craig (1994) because it allows us to explore transience in lake levels as climate changes, as well as to determine the time required to reach steady-state-lake conditions given a particular suite of paleoclimatic variables. These authors were confined to looking at end-member, steady-state climatic scenarios, and thus could not generate a lake-level history for the Owens Lake chain of lakes. Furthermore, we make use of the new oxygen isotopic data set from Owens Lake carbonates (Benson and Bischoff, 1993) to constrain our modeled estimates of lake level history, rather than the Searles Lake isotopic record used by Phillips and others (1991). Because Searles received its input runoff from overspill from Owens Lake, the Searles Lake record is an incomplete record of climate change in the region. Finally, the model of Hostetler and Benson (1994), while excellent at predicting modern isotopic values in lakes, requires meteorological data that we presently have no way of constraining for Pleistocene paleoclimates. We therefore use a more simplified and general approach to determine to first order the relative importance of changes in runoff and evaporation on lake size. Though our modeling approach differs from the approaches employed by Orndorff and Craig (1994), Smith and Street-Perrott (1983), Benson and Hostetler (1994) and Phillips and others (1992), we draw on their works as sources of climatic data and for comparison of results.

MODEL

Theoretical Background

Lake Level Calculations. The water balance of a closed-basin lake can be described by the following equation (Mifflin and Wheat, 1979; Phillips and others, 1992; Benson and Hostetler, 1994):

$$\frac{dV}{dt} = \alpha P A_B - E A_L + G_{in} - G_{out} \pm V_{spill} \quad (1)$$

where P is the precipitation rate, A_B is the drainage basin area, α is a runoff coefficient, E is the lake evaporation rate, A_L is the lake surface area, G_{in} and G_{out} are the fluxes of groundwater into and out of the lake, and V_{spill} is the volume of water spilled into or

out of the lake basin. As we have no way of constraining either G_{in} or G_{out} for the Owens chain of lakes, we ignore these terms. Danskin (personal communication) has informed us that the quantity of water entering or leaving Owens Lake through the groundwater system is small (<10%) compared to the flux of surface water into the lake/aqueduct system. This is primarily due to the fact that the permeability of the sediments underlying Owens Lake is quite low. Looking at Mono Lake however, a possible analog for Owens Lake when it contained water, Rogers (1993) estimated that the groundwater contribution of inflow might be as high as 28% of the total. We ignore groundwater contributions to the water balance of Owens Lake through time, although we realize that such contributions may have been larger during wet periods of the Pleistocene, perhaps more akin to the situation in the Mono Basin. Phillips and others (1992, p. 59) make the same assumption of negligible groundwater contribution for Searles Lake based upon the observation that the basin is arid and that runoff into Searles Lake came primarily from overspill from China Lake rather than from precipitation over Searles' drainage basin. The same assumption can be applied to China, Panamint, and Manly Lakes. For a steady state lake which is neither gaining nor losing water by spillover, equation 1 simplifies to (Benson and Paillet, 1989),

$$\alpha P A_b = E A_L \quad (2)$$

in which the lake surface area, A_L , is easily predicted from known parameters. Any perturbation in runoff or evaporation rate will result in a new steady-state lake area.

We would like to know how long a particular lake system takes to respond to changes in climatic forcing (step-changes in runoff and evaporation rates), i.e., we seek the system-response time constant, t . Casting lake area in terms of lake volume, equation 1 becomes a non-linear non-homogeneous differential equation which can be solved to determine the response time of a lake to climatic perturbations. For the Owens chain of lakes, area and volume are approximately related by the expression

$$A_L = m * V^q \quad (3)$$

where m is a constant, and q is a power less than one (Fig. 3.4).

Area-volume relationships for the Owens Lake chain

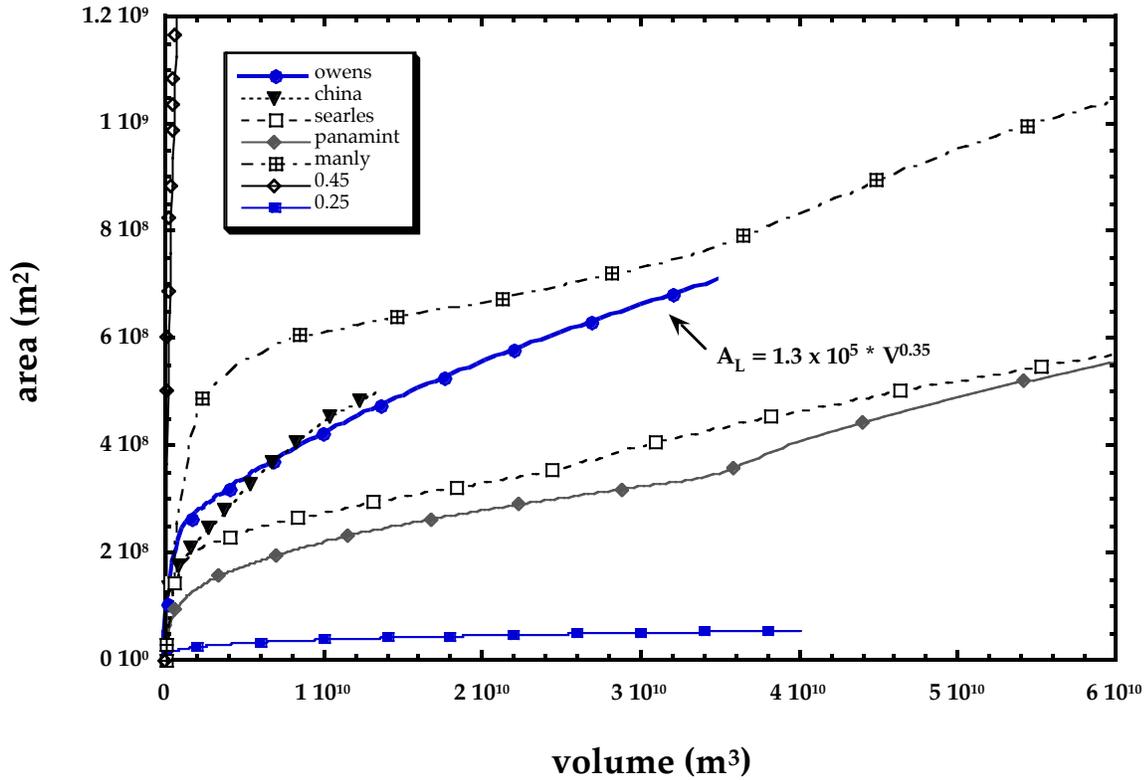


Fig. 3.4: Area-volume relationships for each lake in the Owens chain. The relationship generally can be characterized by the equation $A_L = m * V^q$ where A_L is the area of the lake, V is the volume of the lake, m is a constant, and q is a power < 1 . The dashed line parallel to the Owens Lake area-volume curve represents the best fit to the equation above for Owens Lake. Also shown are two hypothetical area-volume relationships for different values of the power q . Low q values give rise to very small increases in lake surface area for large changes in volume. High q values create lakes whose surface areas change rapidly with small increases in volume.

Substituting equation 3 into equation 1, ignoring both the groundwater fluxes into and out of the lake, and additions by overspill, defining $C = \alpha P A_B$ (corresponding to total basin runoff), and $a = m * E$, yields

$$\frac{dV}{dt} + a * V^q = C \quad (4)$$

At steady state, $a * V_{ss}^q = C$ or $V_{ss} = \left(\frac{C}{a}\right)^{\frac{1}{q}}$, where V_{ss} is the volume of the lake at steady state. As will be shown below in the section on steady-state climate experiments, the volume versus time behavior of an initially dry lake subjected to steady climate forcing appears to be approximated well by

$$V = V_{ss} * \left(1 - e^{-\frac{t}{\tau}}\right) \quad (5)$$

where t is a time constant dependent on amount of runoff, evaporation rate, and basin geometry. Differentiation of equation 5 and substitution into equation 4 yields:

$$\frac{1}{\tau} V_{ss} e^{-\frac{t}{\tau}} + C \left(1 - e^{-\frac{t}{\tau}}\right)^q = C \quad (6)$$

In order to solve this equation for t , we must define what we mean by steady state. We arbitrarily assume that we have reached steady state when the volume of the lake has reached 95% of its asymptotic steady state volume, V_{ss} . Rewriting equation 5 as

$$\frac{V}{V_{ss}} = \left(1 - e^{-\frac{t}{\tau}}\right) = 0.95 \quad (7)$$

and substituting this into equation 6 to solve for t gives:

$$\tau = \left[\frac{0.05}{1 - (0.95)^q} \right] * \frac{C^{\frac{1}{q}-1}}{(m * E)^{\frac{1}{q}}} \quad (8)$$

Once we have determined t , rearrangement of equation 7 reveals how long it takes an empty lake to achieve a steady-state volume in the face of steady climate forcing:

$$t = -\tau * \ln(0.05) \quad (9)$$

Equation 8 provides insight into the relative importance of climatic and geometric parameters in determining the response of a lake to climatic forcings. Because lake surface area is equivalent to lake volume raised to some power q , lake response time is extremely sensitive to basin geometry. In Figure 3.4, we show the five Owens chain- and two hypothetical-lake area-volume relationships. The smaller the q value, the smaller the increase in surface area per change in volume. The effect of these different geometries on response time is shown in Figure 3.5a. The lake with q value of 0.15 has an extremely long response time, reflecting the fact that its surface area-to-volume ratio is very low, and that it takes a long time for evaporation from the lake's surface to balance inflow. The converse is true for the lake with $q=0.55$. Owens Lake, with its q value of 0.35, has an intermediate response time. It should take approximately 100 years for the lake to reach 95% of its steady-state size.

In Figure 3.5b we explore the importance of changes in runoff and evaporation rate on response time for Owens Lake. For the same evaporation rate and basin geometry, small amounts of runoff are quickly balanced by evaporation while larger amounts of runoff require substantially longer times to reach steady state. The response time shows the opposite behavior for changes in evaporation rate when runoff rate and basin geometry are held constant; increases in evaporation rate cause decreases in t . Examination of Figures 3.5a and b shows that the hypsometry of the lake is of primary importance in determining the response time. Climatically reasonable changes in runoff and evaporation lead to changes in response time of order tens of years, while changes in hypsometry could cause changes in response time of orders of magnitude. However, hypsometric changes, such as might result from tectonic factors or changes in sedimentation in the basin, occur over time periods much longer than climatic change, and therefore do not present a problem when trying to model water balance.

Response time of a lake for different geometries

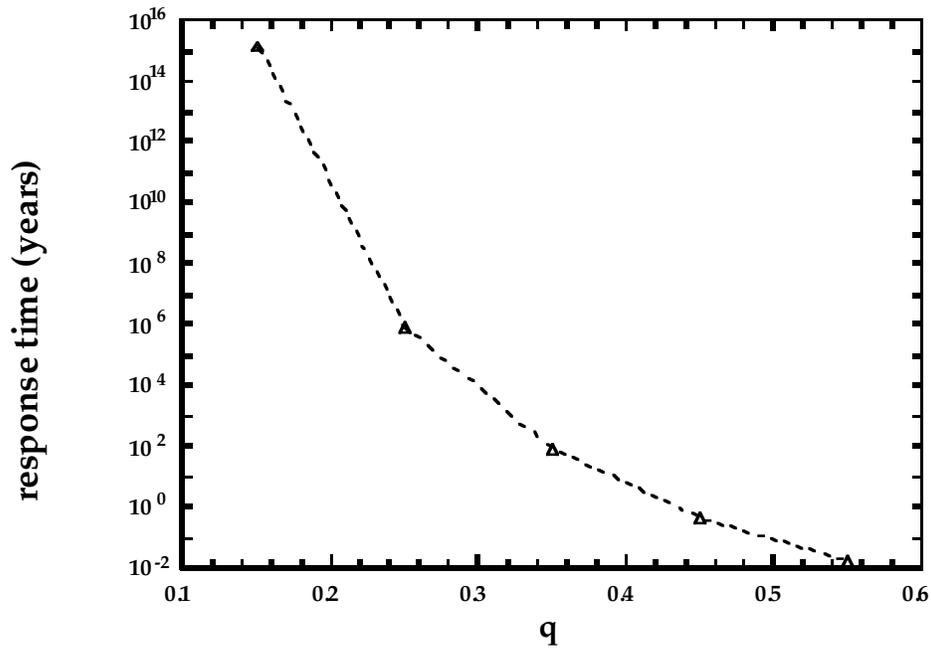


Fig. 3.5a: Relationship of lake response time to lake hypsometry. Evaporation and runoff are specified as modern values for Owens lake. Shown are response times for different q values. The smaller the q value, the longer it takes evaporation from the lake's surface to balance inflow.

Response time versus amount of runoff and evaporation

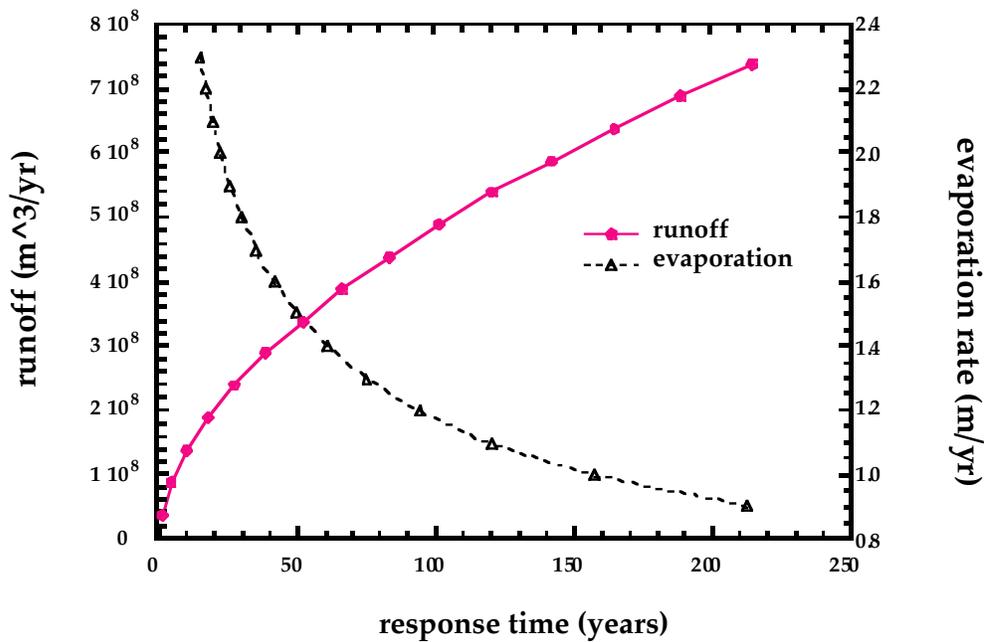


Fig. 3.5b: Relationship between lake response time and amount of runoff flowing into the lake and evaporation rate. For the runoff curve, basin hypsometry and evaporation rate remain constant. For the evaporation curve, basin hypsometry and runoff remain constant.

Isotopic Calculations. In addition to modeling volume changes with different climatic forcings, we can also model the oxygen isotopic composition of Owens Lake water through time for comparison to the isotopic values measured on carbonates in the Owens Lake core (Benson and Bischoff, 1993). We use the equations of Gonfiantini (1986) which describe the evolution of oxygen isotopic composition in a water sample subjected to evaporation:

$$\delta = \left(\delta_0 - \frac{a}{b} \right) f^b + \frac{a}{b} \quad (10)$$

where

$$a = \frac{(h * \delta_a) + \Delta \varepsilon + \frac{\alpha - 1}{\alpha}}{1 - h + \Delta \varepsilon} \quad (11)$$

$$b = \frac{h - \Delta \varepsilon + \frac{\alpha - 1}{\alpha}}{1 - h + \Delta \varepsilon} \quad (12)$$

and

$$\Delta \varepsilon = (1 - h) * \left(\frac{\rho_i}{\rho} - 1 \right) \quad (13)$$

h = relative humidity

ρ_i and ρ are resistance coefficients which describe the relative efficiencies of transport of the rare and common isotopic species

α = fractionation factor for water / vapor system,

$$\ln(\alpha) = \frac{1.137 * 10^3}{T^2} - \frac{0.4156}{T} - 2.0667 * 10^{-3} \quad (\text{Hostetler and Benson, 1994})$$

δ_a = the oxygen isotopic composition of the atmosphere over the lake

δ_0 = the initial oxygen isotopic composition of the lake water

f = the fraction of liquid remaining after evaporation

We employ this relationship to keep track of the oxygen isotopic composition of Owens Lake water through time, using the modern $\delta^{18}\text{O}$ value of Owens River water for the input precipitation, -16‰ (Hay and others, 1991), an atmospheric isotopic value of -21‰ (value used by Hostetler and Benson (1994) to model isotopic evolution of Pyramid Lake), and an empirically derived value of r_i/r of 1.0141 (Gonfiantini, 1986). We further assume that the relative humidity of the atmosphere over the lake is 35%. For comparison, the Owens Valley experiences relative humidities in excess of 40% during the winter and smaller than 30% in the summer. In order to facilitate comparison of the modeled $\delta^{18}\text{O}$ values with the measured carbonate $\delta^{18}\text{O}$ values of Benson and Bischoff (1993), we convert the water oxygen isotopic composition to an equivalent calcium carbonate isotopic composition using the following equations (Friedman and O'Neil, 1977):

$$1000 \ln(\alpha) = \frac{2.76 \times 10^6}{T^2} - 2.89 \quad (14)$$

$$\delta^{18}\text{O}_{\text{calcite}} = \left[\left(\delta^{18}\text{O}_{\text{water}} + 1000 \right) * \alpha \right] - 1000 \quad (15)$$

where α is the fractionation factor for the water-calcite system, and T is the calcification temperature, here taken to be the modern mean annual temperature near Owens Lake, 15°C .

Construction of the Model

The water balance model described in this paper is a finite difference computer model written in the Fortran computer language. In each model timestep, runoff is applied to Owens Lake and its volume is updated. We next employ a detailed lake basin hypsometry to calculate the new depth and surface area of the lake. Once surface area has been established, the evaporative flux of water from the lake is calculated by multiplying surface area by an assumed evaporation rate. When Owens Lake's volume

exceeds the maximum volume dictated by its spillway elevation, the excess volume is transferred to the next lake in the chain (China Lake). The overspilled volume represents the runoff into the next lake and the model then goes through the same process of updating that lake's volume, determining the new depth and surface area, and calculating an evaporative flux and overspilled flux from the lake. The volume, surface area, and depth calculations are conducted for each lake in sequence of its position within the lake chain before the next timestep begins. The model timestep is one year, and different experiments are run for timespans of several hundred years to 200 kyrs. The $\delta^{18}\text{O}$ of the model Owens Lake is also calculated for comparison with $\delta^{18}\text{O}$ values measured on carbonates from core OL-92 which was taken in 1992. The following sections describe the hypsometric and climatic data used in the exercise.

Lake hypsometries. We use U.S. Geological Survey 3 arc-second digital elevation data to construct altitude versus area hypsometries for each lake. We delineate drainage basins for each of the five lakes using Geographical Information System software, and calculate the area contained within each 1 m contour, yielding an area-elevation relationship for each lake. Next we determine the volume of the lake at each 1 m contour by averaging the surface area at that contour with the surface area of at the contour below, and multiplying that average by the contour interval. Figure 3.2 shows the hypsometries determined for each lake. Hypsometric data such as valley floor elevation, spillway elevation, maximum lake depth at spill elevation, and maximum volume attainable by each lake are found in Table 3.3.

Runoff. In the model, we lump changes in runoff coefficient, α , and precipitation rate, P , into changes in the product of these values, αP , because we cannot constrain changes in the individual parameters in the past. Using unpublished Los Angeles Department of Water and Power records of yearly discharge through the Owens River aqueduct system since 1932, we determine the mean annual runoff that would be available to Owens Lake were it not diverted to be $3.98 \times 10^8 \text{ m}^3/\text{yr}$.

TABLE 3.3: Hypsometric information for each modeled lake

	<u>Owens</u>	<u>China</u>	<u>Searles</u>	<u>Panamint</u>	<u>Manly</u>
lake base (m) ¹	1081	657	493	317	-86
spill elevation (m) ¹	1145	665	690	609	87 ³
max depth (m) ¹	64	8	197	292	173
max lake area (km ²) ²	610	195	650	780	1560
max lake volume (km ³) ²	25	1.2	76	117	153

¹from Smith and Street-Perrott, 1983

²determined from the 3 arc second DEMs and ArcInfo

³maximum shoreline elevation

Langbein and others (1949) reported a set of empirical curves relating runoff in the United States to mean annual air temperature and precipitation. Benson and Thompson (1987a) have suggested for various reasons that these curves are inaccurate for reconstructing climate in the Great Basin. In spite of their warnings, in order to make a first approximation, we make use of these curves to estimate what runoff during the last glacial maximum might have been, and then refine our estimates later when we compare our modeled isotopic values to the measured isotopic values of Benson and Bischoff (1993). Based upon geologic and climate-model evidence for a 5°C cooling (Porter, 1983; Dohrenwend, 1984; Spaulding and Graumlich, 1986; Stamm, 1991) and 70% increase in precipitation (Stamm, 1991), average runoff in the Owens Valley would have increased by ~10-fold over modern.

Evaporation. The product of the instantaneous surface area of each lake with the annual evaporation rate determines the evaporative flux of water out of the lake at each time step. We use the evaporation rates suggested by Smith and Street-Perrott (1983) for both modern conditions and for a 5°C cooler annual climate, with the exception of the modern Owens Lake evaporation value. We determine this value by solving the

steady state lake equation for evaporation rate given the historic lake size and mean annual runoff values. Lakes to the south and east of Owens Lake have progressively larger evaporation rates, reflecting both their decreasing elevations and rain shadow effects. Table 3.4 lists the measured modern and inferred glacial runoff and evaporation rates for each of the lakes in the model.

Assumptions

We make the following set of simplifying assumptions to facilitate the modeling. First, we ignore tectonic activity in each drainage basin, taking modern curves for our paleo-hypsometries. By making this assumption, we assume that the maximum volume of each lake has remained constant over the 200 kyrs represented by the model. Bierman and others (1991) report vertical displacement rates on range-bounding faults in Owens Valley of ~ 0.3 mm/yr. For a 200 kyr model run, this translates into 60 m of subsidence. However, as Owens Lake has received ~ 0.4 mm/yr of sediment over the last 800 ka, subsidence rates appear to be approximately matched by sedimentation rates (at least in Owens Valley), and we can assume that the lake hypsometry has not

TABLE 3.4: Hydrologic parameters for each lake for fixed runoff and evaporation case

	<u>Owens</u>	<u>China</u>	<u>Searles</u>	<u>Panamint</u>	<u>Manly</u>
average modern runoff (m^3/yr) ¹	3.98×10^8	0	0	0	0
average modern evaporation rate (m/yr) ²	1.34 [§]	1.41	1.65	1.65	1.97
inferred glacial runoff (m^3/yr)	3.98×10^9	0	0	0	0
inferred glacial evaporation rate (m/yr) ³	0.94	0.99	1.15	1.15	1.38

changed significantly. Similar sedimentation rates have been reported for Searles (0.2 mm/yr, Phillips and others, 1992), and Panamint (0.4 mm/yr, Jannik and others, 1991) Valleys, with a somewhat higher rate (~1 mm/yr) reported for Death Valley (Ku and others, 1994). Subsidence rates in these valleys are poorly constrained (Ellis, personal communication). However, Hodges and others (1989), citing previous work, state that between 0 and 2 km of vertical movement has occurred along the Hunter Mtn. fault zone in northern Panamint Valley. The subsidence of Panamint Valley, relative to Hunter Mtn. itself is thought to have occurred over the last 3-3.6 Ma, giving maximum subsidence rates on order 0.5-0.6 mm/yr for Panamint Valley, in general agreement with sedimentation rates in the valley. Hooke (1972) used tilted fan segments in Death Valley to estimate a vertical slip rate of 7 mm/yr on the Black Mountain fault. If this slip rate were indicative of subsidence of the entire valley, then Death Valley would have to be subsiding much more rapidly than it is filling with sediment. In this case, our assumption of a constant hypsometry over the last 200 kyrs would be invalid. However, as Death Valley is the last lake in the chain and does not overflow, the position of the valley floor relative to the spillway is unimportant in our modeling exercises.

We set runoff contributions from all drainage basins other than Owens Valley to zero. Spaulding and Graumlich (1986) analyzed pack-rat middens to determine the spatial and temporal distribution of temperature and precipitation over the last ~25 ka in the southwestern U.S. They inferred a decrease in mean annual precipitation during the last glacial maximum for regions in the lee of the Sierra Nevada, while localities to the south experienced enhanced precipitation. These data would tend to confirm the calculations of Smith (1976), which show that high lake-level stands in the China, Searles, Panamint and Manly Lake basins must have been derived by overspill from Owens Lake rather than by increased precipitation on each lake's individual drainage basin.

We ignore groundwater fluxes into or out of each lake for reasons mentioned above. We further neglect any dependence of evaporation rate on lake salinity, assuming in all of our calculations that fresh water is evaporating. Salts tend to decrease the thermodynamic activity of water, thereby slowing the evaporation rate (Gonfiantini, 1986), but this effect is very small and can be neglected unless waters are exceptionally saline (Gonfiantini, 1986). There is no evidence in core OL-92 that Owens

Lake ever reached salinities high enough for evaporation to have been inhibited. Since Owens Lake is presently a dry playa, we have no information regarding the lake's previous thermal structure. We therefore assume that the lake is an unstratified, homogeneous water body with no gradients in temperature or isotopic composition with depth. Finally, we assume simple Rayleigh distillation of lake water with a relative humidity of 35% in the overlying atmosphere. Sensitivity analyses below will address some of our assumptions.

EXPERIMENTS

In order to develop our understanding of the dynamics of the 5-lake system, we report a series of experiments employing both steady and variable climatic forcing, and displaying each lake's depth through time. In addition, for Owens Lake, we continuously monitor the oxygen isotopic composition of lake water for comparison to the data set of Benson and Bischoff (1993). In the first set of experiments (steady-state calculations), we use modern climate data to assess the response time of lakes to dramatic increases and decreases in precipitation. In the second set of experiments (sinusoidally varying runoff and evaporation), we allow runoff and evaporation to vary sinusoidally with time between modern and hypothetical glacial values. In the final set of experiments (runoff and evaporation driven by the marine $\delta^{18}\text{O}$ record), we use the $\delta^{18}\text{O}$ record of deep sea sediments (Imbrie and others, 1984) to drive runoff and evaporation between modern and assumed glacial values.

Steady-state Calculations

We conduct a series of steady-state experiments to determine the response time of each lake to variations in runoff and evaporation. Figure 3.6 shows the history of Owens Lake volume, given an initially empty basin subjected to modern climatic conditions. This curve appears to follow the form

$$V = V_{ss} * (1 - e^{-t/\tau}) \quad (16)$$

where V is the volume of the lake at any particular timestep, V_{ss} is the steady state volume, t is time, and τ is a time constant dependent on basin geometry, amount of runoff, and evaporation rate.

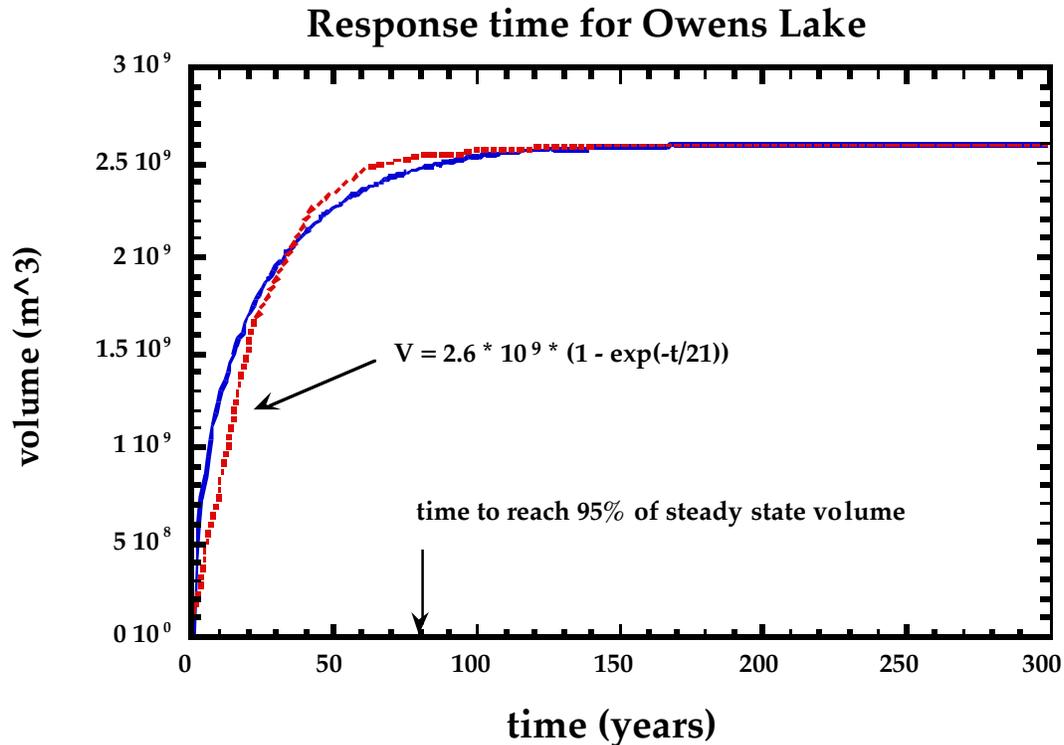


Fig. 3.6: Volume of Owens Lake with time, given an initially empty lake basin subjected to modern climatic conditions: $3.98 \times 10^8 \text{ m}^3/\text{yr}$ of runoff, evaporation rate of $1.34 \text{ m}/\text{yr}$. Note that the lake achieves a steady-state volume in about a century. Therefore, climate changes occurring with timescales of centuries or longer should be reflected in the sediments of the Owens Lake core.

The lake achieves a steady-state volume in approximately one century. Holding evaporation rate constant, we explore the relationship between Owens Lake response time and amount of runoff (Fig. 3.7). For a fixed evaporation rate, response time increases as runoff increases. This agrees with the derivation of response time in equations 8 and 9 above. In fact, the only difference between Figure 3.7 and Figure 3.5b is that to create Figure 3.7 we use the actual hypsometry of Owens Lake rather than an idealized hypsometry of the form $A_L = m * V^q$ used in Figure 3.5b. At runoff values $>210\%$ of modern, Owens Lake exceeds the confines of its basin and spills into China Lake. Below $\sim 40\%$ of modern runoff, Owens Lake becomes ephemeral, with evaporation removing all runoff that has entered the lake during each annual timestep.

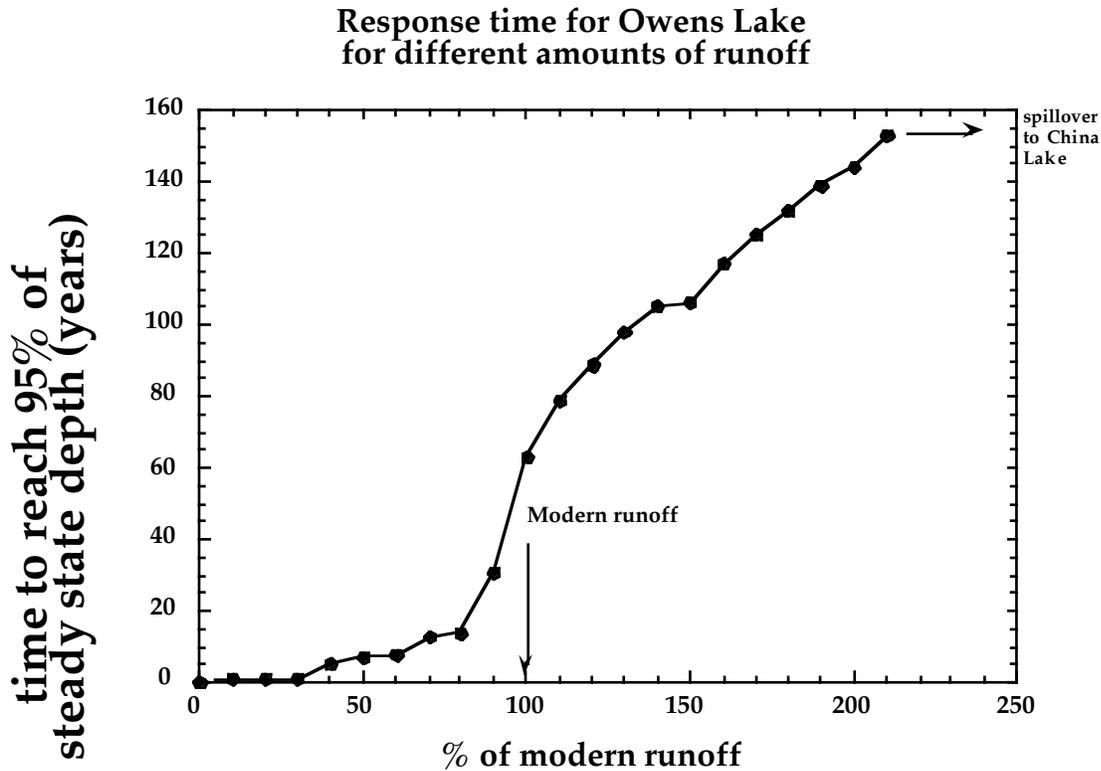


Fig. 3.7: Relationship between lake response time (here defined as the time necessary to reach 95% of the steady state volume) and amount of runoff for Owens Lake. Evaporation rate remains constant at 1.34 m/yr while runoff varies between 0% and 210% of modern, with modern runoff = $3.98 \times 10^8 \text{ m}^3/\text{yr}$. The higher the amount of runoff, the longer the lake takes to respond (see equations 8 and 9). This results from the fact that the steady-state lake size for a higher runoff lake is larger than for a lower runoff lake. Above 210% of modern runoff, Owens Lake exceeds its confines and flows into China Lake.

Similar plots for the other lakes in the chain show similar behavior: all lakes have response times of approximately one century. In Figure 3.8, we show the response of all lakes in the chain to steady state climatic forcing over the Owens Lake drainage basin (no local precipitation falls on any of the other lake basins), with runoff equal to 10x modern and evaporation equal to modern values. The spilling over and successive

filling of each lake in the chain is readily apparent. It takes the entire lake chain about 280 years to respond to this climate forcing.

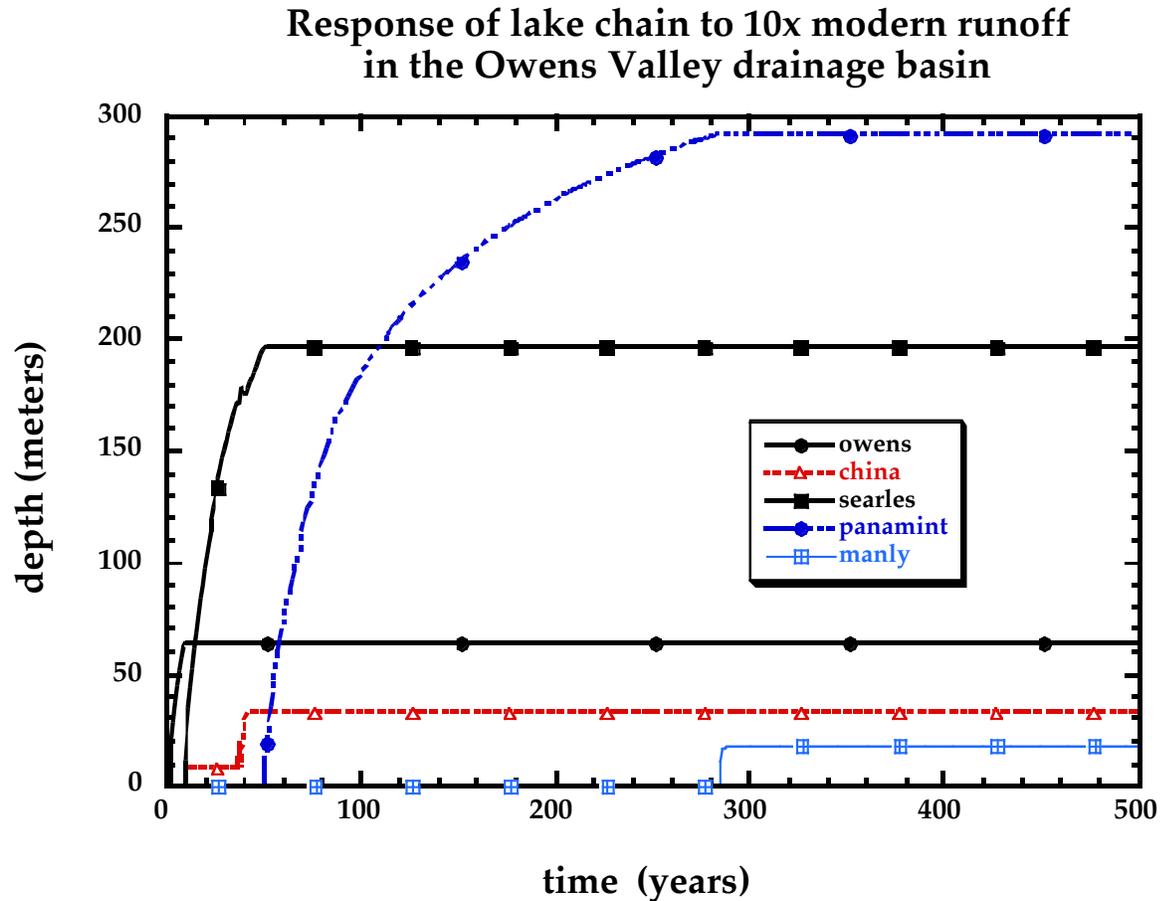


Fig. 3.8: Response of the lake chain system to an input of water to Owens Lake. All lakes are initially empty and receive water solely by overspill from one basin to the next. Likewise, all lakes are subjected to modern evaporation rates (see Table 2). Owens Lake receives 10 x modern runoff and rapidly fills to its spill point. China then instantaneously fills to its spill point and introduces water to Searles Lake. At 20 years, Searles has reached the China Lake spill point and the two lakes coalesce and grow to a combined spillover level. The maximum depth acquired by China Lake, then, is 30 m during periods when it is attached to Searles Lake, and 8 m when the two lakes are separated. The jump from 8 to 30 m is readily apparent at 20 years into the model run. The jaggedness of the lake depth versus time curves at 20 years for both the China and Searles Lake records is an artifact of the way these lakes are "welded" together in the model; it is not reflective of a short term climatic phenomenon. The combined Searles-China lake spills into Panamint Valley at 25 years. Because of the enormity of the Panamint Valley basin, it takes another 250 years for Panamint Lake to reach its spill point and flow into Death Valley. A small (<20 m) lake forms in Death Valley and the whole chain achieves steady state at ~280 years.

Figures 3.7 and 3.8 indicate that for long term changes in climate, such as 20, 40 and 100 kyr variations induced by orbital cycles, the lake chain is perpetually in steady state. However, for short timescale phenomena, such as annual to decadal fluctuations

in runoff and evaporation, steady state will not have been achieved by the time the forcing changes. In Figure 3.9, we depict the desiccation of Owens Lake that would occur if the lake were at its maximum high-stand volume and then ceased to receive further runoff. The modern evaporation rate dries the lake completely within 50 years. Stine (1994) has found evidence for two extreme droughts in California which lasted more than 200 years prior to 1112 A.D. and for more than 140 years prior to 1350 A.D. Inasmuch as the response time for Owens Lake is on the order of one century, we might expect to find evidence of these drought events in the Owens Lake core. At present, these events have not been identified.

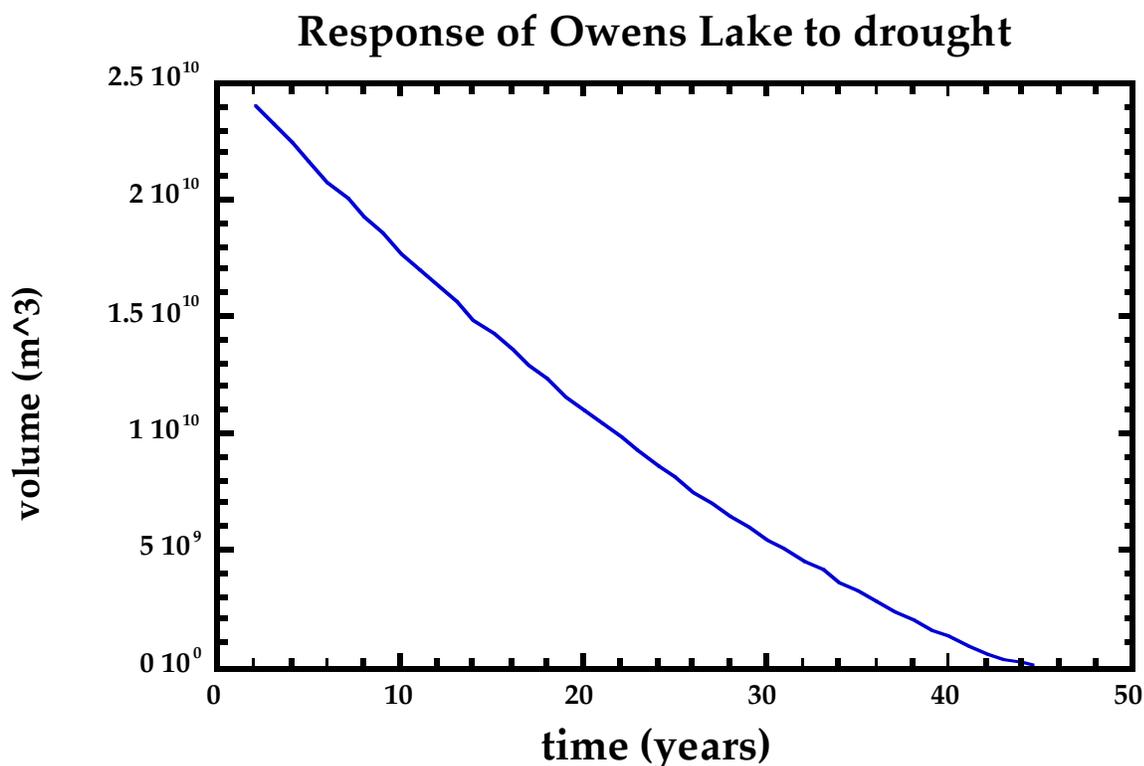


Fig. 3.9: Response of Owens Lake to drought conditions. Lake begins model run at its highstand volume. Modern evaporation rate removes water from lake. No runoff comes into the lake.

Finally, the oxygen isotopic composition of Owens Lake water for different amounts of runoff is depicted in Figure 3.10. Higher runoffs lead to more depleted lake water

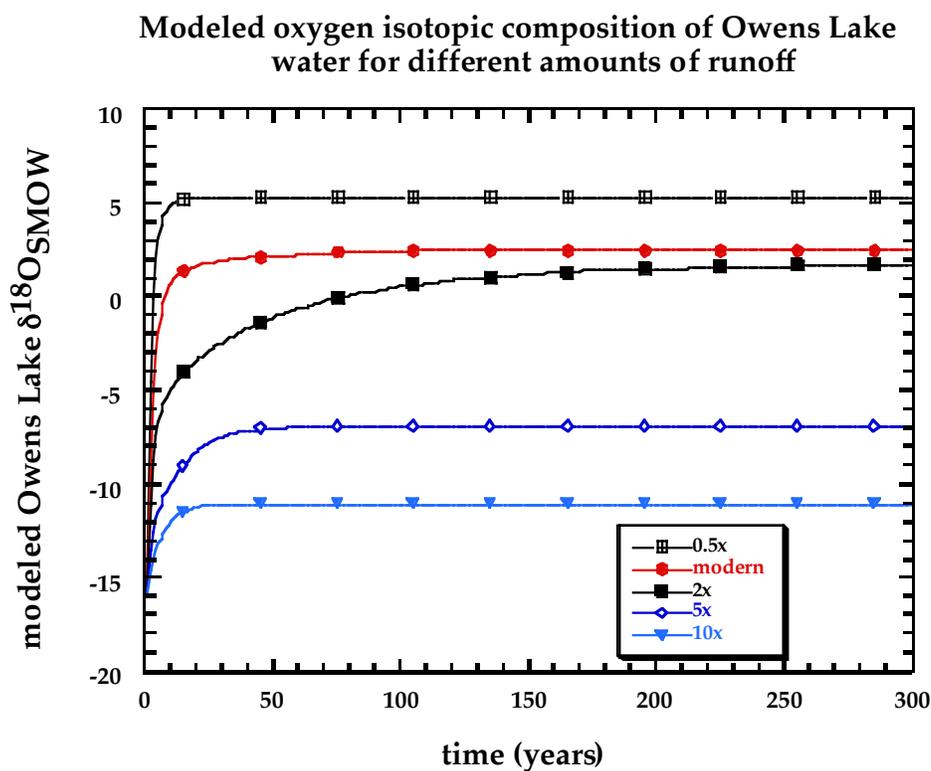


Fig. 3.10: Oxygen isotopic composition of Owens Lake water for different amounts of runoff (modern = $3.98 \times 10^8 \text{ m}^3/\text{yr}$, 0.5x modern, 2x modern, 5x modern, and 10x modern), reported in SMOW units. In all cases, evaporation rate = 1.34 m/yr. Isotopic composition of the input Owens River water is -16‰ . The lake quickly reaches isotopic steady state, with larger amounts of runoff requiring more time to equilibrate. The higher the amount of runoff, the more depleted the lake water at steady state. The steady-state isotopic value for the 2x modern case eventually reaches $\sim 1.8\text{‰}$ which is very similar to the value for modern runoff of $\sim 2.3\text{‰}$.

values (i.e., values more similar to the initial Owens River value of -16‰), while lower runoff leads to more enriched lake water values. The degree of enrichment of lake water reflects the relative importance of runoff and evaporation rates on lake volume. For high runoff (5x and 10x modern runoff), Owens Lake is overflowing to the China Lake basin, and the residence time of water in the lake is very short. As a result, water does not remain in Owens Lake long enough for significant evaporative enrichment to occur. In contrast, for lower runoff (0.5x, 1x, and 2x modern), Owens Lake remains below its spill level, and evaporation causes a 15-20‰ enrichment over the input water isotopic value. In all cases, however, lake isotopic composition reaches a steady state value in roughly 50 years or less. Thus, climatic variations of the sort described by Stine (1994), with durations of 140 to >200 years, should also be recorded in the isotopic composition of carbonate minerals precipitated in equilibrium with Owens Lake water.

Sinusoidally Varying Runoff and Evaporation

To illustrate how the Owens Lake system might respond to fluctuations in climatic parameters, we next conduct a series of experiments in which runoff and evaporation are driven sinusoidally with different periods, 100 years and 20 kyrs. In the first experiment, αP in the Owens Lake drainage basin varies between modern values and 10 times modern with a period of 100 years. We choose this period in order to confirm the response time determinations from the steady state climate experiments. If the response time of the lake chain is in fact ~ 280 years for a 10x modern runoff forcing, then none of the lakes in this experiment should achieve their steady state sizes. For all lakes in this experiment, evaporation is driven between modern and 70 percent of modern values. The minimum evaporation values are taken from Smith and Street-Perrott (1983), who estimated the decrease in evaporation rate associated with a 5°C cooling in the area. Figure 3.11a shows the runoff and evaporation forcing functions at Owens Lake over the 1000 year long model run; we have specified that the maximum runoff conditions prevail during periods also dominated by minimum evaporation rates, i.e., that the two forcings are 180° out of phase. Figure 3.11b shows the lake depth

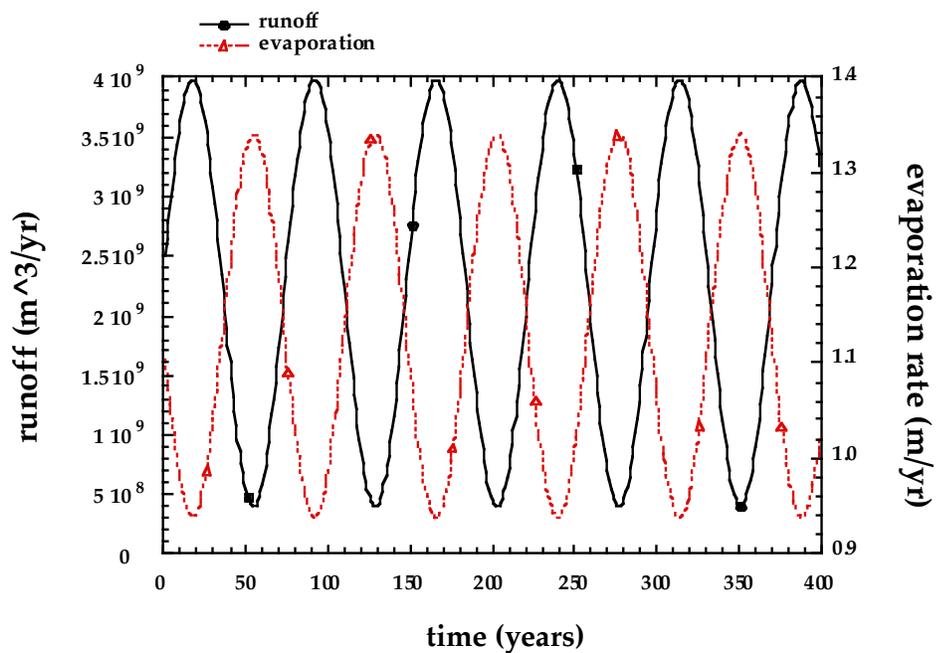


Fig. 3.11a: Climatic forcing functions for oscillating climate with 100 year period. All lakes are initially empty and lakes down-chain from Owens receive water solely by overspill from one basin to the next. Runoff varies between modern and 10x modern values. Evaporation varies between modern and 0.7x modern values and is shown only for Owens Lake.

Lake level history for sinusoidally driven climate.

Period = 100 years

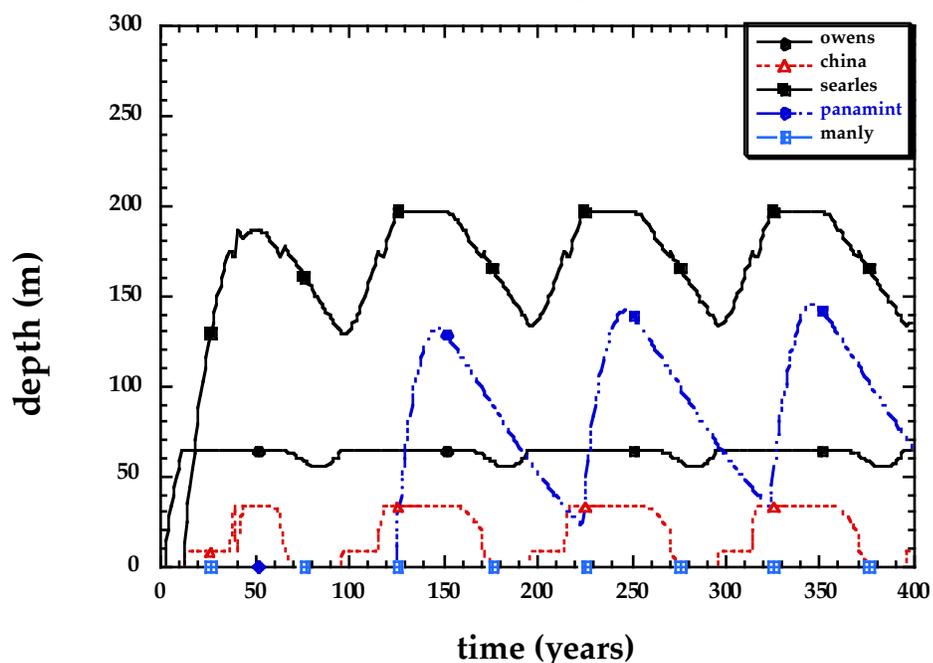


Fig. 3.11b: Lake depth versus time for oscillating climate with 100 year period. Note that Owens Lake never reaches its modern depth of 14 m, even though the runoff and evaporation rates go through modern values every 50 years. Likewise, large lakes persist in the China, Searles, and Panamint Lake basins even though these lakes should be completely dry during periods marked by modern climatic conditions. The persistence of large lakes in these basins indicates that the lake system has not achieved steady state.

histories corresponding to this climatic forcing. Note that while the climatic forcing goes through modern values every 100 years (Fig. 3.11a), Owens Lake never reaches its modern depth of 14 m. Instead, its surface altitude hovers around the spillpoint elevation, declining only by ~ 10 m during minimum runoff and maximum evaporation conditions. Likewise, in spite of the observation that modern climatic conditions lead to no lakes in any basin but Owens Valley, large modeled lakes persist in Searles and Panamint Valleys given these parameters. These results suggest that the system of lakes has not reached steady state values for lake depth, volume, and surface area. Remembering the response time calculations detailed above, in which we determined that response to climatic perturbation occurs within about a century, the behavior

shown in Figure 3.11b is explicable as a transient response, i.e., the time scale of the forcing is too short for the lakes to attain their steady state values.

Figure 3.11c depicts residence time for water in the lake (calculated by dividing instantaneous lake volume by instantaneous inflow rate). Residence time is longest at 75 years and every 100 years thereafter, corresponding to intermediate values of runoff and evaporation (Fig. 3.11a). In Figure 3.11d we show the evolution of the oxygen isotopic value of Owens Lake through time (here reported in ‰ units relative to SMOW). The most enriched values occur immediately after the longest residence time periods, perhaps indicating again that the lake chain has not reached steady-state.

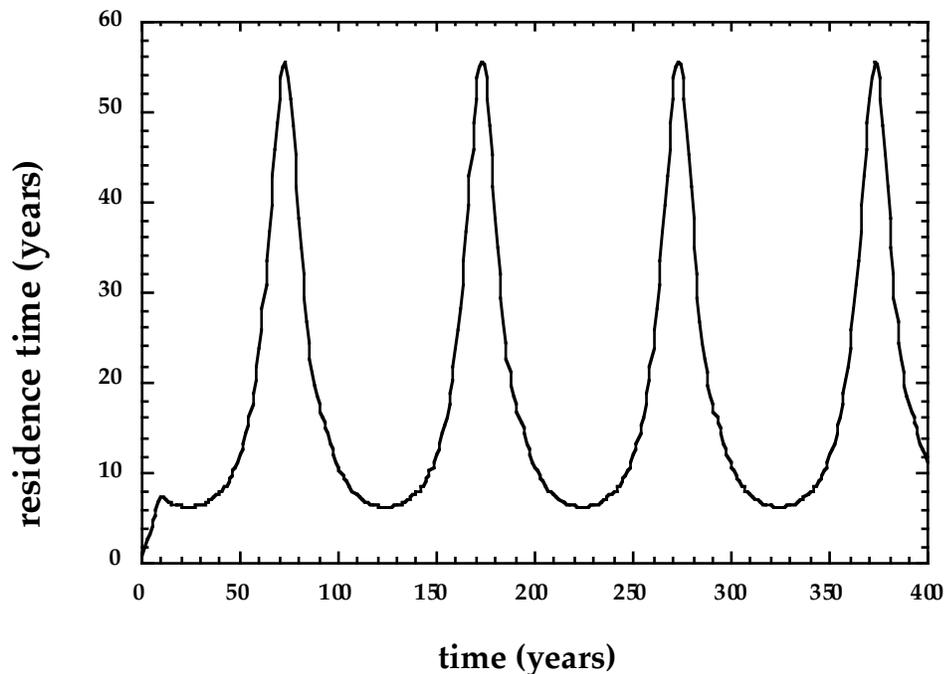


Fig. 3.11c: Residence time for water in Owens Lake for oscillating climate with 100 year period (calculated by dividing lake volume by inflow rate). Comparison of residence time to isotopic composition of lake water (Fig. 3.8c) shows that Owens Lake water becomes most enriched shortly after residence time reaches maximum values.

In the second experiment (Fig. 3.12), runoff and evaporation vary between the same extremes as in the first experiment, but the period of fluctuation is changed to 20 kyrs, and the model run time is adjusted to 40 kyrs. Owens Lake rapidly reaches its

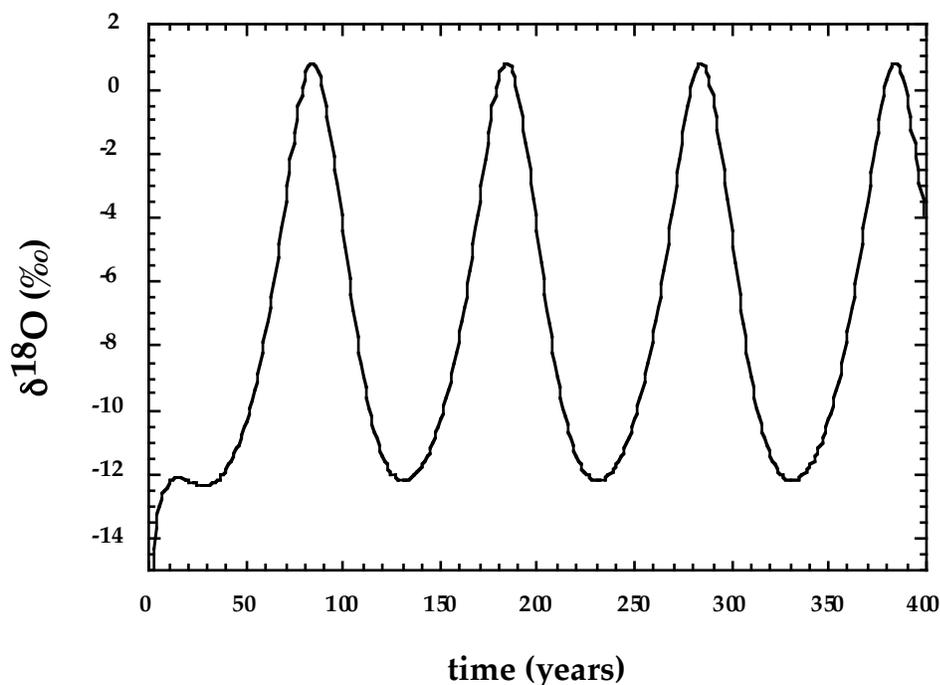


Fig. 3.11d: Oxygen isotopic value of Owens Lake water through time for oscillating climate with 100 year period (here reported SMOW units). Lake water becomes most enriched at ~80 years and every 100 years thereafter. Comparison of the isotopic values to the lake level history shows that the isotopic enrichment occurs when Owens Lake is below its spill point.

spill point, retreating from this value only at 13 and 33 ka, when modeled runoff declines and evaporation increases. China, Searles, and Panamint lakes also fill to overflowing, resulting in a 90 m deep Lake Manly in Death Valley every 20 kyrs. Owens Lake achieves a steady state modern lake level of 14 m at 15 and 35 ka when runoff and evaporation reflect modern values. Comparison of the lake level graph (Fig. 3.12b) to a plot of residence time for water in Owens Lake (Fig. 3.12c) shows that residence time is influenced by two competing processes. Between 0 and 10 ka, and then again between 20 and 30 ka, residence time for Owens Lake water is very short due to high runoff pushing water rapidly through the system. From 10-12 ka, and 30-32

ka, runoff declines while evaporation rises, causing increasing residence times. At 15 and 35 ka, residence time has again become very short. At these times, however, the calculated residence time is dominated by the small volume of Owens Lake and the relatively large evaporation rate. The inflection points at 12, 17, 32 and 37 years, then, represent changeover from residence times dominated by the runoff signal to residence times dominated by evaporation. Figure 3.12d shows the oxygen isotopic composition of Owens Lake water reflecting this residence time behavior. Isotopic composition is nearest that of the input Owens River water (-16‰) when residence times are short and dominated by runoff (0-10 and 20-30 ka). $\delta^{18}\text{O}$ increases as runoff declines and evaporation increases (10-12 and 30-32 ka) and becomes most enriched at 15 and 35 ka, when residence time is dominated by evaporation.

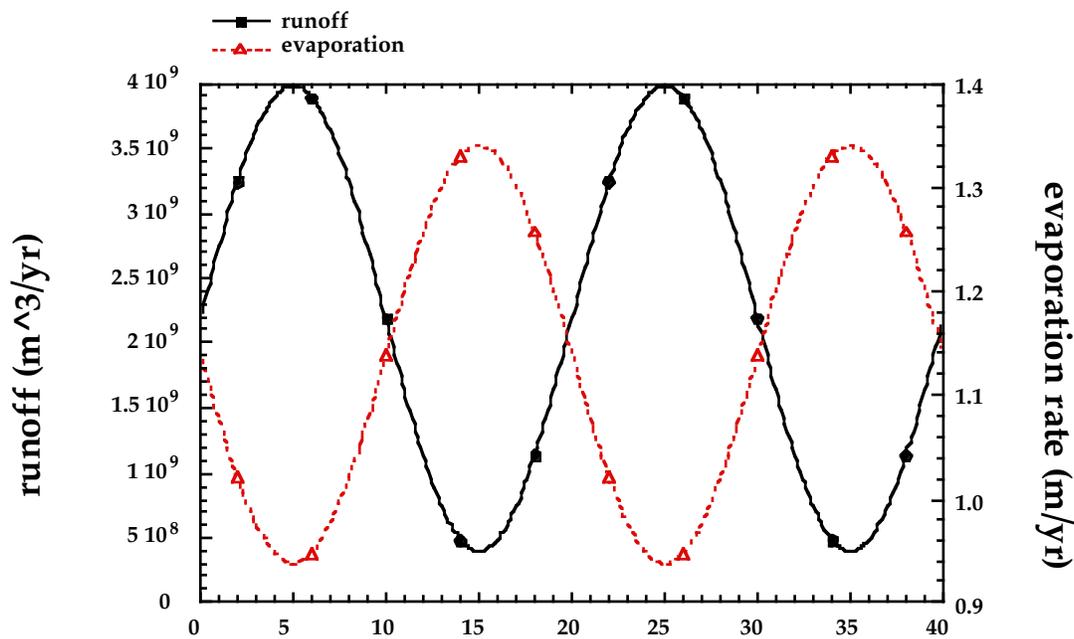


Fig. 3.12a: Climatic forcing functions for oscillating climate with a 20 kyr period. All lakes are initially empty and lakes down-chain from Owens receive water solely by overspill from one basin to the next. Runoff varies between modern and 10x modern values. Evaporation varies between modern and 0.7x modern values and is shown only for Owens Lake.

Lake level history for sinusoidally driven climate.
Period = 20 kyrs

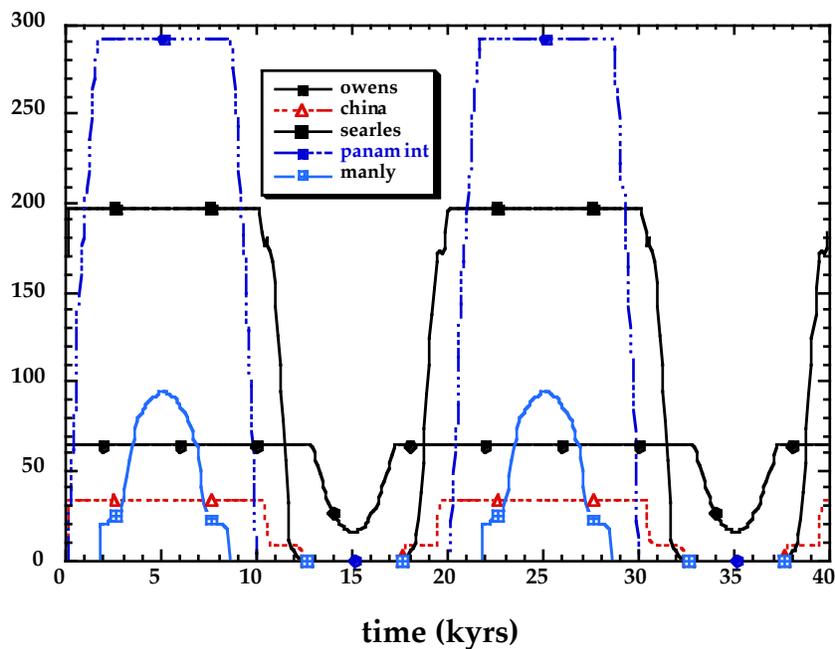


Fig. 3.12b: Lake depth versus time for oscillating climate with 20 kyr period. Runoff and evaporation go through modern values at 15 and 35 ka. At these times, Owens Lake reaches its historic lake level (14 m), and all other lakes in the chain dry out completely. Conversely, at 5 and 25 ka, runoff and evaporation are at assumed maximum pluvial values. All lakes upchain from Manly are at their spill points, and Manly Lake achieves a depth of ~95 m. The fact that the lakes reach their historic sizes with historic climate parameters indicates that the chain of lakes is at steady state, unlike the situation portrayed in Figure 3.8.

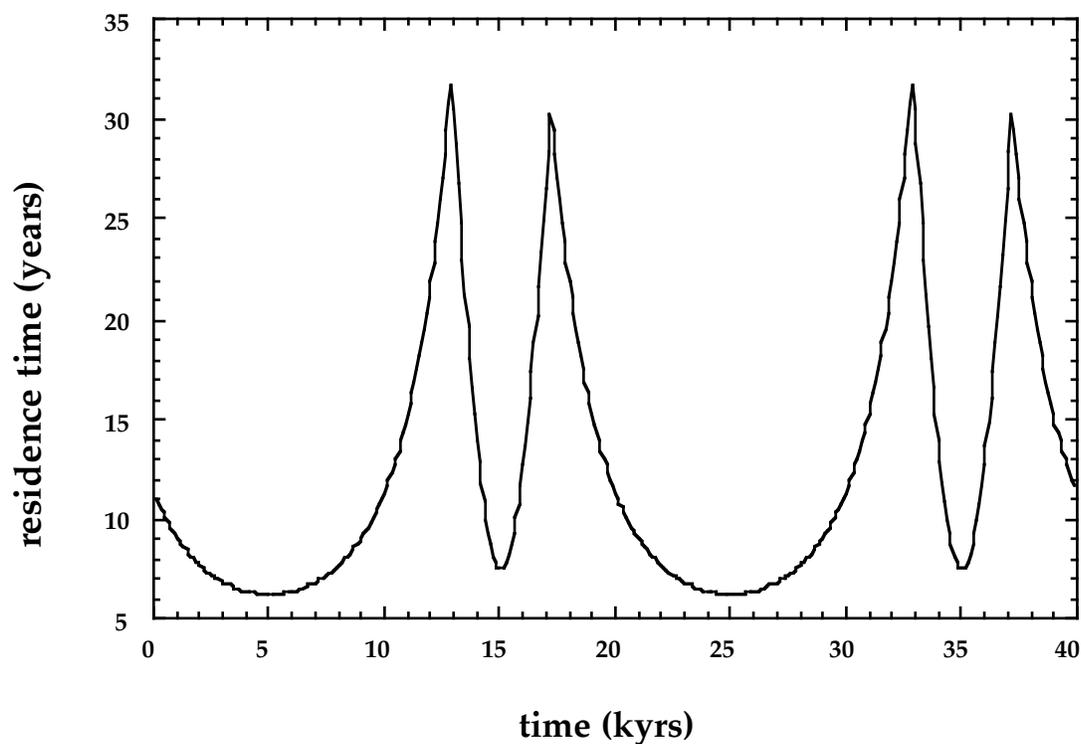


Fig. 3.12c: Residence time for water in Owens Lake for oscillating climate with 20 kyr period (calculated by dividing lake volume by inflow rate). Residence time is influenced by two competing processes. Between 0 and 10 ka, and then again between 20 and 30 ka, residence time for Owens Lake water is very short due to the large amounts of runoff pushing through the system. From 10-12 ka, and 30-32 ka, runoff declines while evaporation rises causing increasing residence times. At 15 and 35 ka, residence time has again become very short. At these times, however, residence time reflects the small volume of Owens Lake and the relatively large evaporation rate. The inflection points at 12, 17, 32 and 37 years, then, represent changeover from residence times dominated by the runoff signal to residence times dominated by evaporation.

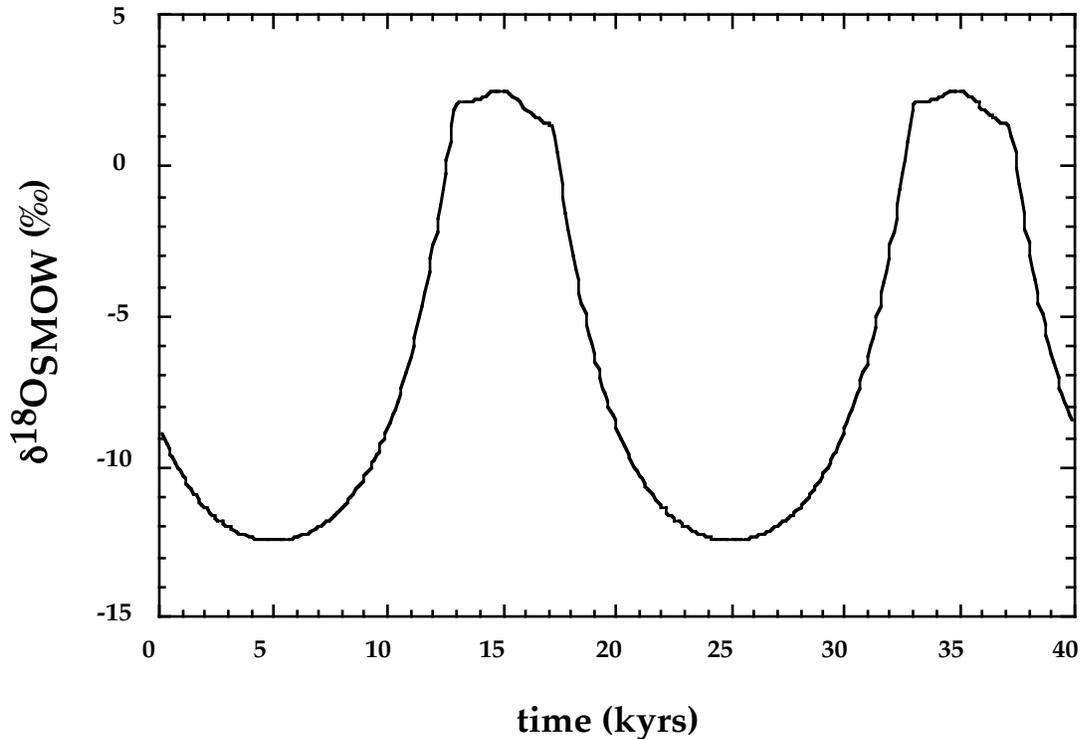


Fig. 3.12d: Oxygen isotopic value of Owens Lake water through time for oscillating climate with 20 kyr period (here reported SMOW units). Lake water becomes most enriched at 15 and 35 kyrs when climate goes through modern values of runoff and evaporation (i.e. minimum runoff and maximum evaporation). Comparison of isotopic composition of lake water to residence time (Fig. 3.9c) shows that Owens Lake water becomes most enriched during the short residence time periods dominated by evaporation.

Runoff and Evaporation Driven by the Marine Oxygen Isotopic Record

We now examine a more complicated climatic forcing function. Antevs (1955), Kutzbach (1987), and Spaulding and Graumlich (1986) suggested that the growth of the Laurentide ice sheet might have deflected the jet stream from its present mean location at $\sim 45^\circ\text{N}$ latitude to a more southerly location. Southerly latitudes with an arid modern climate would then have received more precipitation during glacial periods as storm tracks moved southward. Lao and Benson (1988) supported this hypothesis by showing that Lake Lahonton reached highstands either directly during or slightly after ice sheets reached their maximum sizes, and that moderate size lakes formed when ice sheets were 80% of their maximum sizes. Likewise, Oviatt and others (1990) called on a migrating jet stream to produce the Stansbury oscillation in lake level in the Lake

Bonneville basin. We have incorporated the effects of a migrating jet stream in the following way: we modulate between the assumed maximum paleo-runoff and modern runoff using a temporal pattern derived from the deep sea oxygen isotope record (Imbrie and others, 1984). Because $\delta^{18}\text{O}$ reflects global ice volume (Bradley, 1985, p. 180), in fact, primarily the Laurentide ice sheet, and the position of the jet stream is influenced by that volume (Antevs, 1955; Kutzbach, 1987), we suggest that the $\delta^{18}\text{O}$ record may be used to scale a proxy for jet stream position, and, therefore, precipitation falling over our study area. We also use the marine $\delta^{18}\text{O}$ record to drive evaporation rates between modern and assumed glacial values. Phillips and others (1992, p. 101) used a similar approach, allowing temperature, and therefore evaporation rate, to scale between assumed maximum and minimum values with the marine $\delta^{18}\text{O}$ record.

For both evaporation and runoff, we normalize the marine isotopic curve so that its values fall between 0 and 1, with maximum runoff and evaporation given a value of one, and minimum values of these parameters given a value of zero. We also specify that maximum evaporation rates occur when runoff values are at a minimum, and vice versa. As in the sinusoidally driven climate experiments above, maximum runoff is specified to occur during periods of minimum evaporation, maximum runoff is set to 10x modern runoff, and minimum evaporation taken to be 0.7 x modern. Figure 3.13a shows the runoff and evaporation forcing histories at Owens Lake through time, and Figure 3.13b the depth histories of the 5 lakes. Comparison of this lake level history to those reported by Phillips and others (1991) and Jannik and others (1991) does not show very good agreement. Overflow from Searles Lake to Panamint Lake occurs over much too long a period, and Lake Manly receives overspilled water at many times in this model run. It appears that the maximum glacial estimate of runoff of 10 times the modern value may be too large, or that the estimated evaporation rate is too small. Figure 3.13c shows the modeled oxygen isotopic composition (converted to calcium carbonate equivalent) along with the measured OL-92 oxygen isotopic composition of carbonate from Benson and Bischoff (1993). The modeled $\delta^{18}\text{O}$ values are generally much more depleted than the measured isotopic values, again reflecting that too much runoff is coursing through the system during glacial periods. Finally, the magnitude of

the predicted isotopic enrichment at ~195 ka and at ~125 ka is not borne out in the measured data.

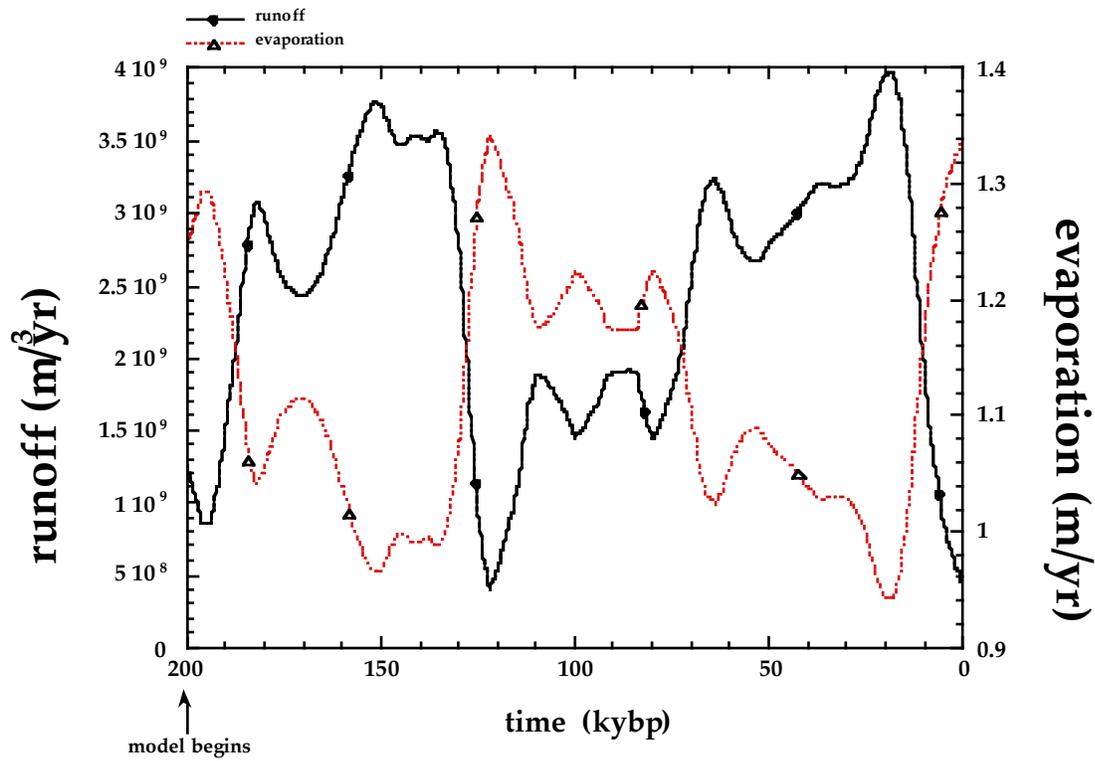


Fig. 3.13a: Climatic forcing functions for potentially more realistic climate variations. All lakes are initially empty and lakes down-chain from Owens receive water solely by overspill from one basin to the next. Runoff varies between modern and 10x modern values. Evaporation varies between modern and 0.7x modern values and is shown only for Owens Lake. Both runoff and evaporation are driven with the marine oxygen isotope curve of Imbrie and others (1984). Maximum precipitation and minimum evaporation occur at the last glacial maximum (18 ka).

Modeled lake level history

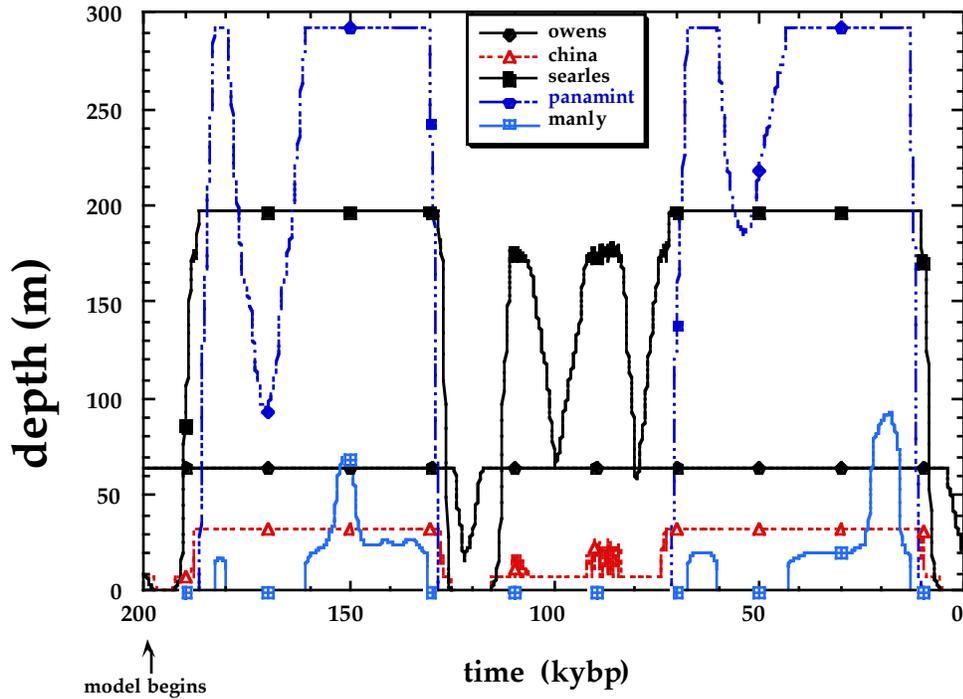


Fig. 3.13b: Lake level history derived from the climatic forcing functions shown in Figure 3.10a. Owens Lake is nearly continuously overflowing except at ~125 ka and at present. Likewise, China and Searles Lakes are overflowing for about 75% of the model run time. Large lakes grow in Panamint Valley, and overflow into Death Valley occurs at roughly 180 ka, from 130-160 ka, from 60-68 ka, and again from 12-43 ka. The data of Bischoff and others (in press) and Jannik and others (1993) do not support the frequent overflows into Death Valley, or the duration of overflow from Searles Lake to Panamint Lake.

Modeled and measured carbonate $\delta^{18}\text{O}$ values

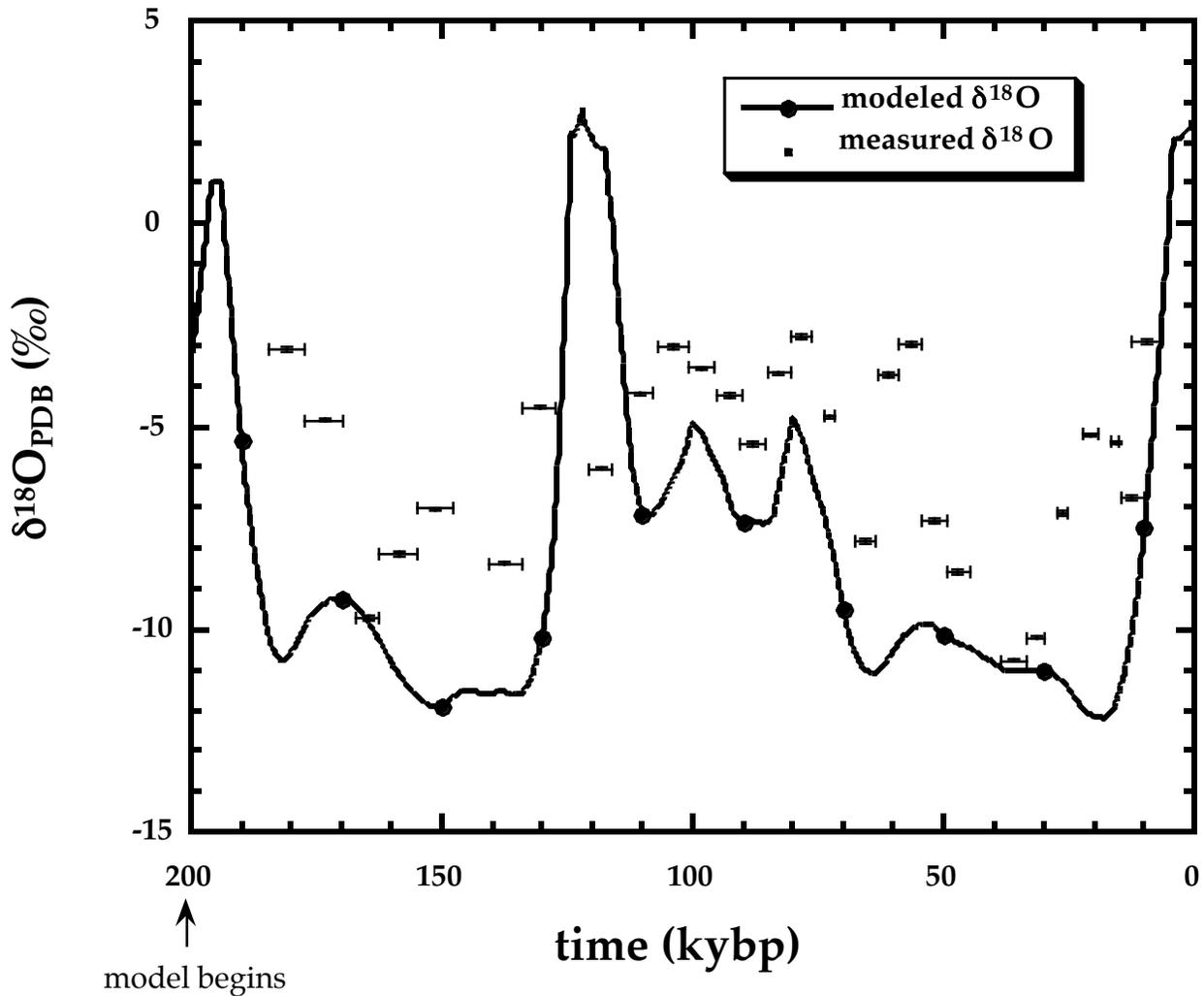


Fig. 3.13c: Comparison of measured carbonate $\delta^{18}\text{O}$ values (Benson and Bischoff, 1993) of Owens Lake carbonates to modeled isotopic composition of carbonates forming in a modeled Owens Lake (shown in PDB units) given the climatic forcing functions found in Figure 3.10a. Input Owens River $\delta^{18}\text{O}$ is specified at -16‰ , carbonates are assumed to be precipitating at 15 °C , relative humidity is set at 35% , and an atmospheric $\delta^{18}\text{O}$ value of -21‰ is assumed. In general, the modeled isotopic values tend to be more depleted than the measured values. Only at 195 and 125 ka do the modeled isotopic values become more enriched than the measured values.

SENSITIVITY ANALYSES

Because our most robust test of the modeled lake level history is the measured $\delta^{18}\text{O}$ of Owens Lake carbonates (Benson and Bischoff, 1993), we must determine the

sensitivity of the modeled $\delta^{18}\text{O}$ to variations in the parameters that control $\delta^{18}\text{O}$. These include runoff, evaporation rate, input Owens River $\delta^{18}\text{O}$, relative humidity, atmospheric $\delta^{18}\text{O}$, and calcification temperature. Variations in relative humidity exercise no control on isotopic composition because the fraction of water evaporating from Owens Lake during any particular timestep is very small relative to the total volume of the lake (the fraction of remaining liquid is ~ 0.95). Were Owens Lake ever to desiccate completely, however, the relative humidity would play an important role in the isotopic evolution of the brine and resulting carbonate precipitates as shown by Gonfiantini (1986).

While relative humidity plays no role in setting the isotopic composition of Owens Lake water, variations in the oxygen isotopic composition of the atmosphere over the lake could play a small role (Fig. 3.14). A range of atmospheric $\delta^{18}\text{O}$ values from -11‰ to -25‰ results in a lake isotopic range of $\sim 2\text{--}3\text{‰}$, with highly depleted atmospheric values giving rise to most depleted lake values. In our model, we have used an atmospheric isotopic value of -21‰ , which Hostetler and Benson (1994) reported for the air over Pyramid Lake, ~ 350 km north of Owens Lake. This value would presumably have been lower during glacial periods as colder air should have carried more isotopically depleted water vapor. In any case, the effect of atmospheric $\delta^{18}\text{O}$ on lake carbonate $\delta^{18}\text{O}$ is very small.

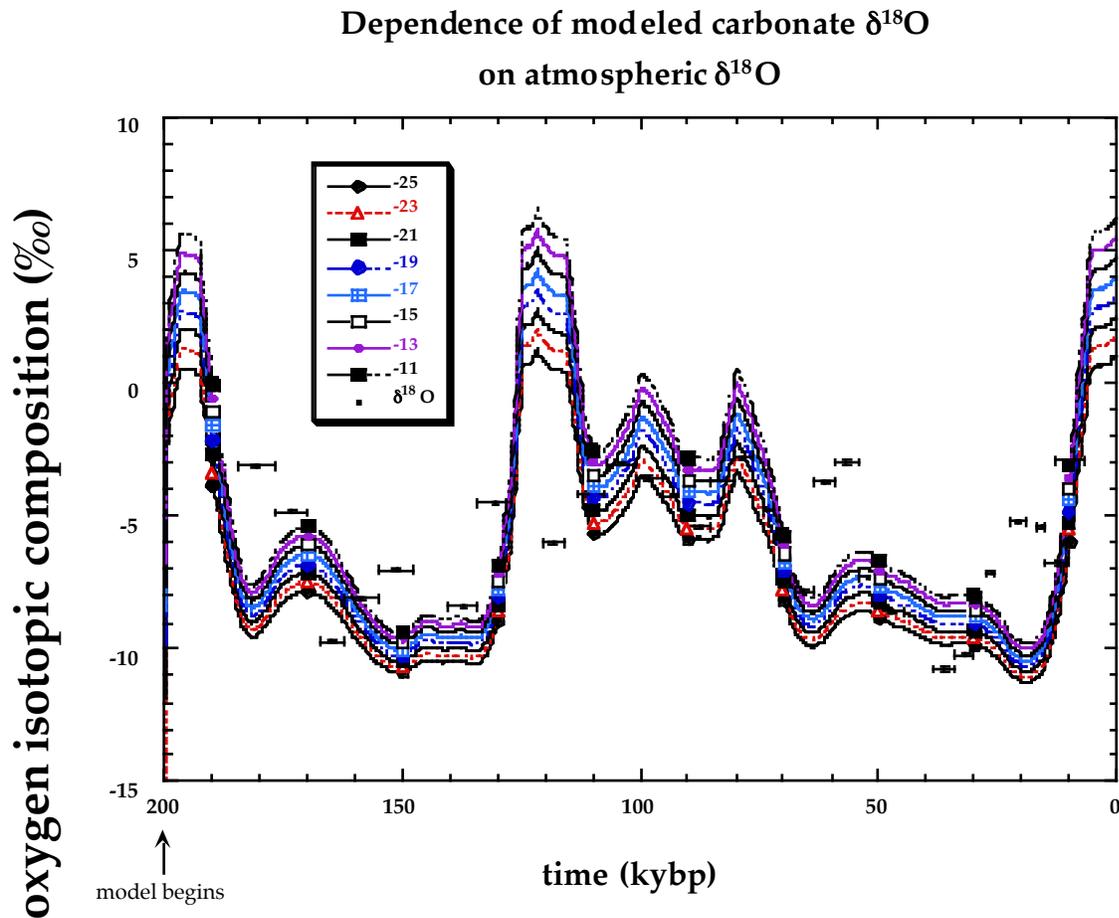


Fig. 3.14: Dependence of modeled oxygen isotopic value of carbonates on atmospheric $\delta^{18}\text{O}$ value. Eight different isotopic values are shown, ranging from -11‰ to -25‰ . As we have no knowledge of the atmospheric $\delta^{18}\text{O}$ over Owens Lake, we give a range that allows both a more enriched value than at Pyramid Lake due to Owens Lake's more southerly location and lower altitude, and a more depleted value possibly reflecting glacial periods. Runoff is modulated between modern and $7\times$ modern values with the marine $\delta^{18}\text{O}$ record. Evaporation varies between modern and $0.7\times$ modern again with the marine $\delta^{18}\text{O}$ record. Input Owens River water has a $\delta^{18}\text{O}$ value of -16‰ . Relative humidity is specified as 35%. Finally, carbonates are assumed to be forming at 15°C . Measured carbonate $\delta^{18}\text{O}$ values (Benson and Bischoff, 1993) are also shown with the age ranges for each sample plotted as error bars parallel to the time axis, and with precision plotted as error bars parallel to the oxygen isotopic composition axis.

Calcification temperature exerts a similarly small effect on isotopic composition (Fig. 3.15). We have assumed that calcium carbonate is precipitating inorganically within Owens Lake at temperatures equivalent to the mean annual temperature at Owens Lake (15°C for the modern case, and inferred to be 10°C for the glacial periods). This is undoubtedly an oversimplification inasmuch as carbonates are sometimes precipitated in saline lakes in "whiting" events (Benson, 1994) when temperatures in the

lake's epilimnion exceed 20°C (i.e., during the summer). Because Owens Lake is currently dry, we have no way of assessing the importance of this process. Nor do we

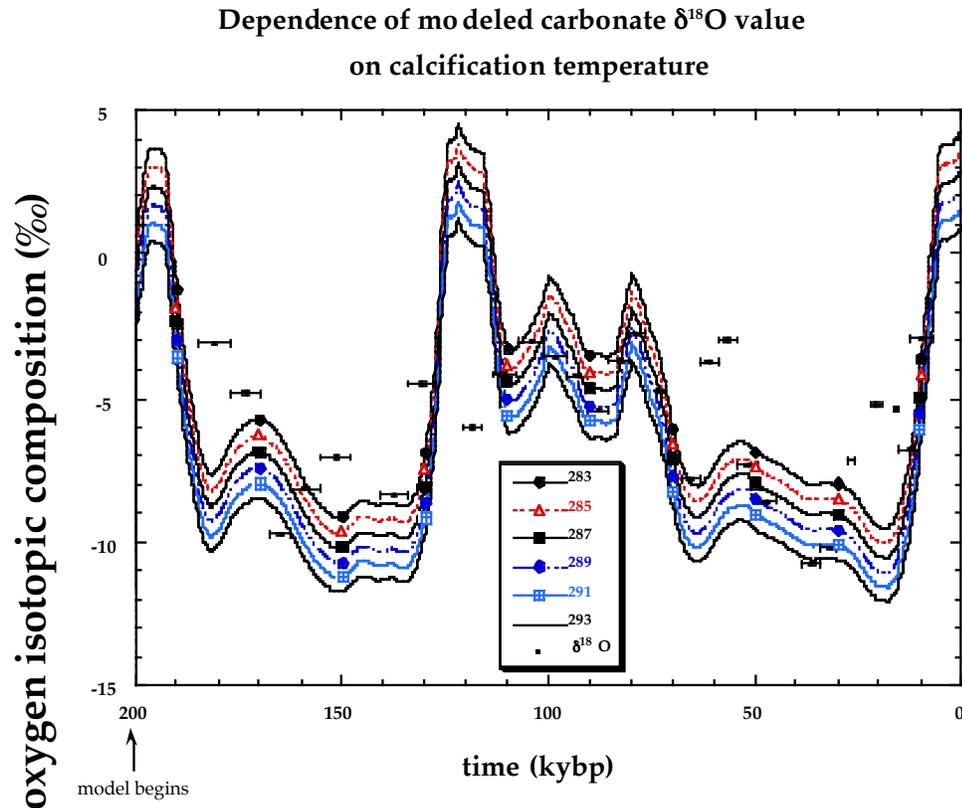


Fig. 3.15: Dependence of modeled oxygen isotopic value of carbonates on calcification temperature. Six different temperatures are shown, ranging from 283 K (10°C) to 293 K (20°C). Runoff is modulated between modern and 7x modern values with the marine $\delta^{18}\text{O}$ record. Evaporation varies between modern and 0.7x modern again with the marine $\delta^{18}\text{O}$ record. Input Owens River water has a $\delta^{18}\text{O}$ value of -16‰. The atmosphere over the lake is assumed to have a $\delta^{18}\text{O}$ value of -21‰. Finally, relative humidity is specified to be 35%. Measured carbonate $\delta^{18}\text{O}$ values (Benson and Bischoff, 1993) are also shown with the age ranges for each sample plotted as error bars parallel to the time axis, and with precision plotted as error bars parallel to the oxygen isotopic composition axis.

have any information about what the lake's thermal structure would have been for different lake levels. We can, however, model the isotopic evolution of lake carbonates for temperatures between 10 and 20°C (shown as 283 to 293 K on the plot). This 10°C range in calcification temperature results in an $\sim 2\text{‰}$ range in isotopic composition, with the highest temperatures giving rise to the most depleted, least fractionated, carbonate isotopic values. Glacial period evaporation rate also exerts a small effect on $\delta^{18}\text{O}$ (Fig.

3.16) since variations in runoff are large enough to swamp the much smaller expected changes in evaporation rate. In general, however, the lower the evaporation rate during maximum glacial periods, the less enriched the carbonates precipitating in Owens Lake. The amplitude of isotopic variation in the measured carbonate samples is $\sim 10\text{‰}$. Relative to this amplitude, the $\sim 2\text{‰}$ variations associated with the calcification temperature, the isotopic composition of the atmosphere, and the glacial period evaporation rate are small, though not negligible. In model runs to follow we will explore the results of allowing all parameters to vary within reasonable ranges.

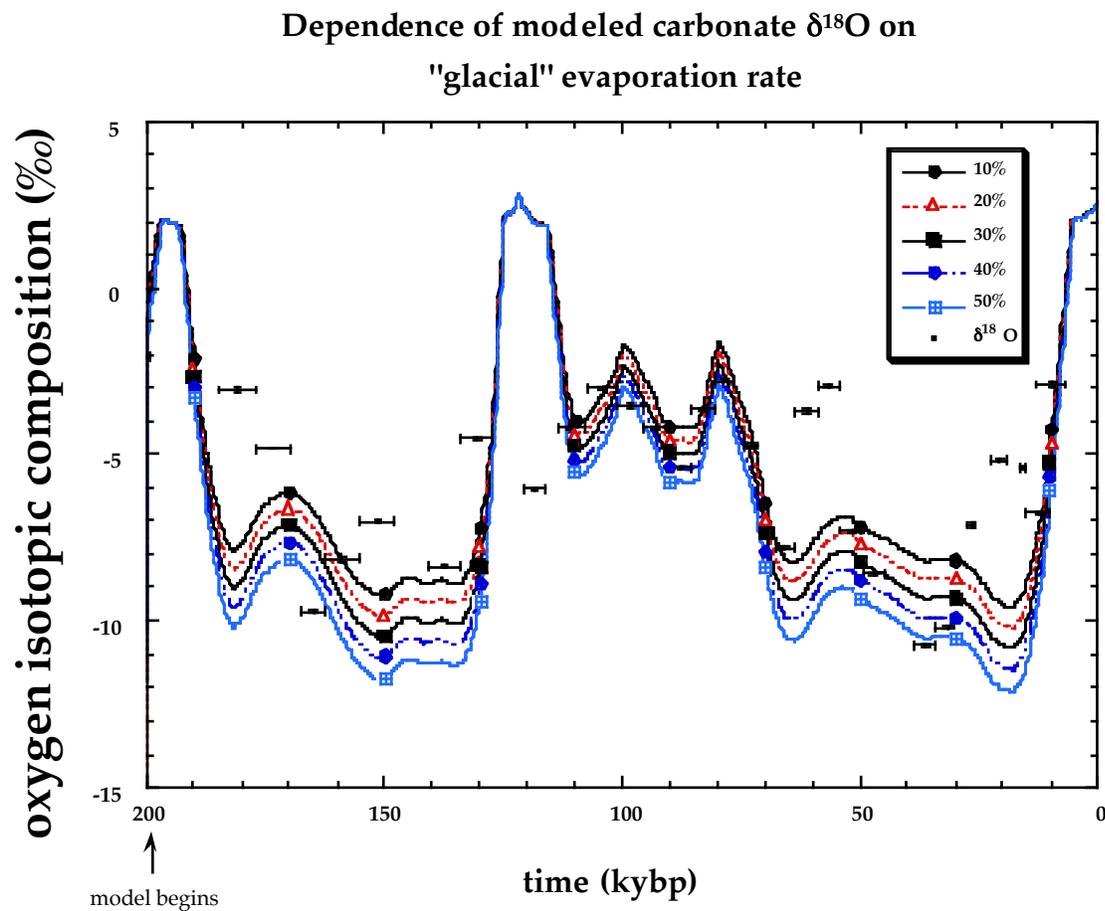


Fig. 3.16: Dependence of modeled oxygen isotopic value of carbonates on evaporation rate. Five different glacial evaporation rates are shown, ranging from 10% lower than modern to 50% lower than modern values. Runoff is modulated between modern and 7x modern values with the marine $\delta^{18}\text{O}$ record. Evaporation is again modulated with the marine $\delta^{18}\text{O}$ record. Input Owens River water has a $\delta^{18}\text{O}$ value of -16‰ . The atmosphere over the lake is assumed to have a $\delta^{18}\text{O}$ value of -21‰ and a relative humidity of 35%. Carbonates are assumed to be forming at 15°C . Measured carbonate $\delta^{18}\text{O}$

values (Benson and Bischoff, 1993) are also shown with the age ranges for each sample plotted as error bars parallel to the time axis, and with precision plotted as error bars parallel to the oxygen isotopic composition axis.

The two most important controllers of modeled oxygen isotopic composition are the amount of runoff (Fig. 3.17) and the isotopic composition of the runoff (Fig. 3.18). We first address changes in runoff quantity. For maximum glacial runoffs only 2x greater than modern (this value does not allow overspill from Owens Lake to China Lake), structure in the isotopic curve between 130 and 80 ka is lost completely due to the fact that the steady-state isotopic composition of a lake receiving 2x modern runoff is nearly identical to that of a lake receiving modern runoff (Fig. 3.10). Furthermore, isotopic values are much more enriched than those observed in the Owens Lake carbonates. In contrast, maximum glacial runoff of 10x modern results in modeled lake isotopic values that are more depleted than those measured, and an isotopic curve with a much higher amplitude or range of values than that found in the data. The runoff value that appears to match the measured data best has a 7x modern precipitation during glacial maxima.

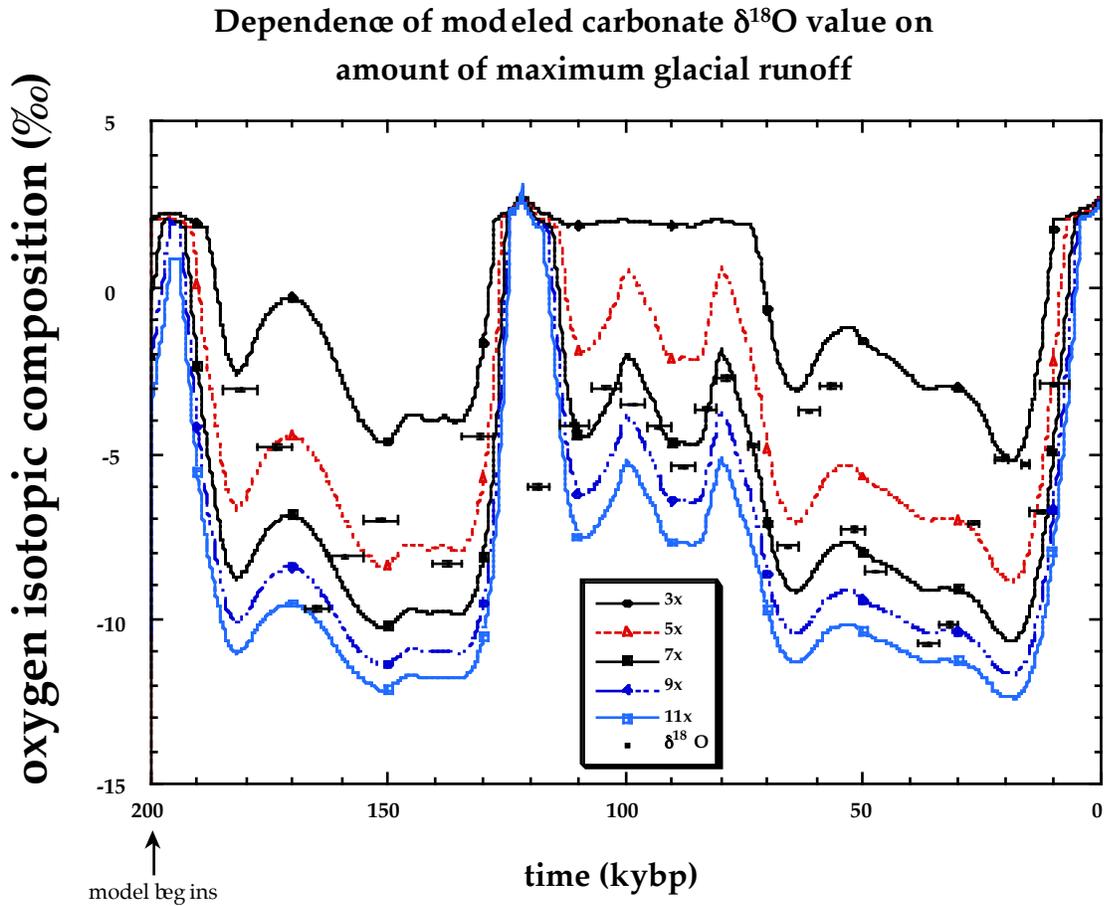


Fig. 3.17: Dependence of modeled oxygen isotopic value of carbonates on amount of runoff. Five different maximum glacial runoff values are shown, ranging from 2x modern runoff to 10x modern runoff. Runoff is again modulated between modern and assumed glacial values with the marine $\delta^{18}\text{O}$ record. Evaporation varies between modern and 0.7x modern also with the marine $\delta^{18}\text{O}$ record. Input Owens River water has a $\delta^{18}\text{O}$ value of -16‰ . The atmosphere over the lake is assumed to have a $\delta^{18}\text{O}$ value of -21‰ and a relative humidity of 35%. Finally, carbonates are assumed to be forming at 15°C . Measured carbonate $\delta^{18}\text{O}$ values (Benson and Bischoff, 1993) are also shown with the age ranges for each sample plotted as error bars parallel to the time axis, and with precision plotted as error bars parallel to the oxygen isotopic composition axis.

Dependence of modeled carbonate $\delta^{18}\text{O}$ on
input Owens River isotopic composition

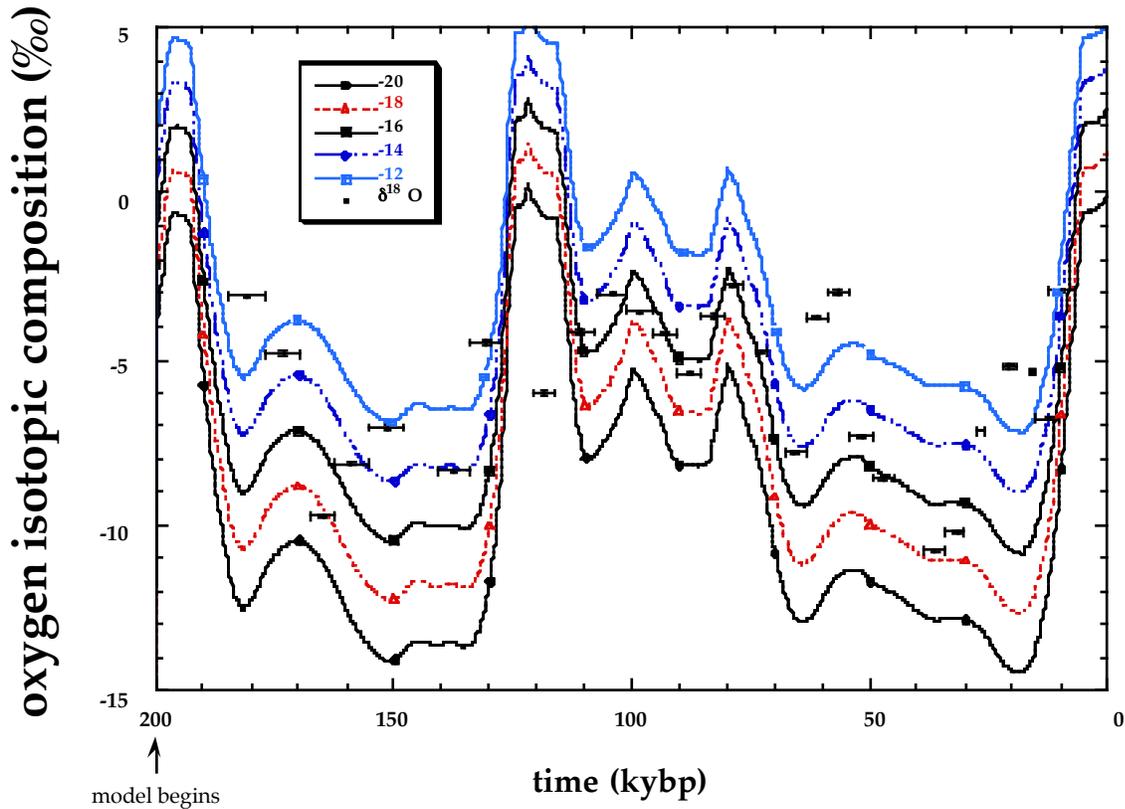


Fig. 3.18: Dependence of modeled oxygen isotopic value of carbonates on input Owens River isotopic value. Five different input isotopic values are shown, ranging from -12‰ to -20‰ . These values have been chosen to reflect both more depleted, glacial values, and more enriched, "super-interglacial" values, or values that might be expected if the Sierra Nevada were at a lower elevation resulting in less orographically-induced fractionation. Runoff is modulated between modern and 7x modern values with the marine $\delta^{18}\text{O}$ record. Evaporation varies between modern and 0.7x modern again with the marine $\delta^{18}\text{O}$ record. Input Owens River water has a $\delta^{18}\text{O}$ value of -16‰ . The atmosphere over the lake is assumed to have a $\delta^{18}\text{O}$ value of -21‰ and a relative humidity of 35%. Carbonates are assumed to be forming at 15°C . Measured carbonate $\delta^{18}\text{O}$ values (Benson and Bischoff, 1993) are also shown with the age ranges for each sample plotted as error bars parallel to the time axis, and with precision plotted as error bars parallel to the oxygen isotopic composition axis.

Benson (1994) showed that the isotopic composition of precipitation falling in the Lake Tahoe drainage basin varied seasonally by 20‰ or more, with a mean value of -14.7‰. Precipitation falling in winter was typically more depleted than that falling in summer, which he ascribed to the lower winter air temperatures. Given the relationship between the isotopic value of precipitation and air temperature, we might expect that the $\geq 5^{\circ}\text{C}$ decline in air temperature associated with glaciation would result in a depletion of isotopic values of precipitation falling in the Owens Lake drainage basin during glacials relative to modern mean annual values. We must therefore explore the sensitivity of modeled carbonate $\delta^{18}\text{O}$ values to changes in input water $\delta^{18}\text{O}$ values. In Figure 3.18 we show the modeled isotopic composition for 5 different input runoff $\delta^{18}\text{O}$ values, ranging from -12‰ to -20‰. It is readily apparent that the value chosen for the input runoff is crucial in determining the absolute isotopic value of the modeled carbonates, although the amplitude of isotopic variation between glacial and interglacial periods would remain the same for any particular input isotopic value. If, however, input runoff isotopic value were allowed to change with time, amplitude of isotopic variation would also change.

A close examination of Figures 3.14 through 3.18 shows that changes in some parameters simply shift the pattern of modeled carbonate isotopic composition toward more depleted or more enriched values, while other parameters change the amplitude of the modeled isotopic pattern. Those parameters that do not influence the residence time of water in the lake, such as the isotopic compositions of the atmosphere and input Owens River water, and the calcification temperature, have no effect on the amplitude of the isotopic pattern. Rather, they modulate the isotopic "starting point" of the model run, and thus also control the absolute isotopic values attainable in the model run. Runoff rate and evaporation rate *do* influence the residence time of water in the lake, and therefore, strongly influence the amplitude of the isotopic curve. As we showed in the sinusoidally fluctuating climate scenarios, residence time of water in a lake is influenced by two competing effects. Residence time is short both when the volume of runoff is large so that water spends a short time in the lake prior to being spilled into the next lake in the chain, and when the lake is small such that evaporation removes a large fraction of the lake water volume in each timestep. Between these two short

residence time periods lies a period of much longer residence time, which reflects the changeover from residence time dominated by runoff to residence time dominated by evaporation. In Figure 3.19 we show the residence time for the five different runoff values leading to Figure 3.17. It is clear that glacial period runoff values of 10x modern show much shorter residence times (~5 years) than runoff values of 2x modern (~10 years). Likewise, at present and at ~125 ka, residence time is again very short, but this time it reflects the small volume of Owens Lake and the comparatively large evaporative flux out of the lake at each timestep. Comparison of Figures 3.19 and 3.17 shows that lake waters become most isotopically enriched during periods of long residence time or short residence time dominated by evaporation.

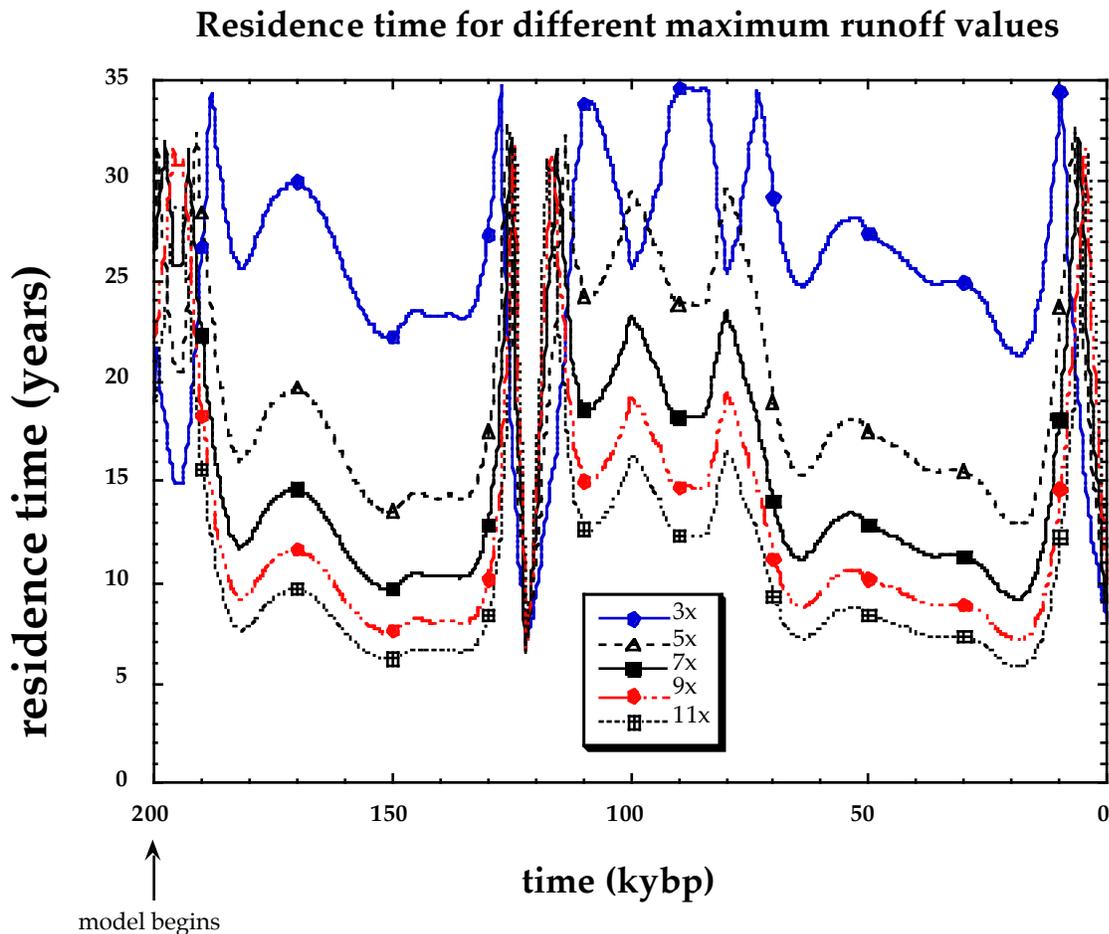


Fig. 3.19: Residence time for water in Owens Lake for 5 different values of glacial maximum runoff ranging from 2x modern to 10x modern runoff. Increasing runoff leads to decreasing residence time as water is pushed rapidly through the lake chain. Short residence times at ~123 ka and at present reflect

the small volume of Owens Lake and comparatively large evaporative flux out of the lake at each timestep.

DISCUSSION AND CONCLUSIONS

Now that the sensitivity of the modeled $\delta^{18}\text{O}$ values to various climatic parameters is known, we may attempt to create a best-fitting scenario to constrain further lake-level fluctuations. Glacial period precipitation and atmospheric water vapor may have been very much more depleted than modern values due to the 5-7°C cooler air temperatures that prevailed at those times (Porter, 1983; Dohrenwend, 1984; Spaulding and Graumlich, 1986; Stamm, 1991). The isotopic composition of precipitation changes with temperature at a rate of 0.7‰/°C (Benson, 1994). Therefore, a 5°C cooling may have led to a glacial Owens River isotopic value of -19.5‰ and an atmospheric isotopic value of -24.5‰. In Figure 3.20, we explore the effects of modulating temperature and isotopic composition of atmosphere and runoff between assumed glacial and modern values. Three pairs of model output are shown, each pair differing by the amount of glacial runoff prescribed. For each pair, one curve allows runoff and evaporation to vary between modern and assumed glacial values while the other parameters--atmospheric and runoff $\delta^{18}\text{O}$ and temperature--are held fixed at modern values. The other curve depicts models in which all parameters are allowed to vary between modern and assumed glacial values. We find that modulating runoff between modern and 7x modern values, modulating evaporation between modern and 0.7x modern values, setting the isotopic composition of the input Owens River water and atmosphere to -16‰ and -21‰ respectively, and setting the calcification temperature to 15°C yields a good visual match between modeled and measured $\delta^{18}\text{O}$ values (Fig. 3.21a). Between 70 and 110 ka, the model matches the measured isotopic data well. Furthermore, the degree of isotopic depletion centered about 38 ka in the measured data is generally matched by the low modeled isotopic values between 70 and 20 ka. The model predicts much more enriched isotopic values at 120 and 195 ka than are found in the data. However, as the measured isotopic data are sparse between 120 and 128 ka and again between 185 and 198 ka, we cannot confidently say that enrichments of magnitude similar to those in the modeled data did not occur.

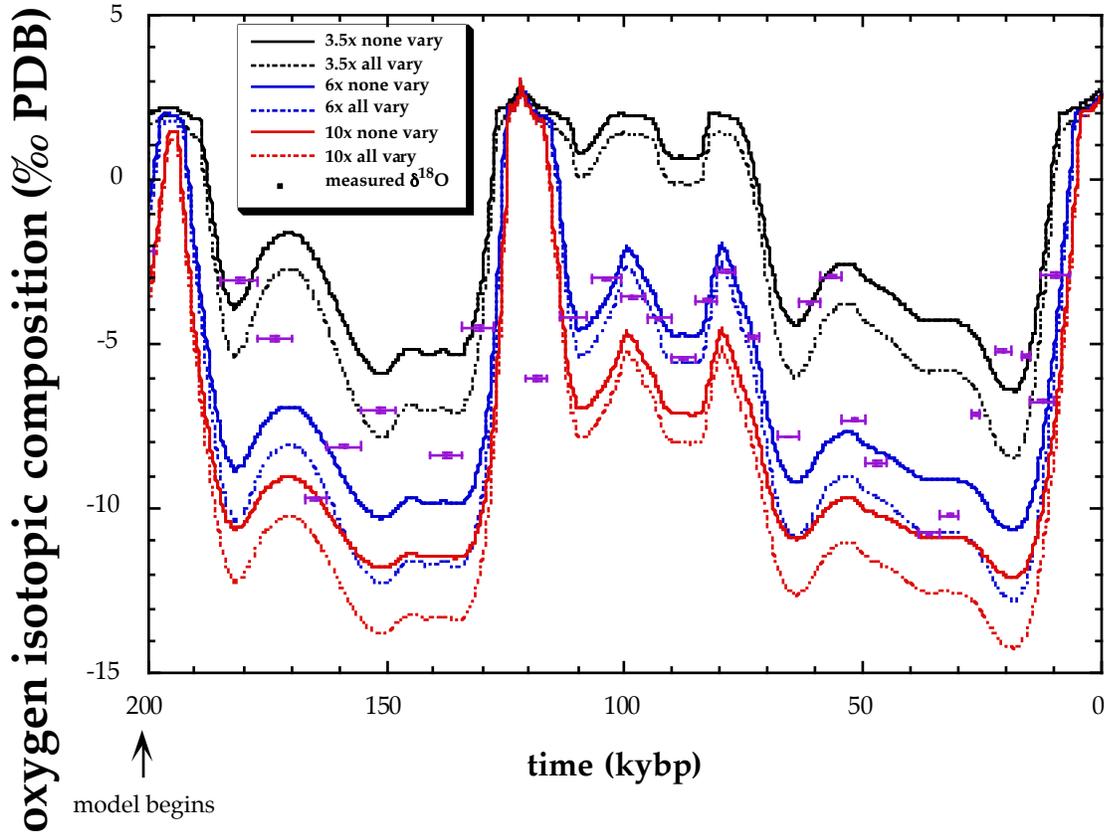


Fig. 3.20: Three pairs of model runs which differ in from each other in amount of glacial runoff prescribed. 3.5x refers to 3.5 x modern runoff during glacial maxima, likewise for 7x and 10x. For all model runs, glacial evaporation rate is specified as 70% of modern. None vary refers to model runs in which temperature and isotopic composition of input runoff and atmosphere are held fixed at modern values. All vary refers to runs in which these parameters are allowed to vary between modern and assumed glacial values (5 °C cooler).

Interestingly, in some ways the carbonate content data determined by Bischoff and others (in press (a)) more closely mimic the modeled isotopic fluctuations than they do the measured isotopic values (Fig. 3.21b). While there is clearly a chronological offset between the measured and modeled records, the carbonate content data show the same pattern of two small peaks (at 60 and 85 ka) preceded by a larger peak (at 100 ka) as do the modeled isotopic data (peaks at 80, 100 and 120 ka). This chronological offset may be due to error in the age/depth model constructed for the Owens Lake core, or to an actual lag in the response of the carbonate content proxy relative to the marine isotopic proxy. Whatever the cause for the offset, the similarities in the two curves are striking.

Figure 3.21c depicts the lake-level history devised for the best-fitting scenario. The history captures 4 periods of overflow from Searles Lake to Panamint Valley, two of which nicely correlate to episodes described by Jannik and others (1991). A small overflow from Panamint Lake to Death Valley (Lake Manly) is modeled at ~20 ka. This event is not seen in the chloride accumulation data of Jannik and others (1991), although a shallow saline lake has been identified in Death Valley that existed from 10-30 ka (Lowenstein and others, 1994; Ku and others, 1994; Li and others, 1994; Robert and others, 1994). We caution that the source of water responsible for creating the 10-30 ka Lake Manly is not presently known (Lowenstein, personal communication), however, (it could have come from the Amargosa river or from overspill from the Tecopa lake system) and therefore, this correspondence cannot be used as a test of the model.

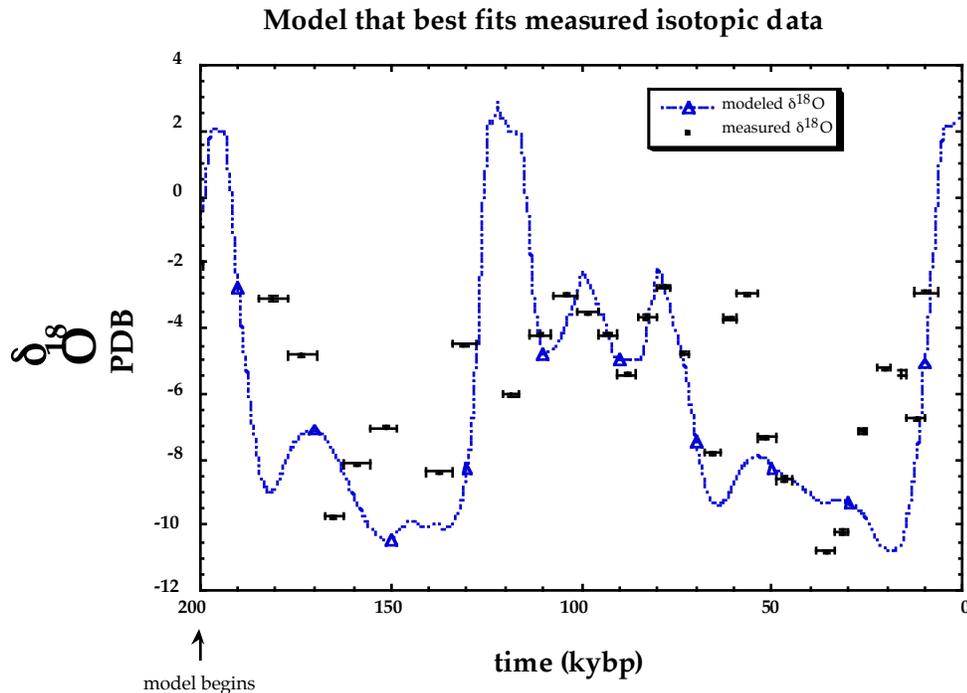


Fig. 3.21a: Measured and modeled oxygen isotopic composition of carbonates in Owens Lake. Measured values come from Benson and Bischoff (1993). Modeled values result from the following parameters: runoff varies between modern and 7x modern values with the marine $\delta^{18}\text{O}$ curve, evaporation varies between modern and 0.7x modern also with the marine $\delta^{18}\text{O}$ curve, input Owens River water and the atmosphere have $\delta^{18}\text{O}$ values of -16‰ and -21‰ respectively, relative humidity of the atmosphere over the lake is 35%, and carbonates are forming at 15°C .

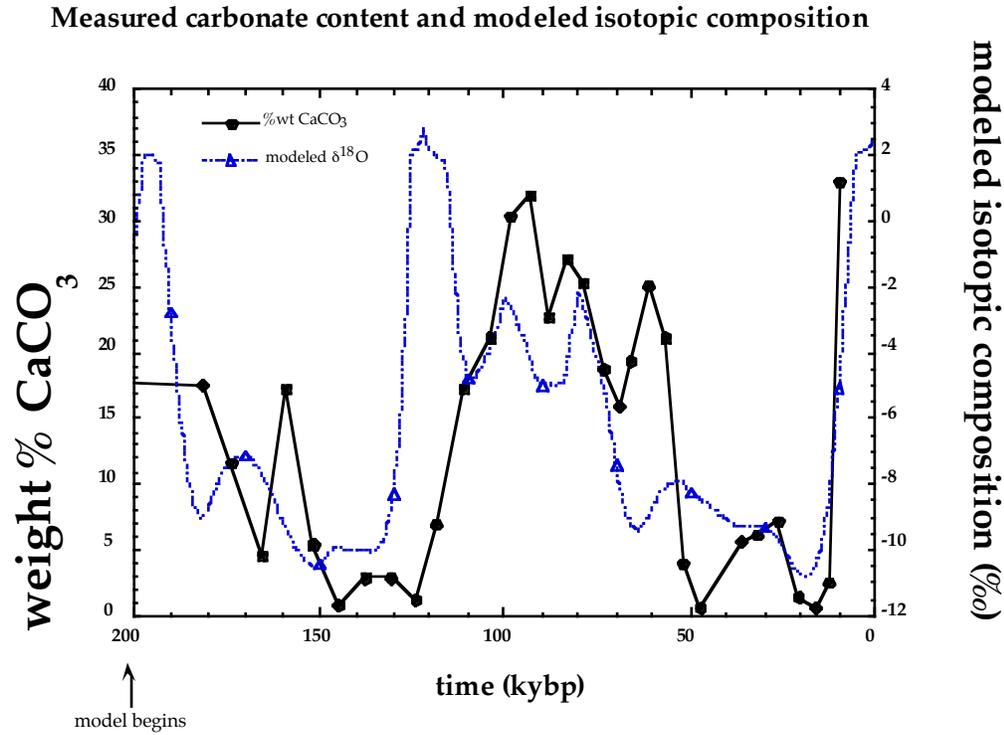


Fig. 3.21b: Comparison of modeled oxygen isotopic composition of carbonates in Owens Lake to measured carbonate content in the Owens Lake core sediments. Note that while the two curves appear to be chronologically offset from one another, their shapes are very similar. The model predicts that isotopic values of Owens Lake carbonates should be very enriched at 125 ka ($>2\text{‰}$), and somewhat less enriched at 80 and 100 ka (-2‰). These are periods of very low runoff and high evaporation rates. The measured isotopic data shown in Fig. 3.17a do not bear out the model predictions for the enrichment at 125 ka, with the modeled isotopic peak nearly the same amplitude as those at 100 and 80 ka. The carbonate content measurements, however, do show that the lake was more saline, and presumably more evaporatively concentrated.

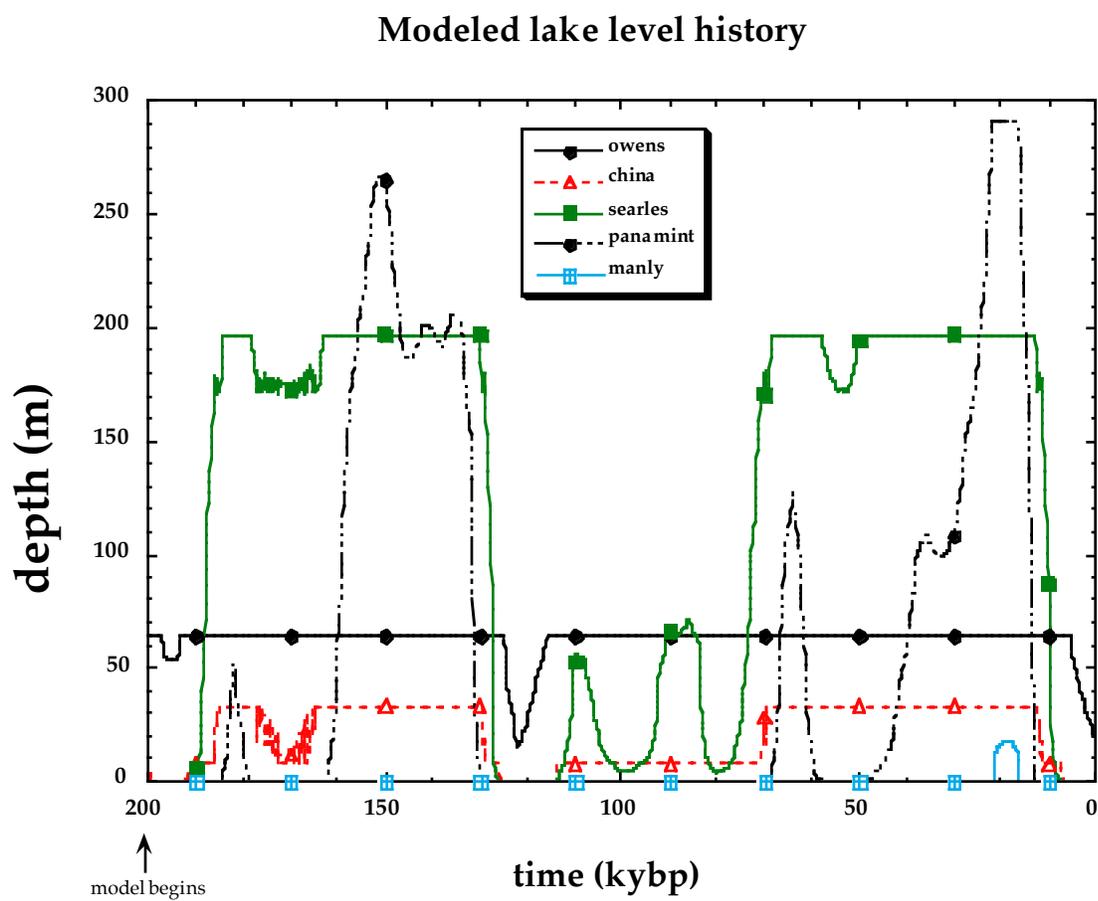


Fig. 3.21c: Lake level history associated with the model run that visually best fits the measured isotopic data. Modeled depths result from the following parameters: runoff varies between modern and 7x modern values with the marine $\delta^{18}\text{O}$ curve, and evaporation varies between modern and 0.7x modern also with the marine $\delta^{18}\text{O}$ curve.

We draw the following conclusions from the modeling exercise:

- The response time of Owens Lake to changes in climatic forcing is on the order of one century. The response time for the entire lake chain is about 300 years. Therefore, the lake chain should be at steady state with respect to long time-scale climatic fluctuations such as those induced by orbital forcing, and lake response should be lagged behind climate forcing by about one century. Thus, long-period forcings should be fully recorded in the sediments of Owens Lake.
- It takes only about 50 years to desiccate Owens Lake with modern evaporation values (smaller evaporation rates, as might have prevailed during glacial periods, require slightly longer periods) (Fig. 3.9). Given this result and the response time data, droughts of durations longer than 1 century (such as those reported by Stine, 1994) should be recorded in the sediments of Owens Lake in the form of evaporite layers. That these layers have not been identified indicates either that droughts did not occur in this area in the last 786 ka, or that the deposits dissolved upon reintroduction of water into the lake. As already noted, there exists no evidence for dissolution features in core OL-92, which favors the absence of >100 yr droughts.
- A maximum glacial runoff value of 10x modern, inferred from the empirical curves of Langbein and others (1949) and from estimates for paleo-temperature and -precipitation, is not supported. This value leads to modeled isotopic compositions that are much less enriched than the measured values (Benson and Bischoff, 1993), and to lakes that are too large. On the other hand, maximum glacial runoff of 3.5x modern, as suggested by Orndorff (personal communication), leads to modeled interglacial isotopic values that are too enriched.
- We capture fairly well the amplitude of oxygen isotope variation between glacial and interglacial periods when we assume a 7-fold increase in runoff and a 30% decrease in evaporation rates during glaciations relative to modern values (Fig. 3.21a); this corresponds to about a 2.7-fold increase in runoff during isotope stages 5a (~80 ka) and 5c (~100 ka). The details of the measured isotopic curve are not as well accounted for,

however. These details may be related to shorter-term climatic fluctuations than the Milankovitch time-scale variations we use to drive runoff. Such shorter-term fluctuations have been found in ice- and oceanic sediment-core studies around the North Atlantic region (Johnsen and others, 1992; Alley and others, 1993; Greenland Ice-core Project (GRIP) Members, 1993; Ruddiman and others, 1977; Broecker and others, 1990; Bond and others, 1993), and are associated with temperature variations of 7-13 °C acting over a few hundred to ~2000 years. These events motivate the higher-resolution study of Owens Lake core sediments described in Chapter 4, which addresses whether short-term fluctuations in water balance are observed in the eastern Sierra Nevada region.

- Comparison of Jannik and others' (1991) chronology to our modeled chronology in Figure 3.21c shows general agreement; our model shows overspill from Searles Lake to Panamint Valley between 130 and 160 ka and between 12 and 30 ka. Jannik and others reported overspill from 120-150 ka, and from 10-24 ka. However, we model two overspills into Panamint (from 60-70 ka, and from 180-185 ka) that are not reported by Jannik and others. Our model is successful in capturing the Holocene aridification noted by these authors and by Smith (1984) inasmuch as lake levels decline dramatically beginning at roughly 10 ka.

- The chronology shown in Figure 3.21c has Owens Lake spilling into Searles Lake for most of the last 200 kyrs. This conflicts with Bischoff and others' (in press (a)) estimates of the overspill using CaCO_3 content; that proxy leads to the conclusion that Owens Lake was closed for up to 50 kyrs at a time. However, studies conducted on cores of Searles Lake suggest that Owens Lake overspilled throughout much of the last 200 kyrs (Smith, 1984; Phillips and others, 1992) supporting the lake-level history reported here. The chronology indicates that any oxygen isotopic value more depleted than $\sim 0\text{‰}$ is associated with an overflowing Owens Lake.

ACKNOWLEDGMENTS

This research was funded by a National Defense Science and Engineering Graduate Fellowship awarded to KMM. Discussions with Greg Dick and Eric Small

proved enormously helpful. Thanks to Wes Danskin, Tim Lowenstein, and Rick Orndorff for sharing their ideas and data, and to Jim Bischoff and Bob Garrison for comments on the manuscript.