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Temperature and rainfall estimates for past 18 000 years in Owens Valley, California with a coupled catchment–lake model

Z. Yu[1,2], W. Dong[3], and P. Jiang[4]

[1]State Key Laboratory of Hydrology-Water Resources and Hydraulic Engineering, Hohai University, Nanjing, China
[2]Department of Geoscience, University of Nevada, Las Vegas, Las Vegas, Nevada, USA
[3]Department of Conservation & Natural Resources, Nevada Division of Environmental Protection, Las Vegas, Nevada, USA
[4]Division of Hydrologic Sciences, Desert Research Institute, Las Vegas, Nevada, USA

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Correspondence to: Z. Yu (zhongbo.yu@unlv.edu)

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Abstract

Closed-basin lakes are intricately linked to the hydrological systems and are very sensitive recorders of local hydro-climatic fluctuations. Lake records in closed-basins are usually used to investigate the paleoclimate condition which is critical for understanding the past and predicting the future. In this study, a physically based catchment–lake model was developed to extract quantitative paleoclimate information including temperature and rainfall over the past 18 000 years (ka) from lake records in a hydrologically closed basin in the Owens River Valley, California, US. The initial model inputs were prepared based on current regional climate data, boundary conditions from the General Circulation Model, and fossil proxy data. The inputs subsequently were systematically varied in order to produce the observed lake levels. In this way, a large number of possible paleoclimatic combinations can quickly narrow the possible range of paleoclimatic combinations that could have produced the paleolake level and extension. Finally, a quantitative time-series of paleoclimate information for those key times was obtained.

1 Introduction

The information on past climate change has been extracted from a variety of archives such as trees, ice, sediments, and corals (Anchukaitis et al., 2013; Marcott et al., 2013; Steinman et al., 2012; Viau et al., 2012). These archives can provide a qualitative interpretation of the history of Earth's climate. In past few decades, climate models emerge as an effective way to quantitatively test the hypotheses of past climate change (Kutzbach, 1987; COHMAP-Members, 1988; Street-Perrott and Harrison, 1985). However, they are not reliable at regional scale especially for variables such as precipitation (Kutzbach, 1987; Groppelli et al., 2011; Jiang et al., 2013). To overcome these shortcomings, an integrated utilization of numerical modeling and paleocliamte archives is proposed to provide quantitative paleoclimate information at region scale. Among these various paleoclimate archives, lake levels in closed basins are the most sensitive

HESSD
12, 6505–6539, 2015

indicators of the water balance between precipitation and evapotranspiration (Street-Perrott and Harrison, 1985).

In southern Great Basin, the regional distinctive tectonic settings and geomorphic characteristics create many hydrologically closed basins that were filled with water during pluvial lake periods (Smith and Street-Perrott, 1983; Street-Perrott and Harrison, 1985; Phillips, 1994; Bischoff et al., 1997; Menking et al., 1997; Lowenstein et al., 1999; Bischoff and Cummins, 2001). Most of them shrunk or dried up under the current dry and hot desert climate over this region. The evolution of lake size is determined by various climate variables such as wind velocities, relative humidity, temperature, and etc. Among them, precipitation is the most important one (Smith, 1991). For instance, lakes in Owens River system, California, are termed as "natural rain gauges" (Smith and Bischoff, 1997) as they can track the changes in precipitation within the catchments they are located at. However, interpretations of climate changes based on the water level changes is generally limited to identification of wetter or drier conditions, and provide little information about the specific nature of the climate change. The reason for this is that lake level in a particular basin is a complicated function of intra-basin and extra-basin climate and basin topography (Magny, 2004; Jones et al., 2001; Angel and Kunkel, 2010; Benson and Thompson, 1987). In order to extract quantitative paleoclimatic proxies from these lake records, one of the best approaches is through numerical modeling.

A variety of models have been used to simulate the paleo-record of closed basin lakes in arid and semiarid areas (Kutzbach, 1980; Benson, 1981, 1986; Hostetler and Bartlein, 1990; Hostetler and Benson, 1990; Hostetler et al., 1993, 1994; Ghile et al., 2014). Physically-based lake models, which explicitly represent the physical processes governing the energy and water balances of the lake, offer a more robust way to predict climate induced changes in water volume, level, and outflow of the lakes. A suitable lake model for paleolake level studies should require a minimum of site-specific parameters (Hostetler and Giorgi, 1993). In this paper, we developed a coupled catchment–lake model and used it to quantitatively estimate paleoclimate information, especially

HESSD
12, 6505–6539, 2015

annual mean precipitation and temperature in southwestern Great Basin since the last glacial maximum (LGM). To test the hypothesis that quantitative paleo-climate variables can be obtained through numerical modeling in the closed basin area, we first derived the past lake extent from field evidence, then we conducted an inverse modeling with a physically based lake model for the simulation of the derived lake extent at a specific time, and finally reconstructed the climatic conditions. Compared to the paleo-climate information derived from various archives or simulated by the climate models, the proposed method can provide quantitative estimate on temperature and precipitation at finer regional resolution.

2 Owens River system

The Owens River system is located at the western margins of the Great Basin. It is a hydrologically closed basin that consists of a chain of lakes including Mono Lake, Owens Lake, China Lake, Searles Lake, Panamint Lake, and Death Valley Lake (Fig. 1). The floors of all these lakes except Mono Lake are now occupied by playa lakes or salt flats. The valley is bound on the west by the Sierra Nevada, on the northeast by the Inyo and White Moutains, and on the southeast by the Coso Range. Presently, Owens River drains an area of about 8550 km^2. Due to the strong rain shadow effect, most of the runoff is derived from about 16 % of the catchment area, which lies on the eastern slope of the Sierra Nevada (Lee, 1912). Modern climate in the floor of the Owens River system is semi-arid with about 15 cm of annual precipitation. Thus, precipitation that fell directly on the surface of the basins is an insignificant contribution to the lake water budget, which could also be true for the lakes in the paleo-Owens River system (Jannik et al., 1991). Street-Perrott and Harrison (1985) termed such lakes as “amplifier” lakes, which describe a simple relationship among basin runoff, lake evaporation, and lake area (Smith and Bischoff, 1997).

Owens Lake, at the base of high mountains, responds firstly to the increasing regional precipitation. The Owens Lake is the terminal of the Owens River, it was about

HESSD
12, 6505–6539, 2015

10 m deep and 290 km^2 in area before agricultural irrigation became extensive in 1912. All of the river's water was diverted to Los Angeles in 1912, and the lake desiccated. Searles Lake was the third in a chain of five permanent lakes receiving water from the Owens River during the late Pleistocene, and Mono Lake was separated from the Owens River system to the south by a high-altitude sill in the late Wisconsin (Benson and Thompson, 1987). During the LGM, Owens Lake, China Lake, and Searles Lake overflowed and the Panamint Valley is a terminal of the Owens River hydrological system. Lake stages at Searles Lake are sensitive to climatic changes because of the storage capacities of the upper lakes in the series. Inflow to Searles Lake depends on overflow from the other lakes and therefore is first affected by a decreasing inflow in their lake system.

Studies on lacustrine outcrops, cores, and landforms have allowed the reconstruction of the past histories of lakes in the Owens River system and its downstream basins (Smith and Street-Perrott, 1983; Smith and Bischoff, 1997). The paleo lake levels are recorded by the geomorphic and sedimentary evidence including staircases of abandoned shorelines and abrupt changes of facies in sediments. Smith and Street-Perrott (1983) provide a chronology of Late Wisconsin to present lake level fluctuations for Searles Lake (Fig. 2).

Benson et al. (1997) identified two hiatuses at 2.25 and 9.2 m based on the ^{14}C data of Core OL84B from Owens Lake. These two hiatuses represent two desiccation events that occurred ~ 15.3–13.5 and ~ 6.1–4.3 ka (all ages used in the text of this paper are on ^{14}C time scale before present (BP)) in Owens Lake. The $\delta^{18}O$ data of sediments between the two hiatuses show four abrupt dry/wet oscillations that have their correspondents in the North Atlantic region. Relatively wet intervals precede each of the dry events. An extreme overflow occurred at about 12 ka, which resulted in the lowest $\delta^{18}O$ (−13 ‰) of lake carbonate (Benson, 1999). Based on the ostracode assemblage from the Owens Lake, Forester (2000) derived more details on lake level changes of the Owens Lake from 25 to 4 ka; Li et al. (2000) provided detailed information on climate

HESSD
12, 6505–6539, 2015

for the past 1000 years; and Benson and Paillet (2002) plotted $\delta^{18}O$ with age for past 18 ka (Fig. 2).

Bacon et al. (2006) suggested that pluvial Owens Lake had dropped 45 m from its latest Pleistocene highstands of 1145 m by 11 600 yr BP. This lowstand was followed by an early Holocene transgression that attained a highstand near 1135 m before dropping to 1120 m at 7860–7650 yr. The lake then lowered another 30 m to shallow and near desiccation levels between 6850 and 4300 yr BP and minor lake-level rise after 4300 yr BP, followed by alkaline and shallow conditions during the latest Holocene.

In summary, the detailed paleolake records in the Owens River system offer a good opportunity to extract quantitative paleoclimate information in the western Great Basin.

3 Description of model and modelling strategies

The surface area of a closed-basin lake under natural conditions is strictly dependent on the dynamic equilibrium between precipitation and evapotranspiration over its entire catchment (Halley, 1714). Any changes in this equilibrium result in a change in terminal lake depth, which directly influences its area, and the cumulative lake area in the drainage basin (Benson and Paillet, 1989). The mean annual water balance of a lake is governed by the equation (Street-Perrott and Harrison, 1985):

$$\Delta V = A_L(P_L - E_L) + (R - D) + (G_I - G_O) \tag{1}$$

where ΔV is the net change in volume of the lake, P_L is precipitation on the lake, E_L is evaporation from the lake, A_L is the area of the lake, R and D are runoff from the catchment and the surface discharge from the lake, respectively, and G_I and G_O are groundwater flows into and out of the lake respectively. For a closed basin lake, G_I and G_O can be assumed negligible, and D is zero, so Eq. (1) reduces to the following form for equilibrium conditions:

$$R = A_L(E_L - P_L). \tag{2}$$

HESSD
12, 6505–6539, 2015

If the runoff from the drainage basin can be represented by

$$R = A_B(P_B - E_B) \quad (3)$$

where A_B represents area of the catchment, P_B is the precipitation over the catchment, and E_B is the evapotranspiration over the catchment, then we have

$$A_B(P_B - E_B) = A_L(E_L - P_L). \quad (4)$$

This simple expression shows that the equilibrium area of a closed lake under natural conditions is strictly dependent on the precipitation and evaporation over its catchment and water surface. In the Owens River system, based on paleolake records, values for A_L and A_B for those paleolakes can be measured quite accurately using a digital elevation model (DEM). Remaining components in Eq. (4) are precipitation and evaporation over the drainage basin. The evaporation mainly depends on temperature, thus the purpose for this paper is to develop a coupled catchment–lake model to resolve these two unknown variables in the Eq. (4).

The evaporation value depends on many climatic factors including solar radiation, temperature, wind speed, and cloud cover. With the exception of temperature, other relevant factors are difficult to reconstruct from geologic data. Many studies assume that paleo-values for the evaporation can be satisfactorily estimated from empirical relationships between modern data on evaporation and air temperature. However, a change in evaporation rates could result from higher wind velocities, higher relative humidities and lower solar radiation values, and greater amounts of precipitation on the lake surface (Smith and Street-Perrott, 1983). Therefore, it is desirable to have a model that considers all of these factors. Hostetler and Bartlein (1990) developed one-dimensional surface energy-balance lake model, where the vertical heat transfer was simulated by eddy diffusion and convective mixing. Several studies using this model have successfully simulated the modern and paleolake level change both in humid and arid regions (Vassiljev, 1997; Hostetler and Benson, 1990; Hostetler et al., 1994). Orndorff (1994) developed a surface hydrologic model (OSHM) that has been successfully applied to

HESSD
12, 6505–6539, 2015

the Owens River system to test the proxy estimates of the LGM against the paleo-lake records. The OSHM has three modules: the snow module that computes mean monthly snowfall, snowmelt, snowpack, ice accumulation, ice transport, and icemelt for each grid cell based on the input of temperature, precipitation, and elevation of that cell, the runoff module that calculates mean annual runoff from available water (rain, snowmelt, and icemelt), and the lake module computes lake extent from the results of basin-wide mean annual runoff calculated by the runoff module and lake evaporation. In the OSHM, empirical relationships between modern data on evaporation and air temperature were used to calculate the evaporation for the pluvial lakes in the Owens River system during the LGM, which may not represent the actual situation as discussed above. In this study, the lake module in the OSHM was modified with the addition of Hostetler's lake model and used to simulate evaporation over the lake surface.

The paleoclimatic evaluation strategy is first to model lake extent under current climatic conditions with the coupled catchment–lake model developed in this study. The input parameters under current climatic conditions are then systematically varied in order to reconstruct lake extent based on lake records under paleoclimatic conditions. The modeling strategy is essentially an inverse approach to inferring paleoclimatic conditions based on past lake extent. The advantage of this modeling strategy is that a large number of possible paleoclimatic combinations can be quickly narrowed to a possible range of temperature/precipitation combinations that could have produced a particular paleolake extent.

A simulation on lake extent begins with dry closed basins In order to prevent water from overflowing into previously considered basins, basins are taken into account in order from the highest to lowest active outlet elevation, which means that the Monon Lake is first full and overflows to Owens Lake, then the Owens Lake is full and overflows to China Lake, then China Lake is full and overflows to Sears Lake, the Sears Lake is full and overflows to Panamant Lake, finally the Panamant Lake is full and overflows to Death Valley Lake. Each basin's runoff volume from its catchment is added to the current lake volume at each time step. The lake volume is compared to the basin

HESSD
12, 6505–6539, 2015

threshold volume, which corresponds to a lake level equal to the controlling elevation of the lowest basin outlet. Overflow occurs into the basin on the other side of the outlet when the lake volume exceeds the threshold volume. The model also checks for lake merging during overflow, which occurs when two lakes with a common active outlet overflow, thus inundating the active outlet. If two lakes merge, the downstream basin becomes a part of the upstream basin, and the remaining outlets of both basins are sorted to determine the new active outlet for the complex basin (Orndorff, 1994). Lake evaporation is then calculated using the eddy diffusion and convective mixing (Hostetler and Bartlein, 1990), and the lake level is adjusted accordingly. The simulation runs in one year time step until the cumulative lake volume equilibrates or the run time exceeds a specified end time. Benson and Paillet (1989) state that "the proper gage of lake response to change in the hydrologic balance is neither lake depth (level) nor lake volume but instead lake surface area", thus this study focused on lake surface area for a comparison of simulated lake extent and derived lake extent based on geologic evidence.

4 Calibration of catchment–lake model

The catchment–lake model used in this study was developed by coupling a distribution hydrology model (Orndorff, 1994) and an energy-balance lake model (Hostetler and Bartlein, 1990). Both of these two models were independently calibrated with observed data (Orndorff, 1994; Hostetler and Bartlein, 1990; Hostetler, 1991), thus they are valid when use them independently. However, the catchment–lake model developed for this study has to be calibrated before applying it to simulate paleolake levels. The Mono Lake is presently only lake with standing water in Owens Valley. The three major streams (Rush, Lee Vining, and Mill Creeks) that delivery water to Mono Lake originate in the high Sierra Nevada (Benson et al., 1990), so the hydrological characteristics of Mono Lake and Owens Lake is similar. Observed data including climate data, hydrological data and lake level data are available for the Mono Lake drainage basin

HESSD
12, 6505–6539, 2015

since 1857 (Mono-Basin-Environmental-Impact-Report, 1993). The data on measured temperature profiles and lake evaporation are also available for some periods of time (MacIntyre et al., 1999). The calibration was done with input data including modern precipitation and temperature matrix data for the Mono Lake drainage basin from the LCM (Stamm, 1992), solar radiation data from the Desert Rock, vapor pressure, wind speed and cloudiness data from stations close to the Mono Lake. The simulated monthly runoff for the Mono Lake drainage basin is compatible with the observed (Fig. 3) and the annual runoff is about 1 % less than the observed. The simulated lake surface area is 224 km^2 that is 1.8 % less than the average of the observed lake surface area from 1940 to 1989 (Mono-Basin-Environmental-Impact-Report, 1993). The lake temperature profile simulated by the model agreed very well with measured temperature profile (Fig. 4a). Furthermore, the simulated evaporation also compares well with the observed evaporation data through the grant pan, but slight higher than the evaporation estimated from water budget method (Fig. 4b). These comparisons indicated that the overall ability of the coupled catchment–lake model developed here to reproduce observed basin-wide mean annual runoff, mean lake surface area, temperature profile and evaporation of lake water in modern Mono Lake drainage basin.

5 Input parameters

A number of input parameters are required for the coupled catchment–lake model. Coarse grid cell (5 km × 5 km) used in the OSHM missed some small snow cover and stream networks, and did a poor job representing some basin shapes (Orndorff, 1994). In this study, the fine resolution (1 km × 1 km) data was used to obtain better results. The topographic data used in this model is from the global 30 arcsec elevation data set (GTOPO30) (https://lta.cr.usgs.gov/GTOPO30). The GTOPO30 has a horizontal grid spacing of 30 arcsec (approximately 1 km). Observed solar radiation, cloud cover, wind speed, atmospheric pressure from near weather stations, and modern monthly temperature and precipitation matrix from local climate models (Stamm, 1992) that are based

HESSD
12, 6505–6539, 2015

on boundary conditions including terrain, wind field, and radiation balance were used to drive the newly developed catchment–lake model and to reproduce the historical lake level of the Owens Lake. Precipitation and temperature matrix data at 18, 15, 12, 9, and 6 ka were prepared based on the proxy data in Table 1 by applying the appropriate perturbation (simple additive change for temperature, and multiplicative change for precipitation) to the modern monthly precipitation and temperature matrix data from the LCM. For example, climate at the LGM might be hypothesized to be 5° colder and 50 % wetter than the present based on proxy data in Table 1. The input climate matrix at the LGM could be prepared for temperature by subtracting 5 °C from the modern temperature matrix, and for precipitation by multipling 1.5 to modern precipitation matrix. Other climate parameters including cloud cover (Fig. 5a), solar radiation (Fig. 5b and c), and wind speed (Fig. 6) were from historic records for modern conditions, and from the Community Climate Model (CCM0) (Kutzbach and Guetter, 1986) and the results of paleoclimate simulation of North America (Bartlein et al., 1998) for paleoclimatic simulation. However, the single monthly value for these parameters was used for whole area. The reason for this is: first, there is no such final resolution data available for these parameters in the study area; second, previous study indicated that precipitation and temperature are two primary factors controlling the glacial extent (Plummer and Phillips, 2003).

As most observation data are available only for Owens Lake and Searles Lake, the comparison between the simulated results and the observation data in this study focused on these two lakes.

6 Results

6.1 Simulation on modern lakes

Mean annual runoff, computed for modern climate from the runoff module is input to the lake model to simulate modern lake extent in the Owens Valley. The resulting lakes

HESSD
12, 6505–6539, 2015

along with hillshed and stream network that were derived from the DEM are shown in Fig. 7a. The lake system converges in 80 years. There is no lake mergence occurring. The results from modern simulation accurately portray Mono Lake, Lake Crowley (in the Long Valley basin), Black Lake (in Adobe Basin), and Owens Lake. Simulated Mono Lake has a surface area of 224 km^2 that is about 2 % less as compared to the average 228 km^2 of the observed lake surface area from 1940 to 1989 (Mono-Basin-Environmental-Impact-Report, 1993). Simulated Owens Lake has a surface area of 302 km^2 that is about 4 % more as compared to an observed pre-diversion surface area of 290 km^2 (Smith and Street-Perrott, 1983).

6.2 Simulations on lakes at Last Glacial Maximum

Simulations on lakes at the LGM in the Owens Valley were done with modern temperature and precipitation matrices perturbed based on proxy-based LGM temperature and precipitation departures (Table 1), and other climate parameters including solar radiation, cloud cover, and wind speed from CCM0 (Kutzbach and Guetter, 1986) and the results of paleoclimate simulation of North America (Bartlein et al., 1998) that were fixed. By varying combinations of temperature and precipitation with appropriate perturbation until the derived lake extent at the LGM from field evidence was reproduced, the final combination with temperature 5.5 °C cooler and 1.25 times precipitation of modern climate conditions was obtained. Simulated final lake extents with hillshed and stream network are shown in Fig. 7f. Owens Lake overflows, and has a lake surface area of 692 km^2. China Lake and Searles Lake coalesce and have an area of 949 km^2. Searles Lake also overflows and a small lake with an area of 144 km^2 was formed in Panamint Valley. These results are very compatible with observed lake extent in Owens Valley at the LGM reported by Smith and Street-Perrott (1983).

HESSD
12, 6505–6539, 2015

6.3 Simulation on lakes at 15 ka

The techniques used to prepare input data for simulations on lakes in Owens Valley at 15 ka are same as them for the LGM. Based on ^{14}C and sedimentary features, Benson et al. (1997) reported a desiccation event would occur for Owens Lake at 15 ka. Because Owens Lake is an upstream lake of Searles Lake, and Searles Lake received most of its inflow from the overflow of Owens Lake, Searles Lake could also desiccate at 15 ka. This is supported by the field evidence that the Searles Lake was at its low water level with an elevation at 510 m. Therefore, the simulation on lake extent in Owens Valley at 15 ka is to find a combination of precipitation and temperature that can create a dry Owens Lake and Searles Lake. After multiple runs, a combination with temperature 1.8 °C cooler and precipitation 20 % less than modern climate condition could produce a dry Owens Lake and Searles Lake (Fig. 7e). The results from this simulation also indicated that a significant decline in the level of Mono Lake. This is consistent with possible hiatuses in cores from the Mono Lake basin (Newton, 1991) and major declines in the levels of Mono Lake and Lake Lahontan (Benson et al., 1998, 1996).

6.4 Simulation on lakes at 12 ka

The $\delta^{18}O$ data from Core OL84B drilled in Owens Lake indicated the lowest values of $\delta^{18}O$ at 12 ka for the last 15 ka (Fig. 2) (Benson et al., 1997). This represents highest ratio of overflow to inflow into Owens Lake, which implied that Searles Lake probably also received its highest inflow for the last 15 ka. The field evidence indicated that the lake level of Searles Lake started to increase at 12 ka and reach its highest level at 11 ka (Fig. 2) (Smith and Street-Perrott, 1983). It can be expected that the overflow from Searles Lake might finally reach its highest, and the largest lake might be formed in Panamint Valley in the last 18 ka. Based on multiple runs, a combination of temperature 4.5 °C cooler and 1.8 times precipitation of modern climate conditions could reproduce a lake system with the highest lake level in Panamint Valley (Fig. 7d) since the LGM. However, the lake level of Panamint Lake was still not high enough for overflow.

6.5 Simulation on lakes at 9 ka

$\delta^{18}O$ values of Owens Lake at 9 ka are around 27 ‰ indicating Owens Lake at its hydrological closure (Benson et al., 1996). A dry event was recognized based on the presence of prismatic cracking that suggests the existence of a soil formed during subaerial exposure of lake sediments at about 9 ka (Benson et al., 1997). In meantime, Searles Lake was at its lowest water level since the LGM (Smith and Street-Perrott, 1983). A combination of temperature 0.5 °C warmer and 1.2 times precipitation of modern climate conditions could reproduce a lake system in Owens Valley at 9 ka (Fig. 7c).

6.6 Simulation on lakes at 6 ka

Owens Lake was at its second desiccation event at 6.1–4.3 ka (Benson et al., 1997), and Searles Lake was at its water level about same as today (Smith and Street-Perrott, 1983), thus the lakes in Owens Valley at 6 ka might have shallower water stand than them under modern climatic conditions. Based on this evidence, a combination of temperature 1.2 °C warmer and 0.9 times precipitation of modern climate conditions could produce a lake system in Owens Valley at 6 ka (Fig. 7b).

7 Sensitivity analysis

Sensitivity analysis was only performed for the simulation of lakes in Owens Valley at the LGM (18 ka). Based on the proxy data at the LGM in Table 1, a combination of the lowest temperature (7.5 °C cooler than modern temperature) and the maximum precipitation (2.40 times modern precipitation) was used to prepare the input data for simulation on lake extent. The results from simulation based on this combination indicate that all basins including Death Valley in Owens River system are full of water,

HESSD
12, 6505–6539, 2015

which is not case in the last 18 ka. Another combination of the highest temperature (3.0 °C cooler than modern temperature) and the lowest precipitation (1.2 times modern precipitation) was used to simulate lake extent in Owens Valley. The results from this simulation indicate that Searles Lake is not full and no lake was formed in Panamint Valley. The results from these two extreme cases of the combination are against to geological evidences. First, Panamint Lake did not overflow, and there is no full lake formed in Death Valley in the last 18 ka (Smith and Street-Perrott, 1983); second, Mono Lake was separated from the Owens River system to the south by a high-altitude sill in the late Wisconsin (Benson and Thompson, 1987); third, Searles Lake was full and overflowed into Panamint Valley where a small lake was formed at the LGM (Smith and Street-Perrott, 1983). Therefore, the coupled catchment–lake model developed in this study is very sensitive to a change of both temperature and precipitation, and it can be easily used to infer paleoclimatic conditions based on past lake extent.

8 Discussion

The coupled catchment–lake model developed in this study has several advantages for paleoclimatic interpretation of paleolake extent. By using physically based models to find the climatic conditions that could produce a particular lake extent at specific time, a quantitative estimate on temperature and precipitation against field evidences was obtained. This approach allows direct consideration of the effects of changes in both precipitation and temperature, as well as numerous climate variables including cloud cover, solar radiation, and wind speed. Simulated lake surface areas and elevations of lake surface for major lakes in Owens Valley in the last 18 ka are listed in Table 2. The combinations of temperature (Fig. 8a) and precipitation (Fig. 8b) that could produce observed lake extent in Owens Valley in the last 18 ka were obtained from the simulations above. We also plotted reconstructed temperature and precipitation based on pollen in Owens Lake core OL84B in Fig. 8. Simulated combination of temperature and precipitation at the LGM (18 ka) is 5.5 °C cooler and 1.25 times of modern climate

HESSD
12, 6505–6539, 2015

conditions. This result is close to a proxy data (Merrill and Péwé, 1977) in Table 1, but different from other proxy data in Table 1. This could be a result that the proxy-based estimates of LGM climate are representative of the area in which the fossil assemblage or glacial feature was studied. A mixed conifer forest from Kings Canyon that includes red fir, western juniper, incense cedar, sugar pine, ponderosa pine, California nutmeg, and single-needle pinon pine (Cole, 1983) indicated a colder than today with near-modern precipitation levels in the southern Sierra Nevada at 18 ka. The 9.2 m hiatus found in Owens Lake core OL84B was dated at 15.5 to 13.5 ka (Benson et al., 1997). Our simulation indicated lake extent at 15 ka is the smallest in the last 18 ka. The combination of precipitation and temperature produced this smallest lake extent is the lowest precipitation in the last 18 ka, and about 1.8 °C cooler than modern temperature. The lake extent at 12 ka in Owens Valley is largest in the last 18 ka. Simulated combination of precipitation and temperature at 12 ka is 4.5 °C cooler and 1.8 times of modern climate conditions. This result is consistent with the pollen data from the core OL84B (Fig. 8a, b) (Mensing, 2001). The pollen data indicated that a mean annual precipitation of 308–370 mm and a > 80 % increase in effective moisture, and 4 to 5 °C cooler than the present mean temperature in Owens Valley start from 13.5 ka. A very wet climate in the western Great Basin at this time is also supported by the δ^{18}O data from Owens Lake (Benson et al., 1996) and Mono Lake (Benson et al., 1998), and ages of tufa from Searles Lake (Lin et al., 1998) and Lake Lahontan (Benson, 1993) that indicate high lake stands from 14 to 13.5 ka. However, the climate in Sierra Nevada shifted from cool, wet conditions to possibly a more seasonal climate with cool, wet winters and warmer summers (Mensing, 2001; Smith and Anderson, 1992), but still wetter and cooler than today (Spaulding, 1983) after 12 ka. Simulated lake extent in Owens Valley at 9 ka indicated a significantly decrease in lake extent from 12 ka. The last deglaciation was interrupted by a worldwide cooling event, the Younger Dryas (YD) from 11 to 10 ka. Studies from western North America have identified late-glacial climatic oscillations roughly synchronous with the YD interval (Stuiver et al., 1995). The pollen data from the core OL84B indicates a series of abrupt climatic oscillations between 10.8 to

HESSD
12, 6505–6539, 2015

9.5 ka, but is not sufficient to clearly define the YD for direct comparison with other sites (Mensing, 2001). Summer insolation reached a maximum between 9 to 8 ka, resulting in higher summer temperatures and probably increased seasonality (Grigg and Whitlock, 1998). Low lake levels and the increased dominance of desert shrubs indicate the beginning of warm, dry Holocene conditions. The results from our simulations indicate 0.5 °C warmer and 1.2 times of modern climate conditions could produce observed lake extent at 9 ka, which is generally agreement with high isolation and increased desert shrubs. Second hiatus found in the core OL84B indicates that Owens Lake was probably dry at 6 ka (Benson et al., 1997), and quantitative analysis of the pollen record from Sierra Nevada suggests temperatures 1.4–2.1 °C warmer than today (Adam and West, 1983). The lake level of Searles Lake was also low at 6 ka (Smith and Street-Perrott, 1983). Simulated lake extent with a combination of temperature 1.2 °C warmer and 0.9 times precipitation of modern climate conditions is consistent with geologic evidences.

The climate in Owens Valley after 6 ka is probably similar to modern climate conditions. However, a slightly increase in precipitation and decrease in temperature could happen, because the historic lake level of Owens Lake is higher than it at 6 ka. An increased frequency of modern extreme storm events in Mojave River watershed in late Holocene was concluded based on lake deposits in the Silver Lake playa, CA (Enzel et al., 1989). The relatively high lake level of Searles Lake from 5 to 3 ka could be a result of an increased frequency of modern extreme storm events and summer monsoon circulation (Bush, 2001).

9 Conclusions and future work

The coupled catchment–lake model developed in this study is capable of accurately simulating lake extent as a function of modern climate and paleoclimate. The purpose for this model is used to quantitatively estimate paleoclimate, especially annual precipitation and temperature against field evidence in a catchement-lake watershed hydrologic system. The simulations on lake extent at 18, 15, 12, 9, 6 ka, and modern

HESSD
12, 6505–6539, 2015

climate conditions are very compatible with observed or derived data. The annual precipitation and temperature of Owens Valley for these times are generally in agreement with the proxy data that are derived from Owens Valley and the places near Owens Valley. The accuracy of our quantitative estimates in paleotemperature and paleoprecipitation in Owens Valley is completely dependent on the accuracy of the field observations, especially the elevation of paleo-shorelines and their chronology for the lakes. Therefore, numeric values for the temperature and precipitation at 18, 15, 12, 9, and 6 ka are only effective for the geologic evidences used this study. However, these numeric values of paleotemperature and paleoprecipitation can be adjusted based on new geologic evidence. The two advantages of the coupled catchment–lake model are: (1) based on the proxy data, the possible range of temperature/precipitation combinations that could produce a particular paleolake extent can be easily obtained by narrowing a large number of possible paleoclimatic combinations, (2) the model developed in this study is a physically based model that requires a minimum of site-specific parameters (Hostetler and Giorgi, 1993), thus it can be applied in any lakes if input parameters are available.

Simulations performed in this study did not consider that the seasonal distribution of precipitation and temperature in the last 18 ka might be different from modern climatic conditions. The reason for this is that we seldom have information on the seasonal distribution of precipitation and temperature in the past. However, the lake levels of the Mediterranean region were significantly affected by the seasonal distribution of temperature (Prentice et al., 1992). Therefore, it is very important to consider the seasonal distribution of paleoclimate into the simulation on the paleolake extent in Owens Valley with accumulation of more data on seasonal distribution of paleoprecipitation and paleotemperature. Besides, an initially dry lake was assumed for the simulation, which is not true for most situations. Thus, the simulation could be improved by running the model continuously a high lake level at 18 ka to a historic lake level. A time-series of temperature and precipitation from the continuous simulation is more useful than the discrete results in this study.

HESSD
12, 6505–6539, 2015

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HESSD
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HESSD
12, 6505–6539, 2015

Temperature and rainfall estimates for past 18 000 years in Owens Valley

Z. Yu et al.

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HESSD
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HESSD

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Z. Yu et al.

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HESSD
12, 6505–6539, 2015

Table 1. Proxy data in the southwest US in the last 18 ka.

Time	Temperature (°C)	Precipitation (%)	Source
Late Wisconsin	−6.25 annual	+37 cm winter	Dohrenwend (1984)
	−3.0 annual	+68 % annual	Mifflin and Wheat (1979)
	−5.5 annual	+27.5 % annual	Merrill and Pewe (1977)
	−6.5 annual	+65 % winter, −45 % summer	Spaulding (1985)
	−6.0 summer	Summer precipitation 10 % of annual	Betancourt (1990)
	−6.7 annual	+32 % annual	Cole (1990)
	+1.0 winter, −1.0 summer	+57 % winter, +56 % summer	Leffler and Cochran (1989)
	−3.0 Jan, −3.0 Jul	+19 mm Jan, −31 mm Jul	Spaulding and Graumlich (1986)
	−5.5 to −4.9 annual	1.4 to 1.7× model annual	Thompson et al. (1999)
18 ka	−3.29 annual	−0.29 mm day^{-1} annual	Thompson et al. (1993)
	−3.17 Jan, −3.01 Jul	+0.25 mm day^{-1} Jan, −0.84 mm day^{-1} Jul	
20.5 to 18 ka	−7.5 annual	2.40×	Thompson et al. (1999)
14 to 11.5 ka	−6.7 annual	2.58×	Thompson et al. (1999)
12 ka	−2.52 annual	−0.18 mm day^{-1} annual	Thompson et al. (1993)
	−3.01 Jan, −0.63 Jul	−0.27 mm day^{-1} Jan, −0.15 mm day^{-1} Jul	
9 ka	+0.43 annual	+0.30 mm day^{-1} annual	Thompson et al. (1993)
	−0.09 Jan, +2.15 Jul	+0.80 mm day^{-1} Jan, −0.27 mm day^{-1} Jul	
6 ka	+0.69 annual	−0.03 mm day^{-1} annual	Thompson et al. (1993)
	+0.30 Jan, +0.68 Jul	−0.16 mm day^{-1} Jan, +0.07 mm day^{-1} Jul	

HESSD
12, 6505–6539, 2015

Table 2. Simulated lake extent and elevation of lake levels in Owens Valley in the last 18 ka.

Age (ka)	Mono Lake Elevation (m)	Mono Lake Area (km^2)	Owens Lake Elevation (m)	Owens Lake Area (km^2)	Searles Lake Elevation (m)	Searles Lake Area (km^2)	Panamint Lake Elevation (m)	Panamint Lake Area (km^2)
18	2040	461	1145	692	688	949	340	144
15	1952	258	1070	30	475	2	310	3
12	2120	689	1145	692	688	949	350	349
9	1978	270	1100	302	525	252	325	94
6	1949	172	1075	96	490	4	320	30
0	1958	227	1097	302	515	225	320	30

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HESSD

12, 6505–6539, 2015

Temperature and rainfall estimates for past 18 000 years in Owens Valley

Z. Yu et al.

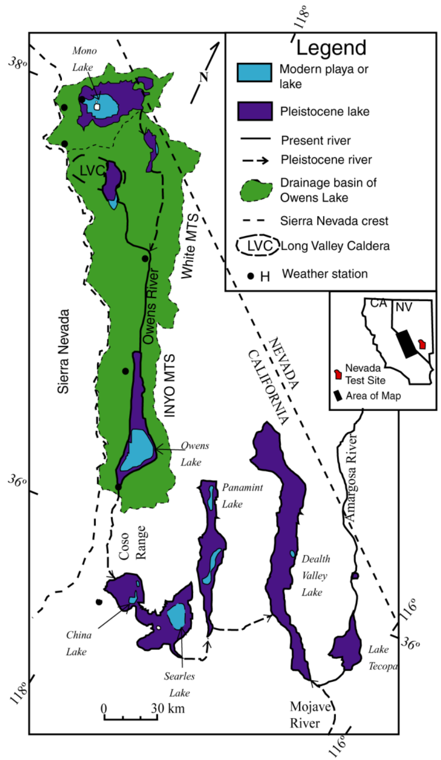

Figure 1. Location map of Owens River system (modified from Smith and Bischoff, 1997).

HESSD
12, 6505–6539, 2015

Temperature and rainfall estimates for past 18 000 years in Owens Valley

Z. Yu et al.

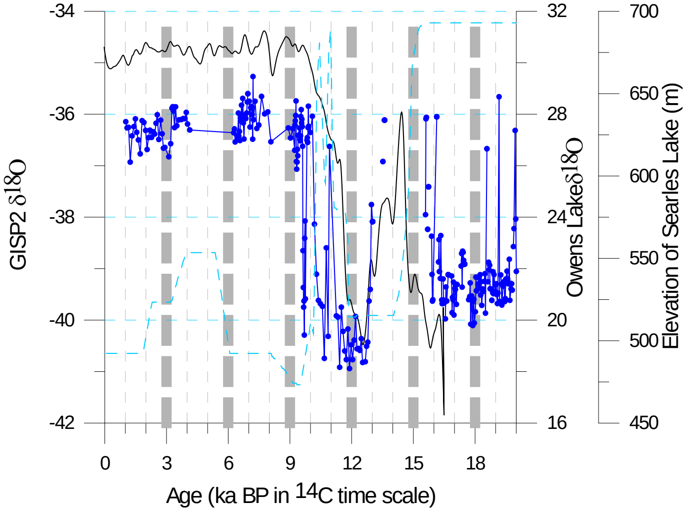

Figure 2. $\delta^{18}O$ for core OL84B (solid line with dot) (Benson et al., 1997), $\delta^{18}O$ from GISP2 (solid line) (Stuiver et al., 1995), and elevation of lake surface for Searles Lake (Smith and Street-Perrott, 1983) in the last 20 ka.

HESSD
12, 6505–6539, 2015

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀| |▶
◀ ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

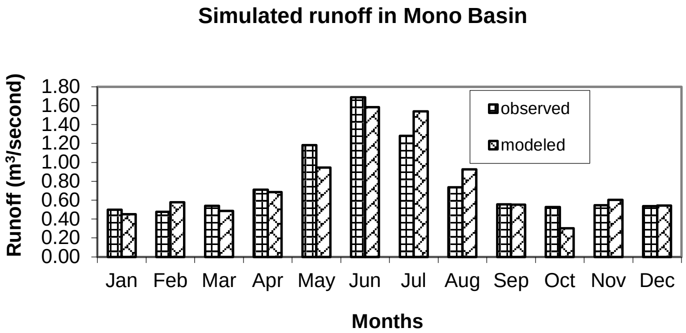

Figure 3. A comparison of observed runoff and simulated runoff in Mono Lake drainage basin.

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

HESSD

12, 6505–6539, 2015

Temperature and rainfall estimates for past 18 000 years in Owens Valley

Z. Yu et al.

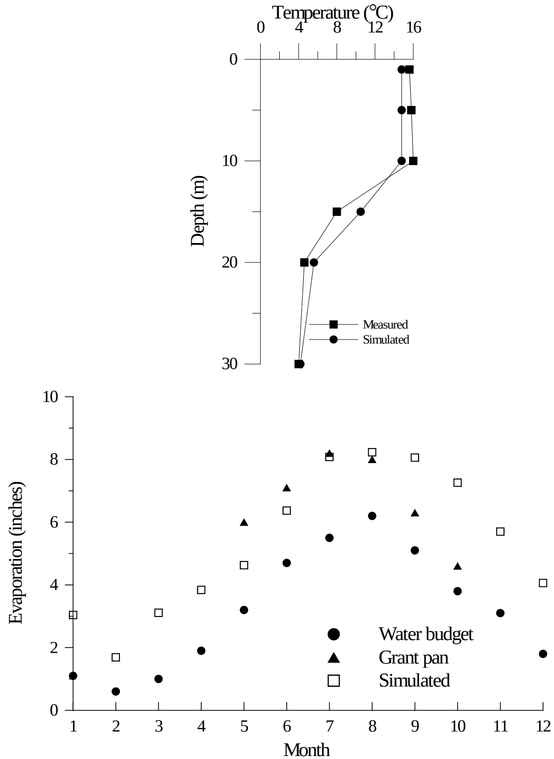

Figure 4. A plot of simulated and measured temperature profile (MacIntyre et al., 1999) in Mono Lake (upper panel); simulated evaporation and observed evaporation for Mono Lake (lower panel).

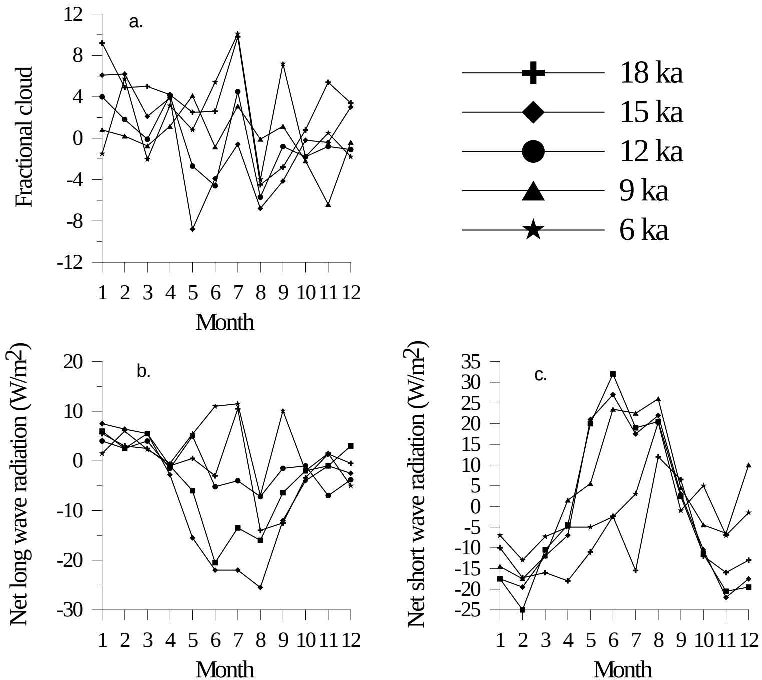

Figure 5. Inputs of cloud cover and solar radiation for the simulations at 18, 15, 12, 9, and 6 ka (Bartlein et al., 1998).

HESSD
12, 6505–6539, 2015

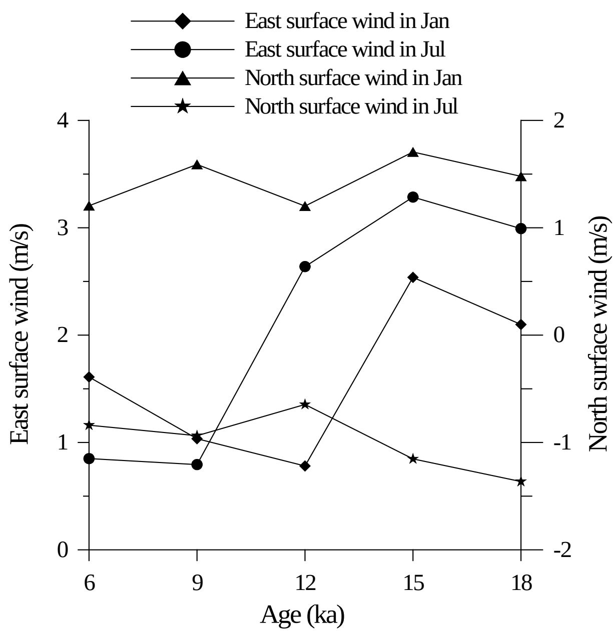

Figure 6. Inputs of wind speed for the simulations at at 18, 15, 12, 9, and 6 ka (Kutzbach and Guetter, 1986).

HESSD
12, 6505–6539, 2015

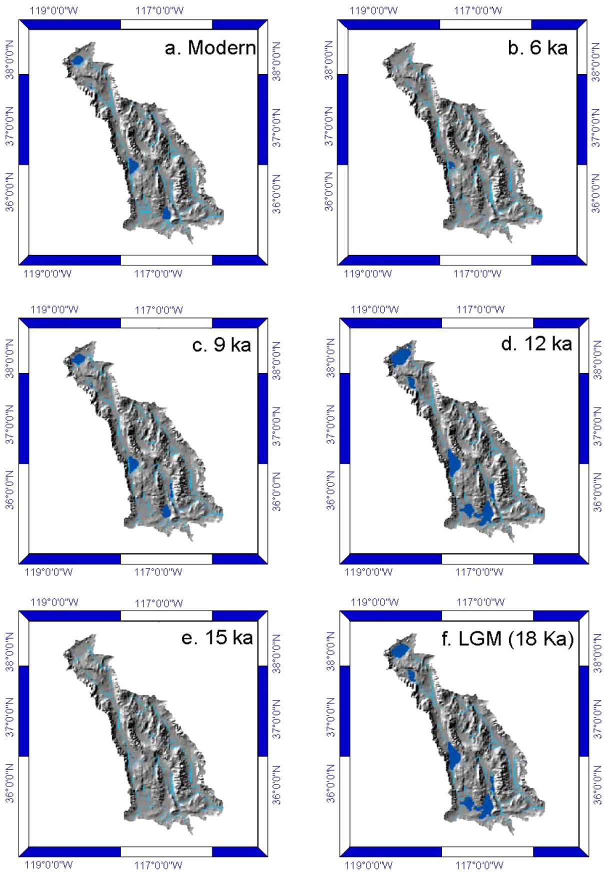

Figure 7. Simulated lake extents (deep blue) in the last 18 ka, and hillshade (gray scale) and stream network (light blue) derived from DEM data.

HESSD

12, 6505–6539, 2015

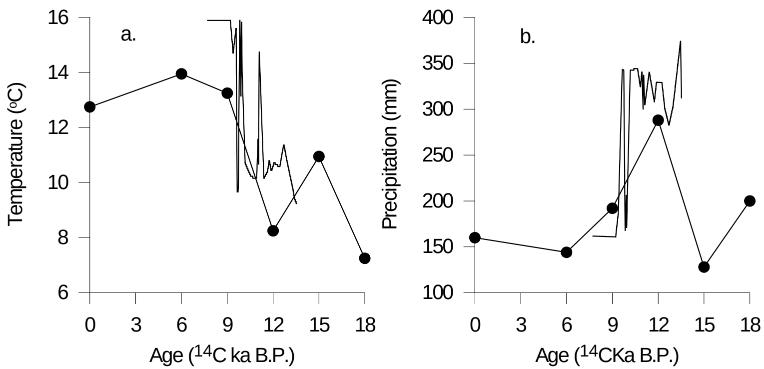

Figure 8. Simulated temperature (solid line with dots) and estimated temperature based on pollen (solid line) (Mensing, 2001) in Owens Lake (left panel) and simulated precipitation (solid line with dots) and estimated precipitation based on pollen (solid line) (Mensing, 2001) in Owens Lake (right panel).

HESSD
12, 6505–6539, 2015

Temperature and rainfall estimates for past 18 000 years in Owens Valley

Z. Yu et al.