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Stable isotopes of oxygen and hydrogen in the Truckee River–Pyramid Lake surface-water system. 2. A predictive model of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in Pyramid Lake

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Abstract

A physically based model of variations in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in Pyramid Lake is presented. For inputs, the model uses measurements of liquid water inflows and outflows and their associated isotopic compositions and a set of meteorological data (radiative fluxes, air temperature, relative humidity, and windspeed). The model simulates change of lake volume, thermal and isotopic stratification, evaporation, and the isotopic composition of evaporation. A validation of the model for 1987–1989 and 1991 indicates that it can reproduce measured intra- and interannual variations of $\delta^{18}\text{O}$ and $\delta^2\text{H}$. Three applications of the model demonstrate its ability to simulate longer term responses of $\delta^{18}\text{O}$ to change in the hydrologic balance and hydrologic characteristics (opening and closing) of the lake.

Annual variations of the isotopic ($\delta^{18}\text{O}$ and $\delta^2\text{H}$) composition of Pyramid Lake are associated by Benson (1994) with annual patterns of precipitation, evaporation, streamflow discharge, and lake mixing (stratification). Larger, long-term changes in isotopic composition are linked to change in the hydrologic balance and the hydrologic state (e.g. whether the lake is open or closed) of the lake. In this paper, we present a model used to simulate variations in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in Pyramid Lake that occur in response to change in the hydrologic balance and climate.

Several steady state and dynamic models of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ have been developed and applied to lakes (e.g. Gat 1984; Zimmerman and Eh-halt 1970; Zimmerman 1977; Fontes et al. 1979; Phillips et al. 1986). The models have been successfully used to evaluate hard to measure components of the hydrologic budget (typically evaporation). In such applications, a model is calibrated on measured hydrologic and isotopic data, and the hydrologic component of interest is estimated indirectly by calculating the residual that is required to balance the isotopic budget of the lake. Lewis (1979), for example, used a two-box representation of the epilimnion and hypolimnion to reproduce the isotopic composition and evaporation rate of Lake Kinneret. The model was calibrated with (among other measured inputs) values of water temperature and thermocline depth. Applications such as this are useful for determining components of the hydrologic budget. Because the models are calibrated with

measured data, however, they are limited to applications within the period of calibration and to lakes where sufficient measured data are available. In addition, except for the model of Lewis, a vertically homogeneous water column is assumed, and the effects of seasonal thermal and isotopic stratification on isotopic composition are not considered.

To broaden the range of applicability of isotope models, we have developed a calibration-free model for Pyramid Lake that can simulate the seasonal behavior of $\delta^{18}\text{O}$ and $\delta^2\text{H}$. Both external forcing (atmospheric and hydrologic) and internal processes (seasonal stratification and mixing) that affect the isotopic composition are treated in the model. For inputs, the model requires inflows and outflows of liquid water and their associated isotopic compositions and a set of atmospheric data (air temperature, humidity, windspeed, solar radiation, and atmospheric radiation). The model simulates evaporation, change of volume (level), and patterns of thermal and isotopic stratification of the lake. The model may be generally applicable to closed or open freshwater lakes and will be modified in the future to reconstruct past hydrologic change from isotopic variations evidenced in the sedimentary record of the Lahontan basin.

In this paper, we describe the model and then validate it. We present three applications that demonstrate the potential of the model for deriving hydrologic information from past variations of isotopic composition. The first application is a simulation of the 1986–1991

change in monthly values of $\delta^{18}\text{O}$ that occurred in response to historical extreme variations in streamflow discharge into Pyramid Lake. The second application is an equilibrium simulation (300 yr) in which the model was run on a daily time step to demonstrate the dependence of isotopic equilibrium on the kinetics of hydrologic equilibrium. In the third application, the model is run on an annual time step to demonstrate the isotopic response of the lake as it changes from a closed basin to an open basin and then back to a closed basin.

Model description

Model equations—The hydrologic and isotopic balances of the lake are computed in the model from the standard mass-balance equation (Gonfiantini 1965):

$$\frac{\partial(\delta V)}{\partial t} = (P\delta_p - E\delta_e)A + I\delta_{si} - O\delta_{so} + G_i\delta_{gi} - G_o\delta_{go}. \quad (1)$$

V is lake volume, δ represents either $\delta^{18}\text{O}$ or $\delta^2\text{H}$, t is time, P is on-lake precipitation, δ_p is isotopic composition of the precipitation, E is lake evaporation, δ_e is isotopic composition of the evaporative vapor, A is surface area of the lake, I is surficial inflow to the lake, δ_{si} is isotopic composition of the inflow, O is surficial outflow from the lake, δ_{so} is isotopic composition of the outflow, G_i is groundwater inflow into the lake, δ_{gi} is isotopic composition of the groundwater inflow, G_o is groundwater outflow from the lake, and δ_{go} is isotopic composition of the groundwater outflow (units are given in the list of notation). The isotopic compositions of the liquid-water terms in Eq. 1 are readily determined by analysis, and the composition of the evaporative flux is evaluated here by modeling.

For daily time steps, our isotope model makes use of an existing one-dimensional model (Hostetler and Bartlein 1990) that simulates the thermal structure and evaporation of a lake. The equation of the thermal model is

$$\frac{\partial T}{\partial t} = \frac{1}{A(z)} \frac{\partial}{\partial z} \left\{ A(z) [k_h + K_h(z, t)] \frac{\partial T}{\partial z} \right\} + \frac{1}{C A(z)} \frac{\partial A(z) \Phi}{\partial z}. \quad (2)$$

Notation

A	Lake surface area, m^2
C	Heat capacity of water, $\text{J (kg } ^\circ\text{C)}^{-1}$
E	Evaporation, m t^{-1}
I, O	Surface inflow and outflow, $\text{m}^3 \text{t}^{-1}$
G_i, G_o	Groundwater inflow and outflow, $\text{m}^3 \text{t}^{-1}$
K_h	Vertical eddy diffusivity, $\text{m}^2 \text{d}^{-1}$
K_z	Vertical eddy diffusion of ξ , $\text{m}^2 \text{d}^{-1}$
N_{mt}	Mass-transfer coefficient, kPa^{-1}
P	Precipitation, m t^{-1}
RH	Relative humidity, dimensionless
T	Lake temperature, $^\circ\text{C}$
T_w	Temperature of the water surface, $^\circ\text{K}$
U_2	2-m windspeed, m s^{-1}
V	Lake volume, m^3
e_a, e_o	Atmospheric air and surface-saturated vapor pressure, kPa
f	Fraction advected air, dimensionless
k_h	Molecular diffusivity, $\text{m}^2 \text{d}^{-1}$
$k_{m\xi}$	Molecular diffusion of ξ , $\text{m}^2 \text{d}^{-1}$
q_{le}, q_h	Latent and sensible heat flux, W m^{-2}
t	Time, d
z	Lake depth, m
Φ	Heat source term, W m^{-2}
α_{eq}	Isotopic enrichment factor, ‰
α_k	Ratio of isotopic diffusivities, dimensionless
α_{sw}	Shortwave albedo, dimensionless
δ_a, δ_e	Isotopic composition of advected air and evaporative vapor, ‰
δ_{gi}, δ_{go}	Isotopic composition of groundwater inflow and outflow, ‰
δ_l, δ_p	Isotopic composition of lake water and precipitation, ‰
δ_{si}, δ_{so}	Isotopic composition of surface inflow and outflow, ‰
ϕ_i, ϕ_s	Incoming atmospheric and solar radiation, W m^{-2}
ϕ_o	Outgoing longwave radiation, W m^{-2}
ξ	$\delta^{18}\text{O}$ or $\delta^2\text{H}$, ‰

T is lake temperature, z is lake depth, $A(z)$ is lake area at depth z (a vertical resolution of 1 m is used here), k_h is molecular diffusivity, $K_h(z, t)$ is vertical eddy diffusivity, C is the heat capacity of water, and Φ is a heat-source term representing subsurface absorption of solar radiation. The surface boundary condition for Eq. 2 is

$$[k_h + K_h(z, t)] \frac{\partial T}{\partial z} = (1 - \alpha_{sw})\phi_s + \phi_l + \phi_o \pm q_{le} \pm q_h. \quad (3)$$

α_{sw} is the shortwave albedo of the lake surface, ϕ_s is incoming solar radiation, ϕ_l is incoming atmospheric radiation, ϕ_o is outgoing longwave radiation from the lake surface, q_{le} is latent heat flux, and q_h is sensible heat flux. For

Pyramid Lake, we assume that no heat is transferred between lake water and underlying sediments which leads to the bottom boundary condition of $[k_m + K_h(z, t)]\partial T/\partial z = 0$. A measured or estimated temperature-depth profile provides the initial conditions for Eq. 2.

Bulk evaporation is computed from mean atmospheric conditions by a mass-transfer equation (Brutsaert 1982):

$$E = N_{mt}U_2(e_o - e_a). \quad (4)$$

N_{mt} is a mass-transfer coefficient determined as a function of lake surface area, U_2 is the 2-m windspeed, e_o the saturated vapor pressure at the lake surface, and e_a the air-vapor pressure of the atmosphere.

Preferential removal (fractionation) of the lighter molecule of water ($^1\text{H}_2^{16}\text{O}$) with respect to the heavier molecules of water ($^1\text{H}^2\text{H}^{16}\text{O}$ and $^1\text{H}_2^{18}\text{O}$) occurs during evaporation (e.g. Gonfiantini 1986). The degree of fractionation is determined by the temperature and isotopic composition of the water, windspeed, and the relative humidity and isotopic composition of the atmospheric boundary layer (Craig and Gordon 1965; Merlivat 1978; Gat 1979; Merlivat and Jouzel 1979). Within the boundary layer, evaporation—and thus isotopic fractionation—is controlled by near-surface conditions (e.g. surface roughness, turbulence) that vary spatially over the lake (e.g. Brutsaert 1975a,b; Merlivat 1978). In the model, evaporation is computed from mean boundary-layer conditions around the lake with a mass-transfer technique (Eq. 4). We address the effects of the isotopic composition of the boundary air in the boundary layer by comparing fractionation δ_e with a modified version of the equation derived by Craig and Gordon (1965):

$$\delta_e = \left[\frac{\delta_l}{\alpha_{eq}} - \text{RH}f\delta_a \right] \cdot \left[\frac{1 - \text{RH}}{\alpha_k} + \text{RH}(1 - f) \right]^{-1}. \quad (5)$$

δ_l is the isotopic composition of the lake water, α_{eq} the isotopic enrichment factor, RH the relative humidity, f the fraction of the air in the boundary layer over the lake that is derived (advected) from outside the basin, δ_a the isotopic composition of the advected air, and α_k (0.994 for $\delta^{18}\text{O}$ and 0.995 for $\delta^2\text{H}$) the ratio of

isotopic diffusivities in the boundary layer (Merlivat and Jouzel 1979).

The isotopic enrichment factor α_{eq} varies with isotopic species and is a function of lake temperature (Majoube 1971). For $\delta^{18}\text{O}$,

$$\alpha_{eq} = \exp(1,137T_w^{-2} - 0.4156T_w^{-1} - 0.00207); \quad (6a)$$

for $\delta^2\text{H}$,

$$\alpha_{eq} = \exp(24,844T_w^{-2} - 76.248T_w^{-1} + 0.05261). \quad (6b)$$

T_w is the temperature of the water surface ($^{\circ}\text{K}$).

Differential equations of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ transport are coupled with the thermal model to account for seasonal stratification and mixing of isotopes in the water column. The generalized isotopic transport equation is

$$\frac{\partial \xi}{\partial t} = \frac{1}{A(z)} \frac{\partial}{\partial z} \left\{ A(z)[k_{m\xi} + K_\xi(z, t)] \frac{\partial \xi}{\partial z} \right\}. \quad (7)$$

ξ represents either $\delta^{18}\text{O}$ or $\delta^2\text{H}$, $k_{m\xi}$ is the molecular diffusion of ξ , and K_ξ is the vertical eddy diffusion of ξ . The boundary condition for Eq. 7 is

$$[k_{m\xi} + K_\xi(z, t)] \frac{\partial \xi}{\partial z} = \text{rhs}. \quad (8)$$

rhs is the right-hand side of Eq. 1. The initial conditions for Eq. 7 are provided by measured or estimated isotope-depth profiles.

Modeling assumptions—The thermal model has been validated and used successfully to predict thermal structure and evaporation in various studies (e.g. Hostetler and Benson 1990; Hostetler and Giorgi 1993; Hostetler et al. 1993) that demonstrate the applicability of the one-dimensional model to lakes with different morphometries and climatic settings. Both wind-driven turbulent mixing (eddy diffusion) and density-driven convective mixing are simulated by the model. The diffusion of heat by eddies and convective mixing both entail the movement of water molecules. We assume that transport of the isotopes of oxygen and deuterium is also accomplished by eddies and convection without preference for isotopic species (e.g. ^{18}O vs. ^{16}O). In the model, therefore, K_ξ is set equal to the computed value of K_h , and the simulated convection that mixes

heat into the lake is also used to mix the isotopes.

The depth at which river discharge flows into a lake is determined by the morphometry of the lake and by the difference in density of river discharge and lake water. The dynamics of river inflow are not explicitly treated in this model; rather, all inflow is routed into the mixed layer. This assumption is generally appropriate for existing conditions at Pyramid Lake (and for past meromictic events).

Model validation and application

The isotope model was validated for two time periods: 12 October 1987 through 12 October 1989 and 8 May through 26 December 1991. Because Pyramid Lake currently has no outlet, Eq. 1 and 5 were simplified by eliminating the surficial outflow and groundwater terms.

Input data—Daily discharge data for the Truckee River at Nixon were obtained from U.S. Geological Survey data (1986–1992). Isotopic values of the discharge were interpolated with cubic splines fitted to measured isotopic data for Farad, Nevada (Benson 1994). Isotopic values for on-lake precipitation were obtained from datasets also reported by Benson (1994). The meteorological datasets were obtained from measurements made at meteorological stations on the shore of Pyramid Lake (Hostetler and Benson 1993). The initial temperature profiles for the model were obtained from a thermistor string affixed to a raft located near the center of the lake (Hostetler and Benson 1993), and the initial profiles of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ were estimated from analyses of water samples (Benson 1994) that also were obtained at the raft site.

Results—Because the lake is closed, its isotopic composition is strongly influenced by evaporative fractionation which is determined, in part, by characteristics of the water vapor contained in the air mass over the lake (Eq. 5). In our model, these characteristics are estimated by specifying the fraction of advected air—and thus the fraction of lake-derived air—in the air mass. A series of simulations was conducted with different values of the parameter f in Eq. 5 to validate the model and to investigate the sensitivity of the model to the make-up of the air mass. For all simulations, the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of the water

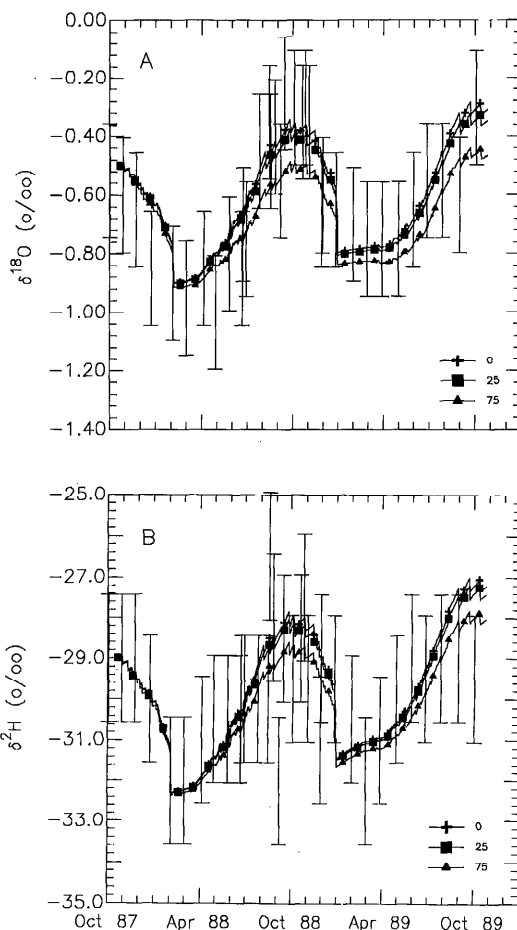


Fig. 1. Simulated epilimnetic (—) isotopic values for different values of the parameter f in Eq. 5 compared with measured values from Pyramid Lake. A. Measured values are plotted with 0.2‰ error bars. B. Measured values are plotted with 2.0‰ error bars. Measured data from Benson (1994).

vapor in the advected air were specified as -21‰ and -151‰ , respectively. These values are consistent with isotopic analyses of water-vapor samples obtained from the boundary layer above the lake (L. V. Benson unpubl. data).

For 1987–1989, the overall annual cycles of measured epilimnetic $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in the lake are captured by the model for a mixture of up to 75% advected air ($f = 0.75$) in the air mass above the lake (Fig. 1). If $>25\%$ of the air mass is derived from advected air, the model yields epilimnetic $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values that are more depleted than measured values. If the air mass above the lake contains 25% or less advected

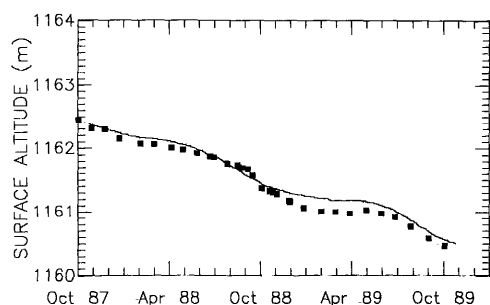


Fig. 2. Simulated (—) and measured (■) 1987–1989 surface-water levels of Pyramid Lake. Measured data from the U.S. Geological Survey (1987–1989).

air, the simulated epilimnetic $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values fall within the error of the measured values (Fig. 1) and follow the general trends in the measured data to more enriched values. These trends correspond to the decrease of lake level that was correctly simulated by the model (Fig. 2). These sensitivity tests indicate that for 1987–1989 the isotopic characteristics of the water vapor in the air mass above the lake were primarily derived from lake water and not from advected air. Based on these results, we used an air mass consisting of 10% advected air for the remaining simulations.

During 1987–1989, two periods of turnover—one at the end of January 1988 and another at the beginning of January 1989—are simulated by the model (Fig. 3). Profiles of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ are not available for this period; however, the simulated time of turnover is consistent with the lightening of the isotopes associated with the mixing of the epilimnion with more depleted hypolimnetic waters. The timing of these mixing events is also consistent with measured thermal profiles at the raft located in the center of the lake (Hostetler and Benson 1990; Pyramid Lake Fish. unpubl. data). The turnover periods are further illustrated in Fig. 4, which shows modeled $\delta^{18}\text{O}$ stratification and progressive entrainment of depleted hypolimnetic water into the epilimnion preceding turnover.

For 1991, the simulated time series of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ also fall within the error of the measured isotopic composition of the lake (Fig. 5). Turnover of the lake is not simulated (and was not observed) during the period considered; however, the model in general is able to reproduce the measured change of $\delta^{18}\text{O}$ in the

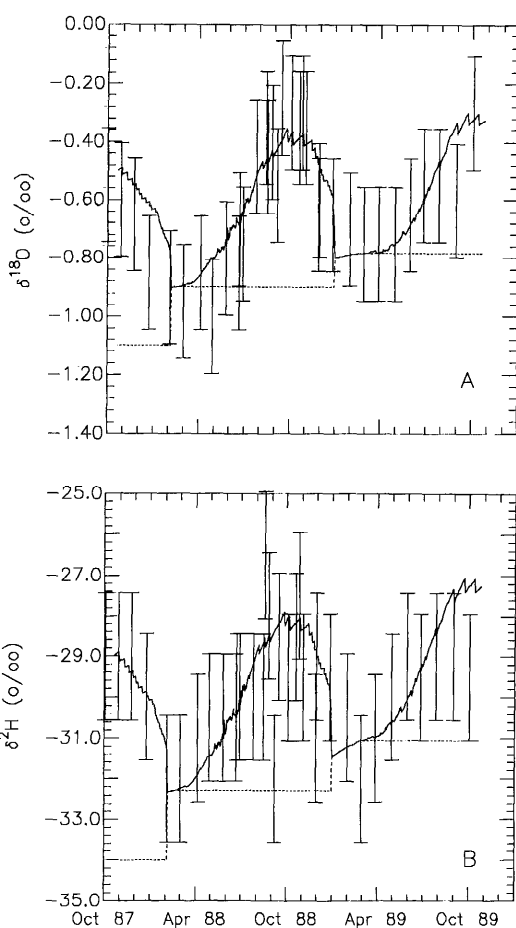


Fig. 3. Simulated epilimnetic (—) and hypolimnetic (---) isotopic values for $f = 0.10$ in Eq. 5 compared with measured 1987–1989 values from Pyramid Lake. A. Measured values are plotted with 0.2‰ error bars. B. Measured values are plotted with 2.0‰ error bars. Measured data from Benson (1994).

water column during erosion of the thermocline in fall (Fig. 6).

The results of the validation indicate that the model is capable of simulating the intra- and interannual variation of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in Pyramid Lake. Discrepancies between the simulated and measured isotopic values could be the result of bias in the measured data (i.e. near-surface samples could have been obtained that were not representative of the average composition of the epilimnion, Benson 1994). For example, the most enriched $\delta^{18}\text{O}$ values are found in 1991 samples obtained from the surface at the edge of the lake, where evaporation may have been most intense. Un-

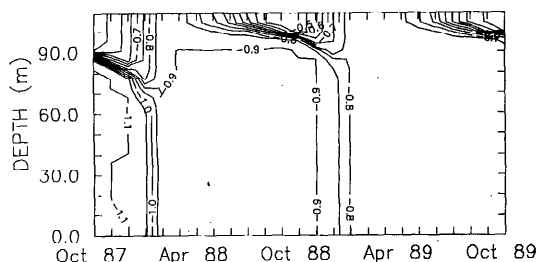


Fig. 4. Simulated 1987–1989 evolution of $\delta^{18}\text{O}$ in Pyramid Lake. Contour interval of 0.05‰.

derestimates in simulated evaporation rates, also would introduce error in that the simulated isotopic composition of the epilimnion would be more depleted than the actual composition.

Model applications—In the first application, the model is used to reproduce the large ($> 2\text{‰}$) $\delta^{18}\text{O}$ variation in Pyramid Lake over the period 1986–1991 (Benson 1994). The intent of this simulation is not to reproduce exactly the measured values of $\delta^{18}\text{O}$ for the 6-yr period, but rather to demonstrate the ability of the model to simulate large changes and general trends in $\delta^{18}\text{O}$ that result from historical change in the hydrologic budget of the lake.

The 6-yr simulation was run on monthly time steps with the mass-balance model of Eq. 1. The amounts and compositions of river discharge and precipitation were obtained from Benson (1994). Monthly inputs of water temperature, thermocline depth, evaporation, and relative humidity were derived by averaging the output from the 1987–1989 daily simulations. The model is able to reproduce the rapid depletion in $\delta^{18}\text{O}$ (2‰) in 1986 (Fig. 7) that was a result of abnormally high discharge from the Truckee River (1.6 times normal) and the subsequent return to enriched values caused by evaporation that followed the onset of the current drought in the Truckee basin. The model also reproduces the general trend toward enriched composition from 1986 to 1992.

In the second application, the model is used to simulate the long-term (300 yr) response of $\delta^{18}\text{O}$ as a lake, given prescribed hydrologic and meteorologic conditions, tends to a pseudo-equilibrium level (volume). This application demonstrates that the model is stable for long simulations and that it simulates the dependence of isotopic equilibrium on the kinetics of hydrologic equilibrium.

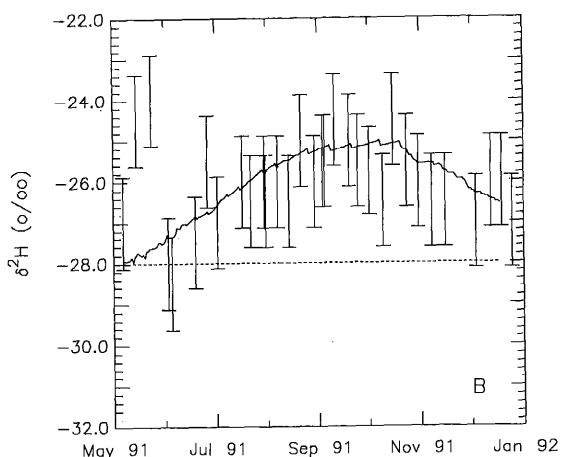
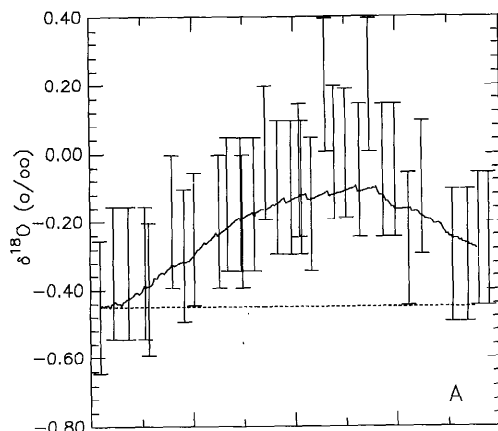


Fig. 5. As Fig. 3, but compared with measured 1991 values.

The daily model was run for the 300-yr simulation with perpetual 1987–1988 meteorologic, hydrologic, and isotopic data as input. Under perpetual 1987–1988 hydrologic conditions (streamflow discharge 0.04 km^3 , a value $\sim 10\%$ of the long-term mean discharge of $0.38 \text{ km}^3 \text{ yr}^{-1}$), the lake would fall rapidly at the evaporation rate of $\sim 1.20 \text{ m yr}^{-1}$ and dry up in $\sim 100 \text{ yr}$. Similarly, if perpetual mean conditions were invoked, the lake would also fall, although more slowly, to very low levels. For the 300-yr model run, therefore, a water balance (streamflow discharge of $0.5 \text{ km}^3 \text{ yr}^{-1}$) was prescribed that would allow the level of the lake to fall and then stabilize. The model was initialized at the 1987 level.

For the 300-yr simulation, annual cycles (high-frequency, low-amplitude variations) in

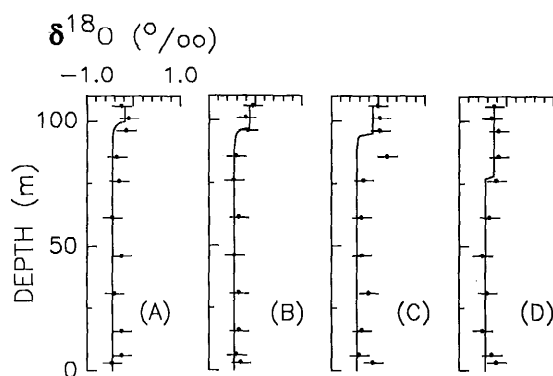


Fig. 6. Simulated (—) profiles of $\delta^{18}\text{O}$ compared with measured (●) 1991 values in Pyramid Lake: 1 August—A; 13 September—B; 10 October—C; 10 December—D. Measured values are plotted with 2.0‰ error bars. Measured data from Benson (1994).

the $\delta^{18}\text{O}$ composition of the epilimnion are superimposed on the underlying asymptotic trend toward equilibrium (Fig. 8). The trend of $\delta^{18}\text{O}$, which follows the simulated decrease and subsequent trend to equilibrium of lake level (Fig. 9), indicates the effect of the hydrologic balance on the isotopic composition of the lake. The trend toward an equilibrium value of $\sim 1.4\text{‰}$ is caused by increased evaporative fractionation, which is a result of longer residence time of water in the lake.

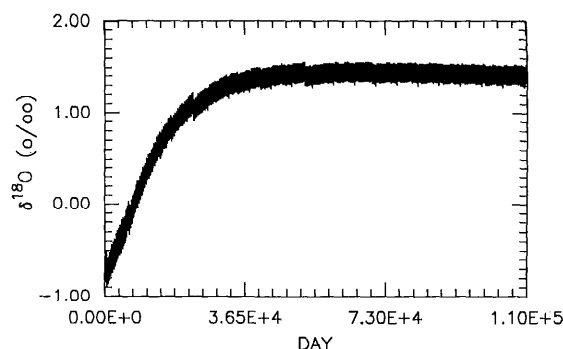


Fig. 8. Daily $\delta^{18}\text{O}$ values for Pyramid Lake for 300 simulated years. Each short-term variation is one annual cycle. Daily data plotted every 30 d of simulation. ($\delta_{si} \approx -10.5\text{‰}$; $\delta_e \approx -12\text{‰}$.)

When a lake changes from a closed to an open state, the residence time of the water decreases, and the lake becomes isotopically depleted as a result of reduced exposure to evaporation. Conversely, residence time increases and a lake becomes isotopically enriched through evaporation when it changes from an open to a closed state. In the third application, the model is used to illustrate the effects on the $\delta^{18}\text{O}$ of lake water as a closed lake opens and spills and then becomes closed again.

The Lahontan basin comprises seven sub-basins that are separated hydrologically by sills (Benson and Mifflin 1986). Under favorable

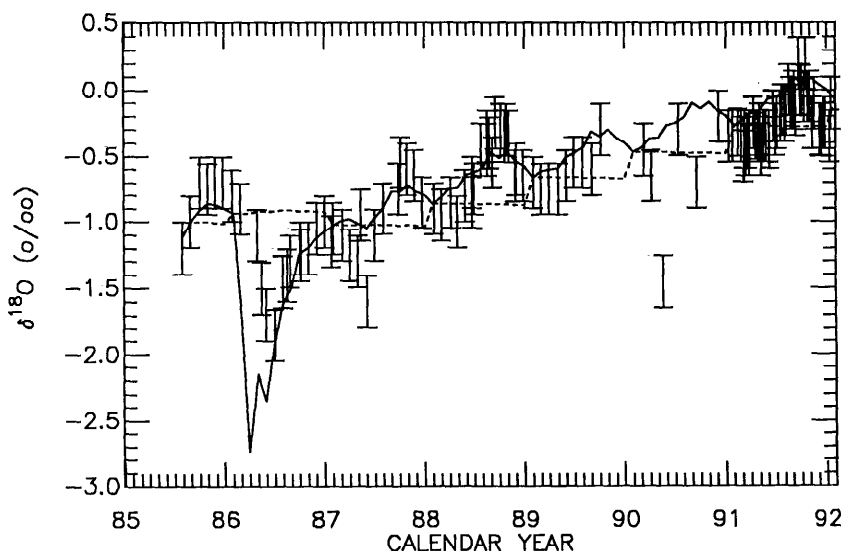


Fig. 7. Simulated epilimnetic (—) and hypolimnetic (---) $\delta^{18}\text{O}$ values for Pyramid Lake compared with 1986–1991 measured values. Measured data are plotted with 0.2‰ error bars. Measured data from Benson (1994).

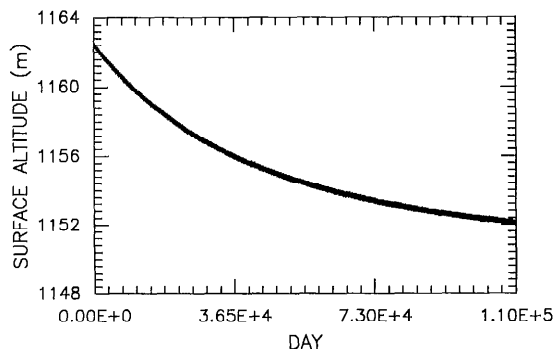


Fig. 9. As Fig. 8, but of lake level.

hydrologic budgets, the lakes fill and spill over the sills to adjoining basins. The model of Eq. 1 was incorporated into a hydrologic routing model of the Lahontan basin (that simulates spill of water from one subbasin to another based on the hydrologic balance and sill configuration of each subbasin) to illustrate the hydrologic and isotopic responses of Pyramid Lake filling and spilling into Winnemucca Lake. We used a streamflow discharge of $1.45 \text{ km}^3 \text{ yr}^{-1}$ (~ 4 times the average discharge of the Truckee River into the lake) with an isotopic value of -10.5‰ and 1987–1989 annual averages for lake temperature, evaporation, thermocline depth, and relative humidity as input to the model.

Pyramid Lake initially becomes isotopically depleted (Fig. 10) in response to the persistent input of isotopically depleted river water that also causes lake level to rise (Fig. 11). When Pyramid Lake reaches surface level of 1,177 m, the lake spills to the Winnemucca Lake

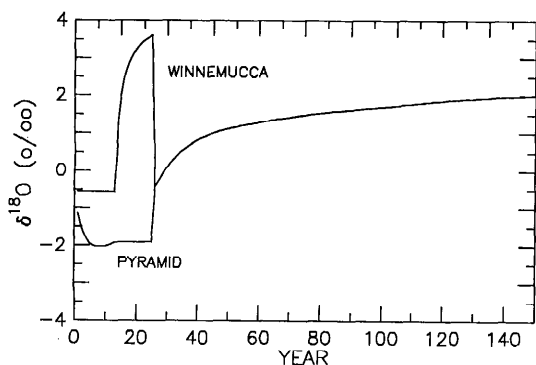


Fig. 10. Simulated isotopic composition of Pyramid and Winnemucca Lakes. ($\delta_{\text{r}} \approx -10.5\text{‰}$; $\delta_{\text{c}} \approx -12.5\text{‰}$.)

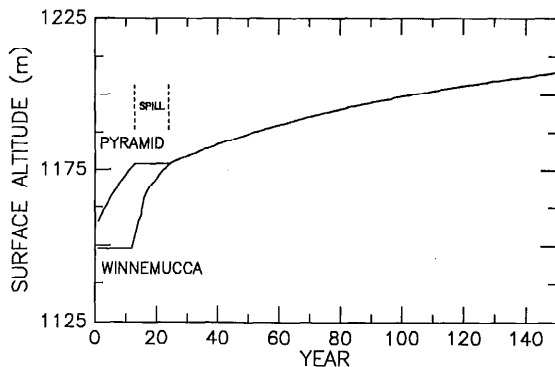


Fig. 11. Simulated lake level of Pyramid and Winnemucca Lakes. Spill indicates the period when Pyramid Lake spills into the Winnemucca Lake basin.

basin, and depletion continues as a result of the reduced residence time of the lake water in the Pyramid Lake subbasin. Isotopic values for Winnemucca Lake water are relatively enriched initially due to the effects of evaporation on the small volume of water contained in the basin. After 25 yr, Winnemucca Lake rises to sill level and the lakes join. The large initial responses of $\delta^{18}\text{O}$ when the lakes join are from mixing the two lakes (the initial isotopic composition of the combined lake is computed in the model as the volume-weighted average composition of the individual lakes). The combined lake, which is now closed, rises toward a new equilibrium level and equilibrium $\delta^{18}\text{O}$ value of -2‰ , which reflects the effect of hydrologic equilibrium and long residence time of the water in the lake.

References

- BENSON, L. V. 1994. Stable isotopes of oxygen and hydrogen in the Truckee River–Pyramid Lake surface-water system, California and Nevada. 1. Data analysis and extraction of paleoclimatic information. *Limnol. Oceanogr.* **39**: 344–355.
- , AND M. D. MIFFLIN. 1986. Reconnaissance bathymetry of basins occupied by Pleistocene Lake Lahontan, Nevada and California. U.S. Geol. Surv. Water-Resour. Invest. Rep. 85-4262. 14 p.
- BRUTSAERT, W. A. 1975a. Theory for local evaporation (or heat transfer) from rough and smooth surfaces at ground level. *Water Resour. Res.* **11**: 534–550.
- , 1975b. The roughness length for water vapor, sensible heat and other scalars. *J. Atmos. Sci.* **32**: 2028–2031.
- , 1982. Evaporation into the atmosphere. Reidel.
- CRAIG, H., AND L. I. GORDON. 1965. Isotopic oceanography: Deuterium and oxygen 18 variations in the ocean and marine atmosphere, p. 277–374. *In* D. R.

- Schink and J. T. Corless [eds.], Marine geochemistry. Univ. R.I. Occas. Publ. 3.
- FONTES, J. C., B. BOULANGE, J. P. CARMOUZE, AND T. FLORKOWSKI. 1979. Preliminary oxygen-18 and deuterium study of the dynamics of Lake Titicaca, p. 203-221. *In* Isotopes in hydrology. IAEA.
- GAT, J. R. 1979. Environmental isotope balance of Lake Tiberias, p. 109-127. *In* Isotopes in hydrology. IAEA.
- . 1984. The stable isotope composition of Dead Sea waters. *Earth Planet. Sci. Lett.* **71**: 361-376.
- GONFIANTINI, R. 1965. Isotopic effects in the evaporation of salt water [in Italian]. *Atti Soc. Toscana Nat. Sci. Pisa* **72**: 5-22.
- . 1986. Environmental isotopes in lake studies, p. 113-168. *In* P. Fritz and J. C. Fontes [eds.], Handbook of environmental isotope geochemistry. V. 2. Elsevier.
- HOSTETLER, S. W., AND P. J. BARTLEIN. 1990. Simulation of lake evaporation with application to modeling lake level variations of Harney-Malheur Lake, Oregon. *Water Resour. Res.* **26**: 2603-2612.
- , G. T. BATES, AND F. GIORGI. 1993. Coupling of a lake model with a regional model of climate. *J. Geophys. Res.* **98**: 5045-5058.
- , AND L. V. BENSON. 1990. Paleoclimatic implications of the high stand of Lake Lahontan derived from models of evaporation and lake level. *Clim. Dyn.* **4**: 207-217.
- , AND ———. 1993. Meteorological and water-temperature data for Pyramid Lake, Nevada 1987-89. U.S. Geol. Surv. Open-File Rep. 92-159. 15 p.
- , AND F. GIORGI. 1993. One-way coupling of landscape-scale hydrologic models with a regional climate model: An assessment. *Water Resour. Res.* **29**: 1685-1695.
- LEWIS, S. 1979. Environmental isotope balance of Lake Kinneret as a tool in evaporation rate estimation, p. 33-53. *In* Isotopes in hydrology. IAEA.
- MAJOUBE, M. 1971. Fractionnement en oxygene-18 et en deuterium entre l'eau et sa vapeur. *J. Chim. Phys.* **197**: 1423-1436.
- MERLIVAT, L. 1978. The dependence of bulk evaporation coefficients on air-water interfacial conditions as determined by the isotopic method. *J. Geophys. Res.* **83**: 2977-2980.
- , AND J. JOUZEL. 1979. Global climatic interpretation of the deuterium-oxygen 18 relationship for precipitation. *J. Geophys. Res.* **84**: 5029-5033.
- PHILLIPS, F. M., M. A. PERSON, AND A. B. MULLER. 1986. A numerical lumped parameter model for simulating the isotopic evolution of closed-basin lakes. *J. Hydrol.* **85**: 73-86.
- U.S. GEOLOGICAL SURVEY. 1986-1992. Water resources data for Nevada. U.S. Geol. Surv. Water-Data Rep. Ser. Annu. Vol.
- ZIMMERMAN, U. 1977. Determination by stable isotopes of underground inflow and outflow and evaporation of young artificial groundwater lakes, p. 87-94. *In* Isotopes in lake studies. IAEA.
- , AND D. H. EHHALT. 1970. Stable isotopes in study of water balance of Lake Neusiedel, Austria, p. 129-138. *In* Isotope hydrology. Proc. Int. Conf. IAEA.

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