1994

Stable isotopes of oxygen and hydrogen in the Truckee River-Pyramid Lake surface-water system. 1. Data analysis and extraction of paleoclimatic information

Larry Benson
U.S. Geological Survey, great.basin666@gmail.com

Limnol. Oceanogr., 39(2), 1994, 344–355

Stable isotopes of oxygen and hydrogen in the Truckee River–Pyramid Lake surface-water system. 1. Data analysis and extraction of paleoclimatic information

L. V. Benson
U.S. Geological Survey, 3215 Marine St., Boulder, Colorado 80303

Abstract

The $\delta^{18}O$ content of streamflow discharge entering Pyramid Lake is a simple mixture of isotopically enriched Lake Tahoe discharge and isotopically depleted snowmelt. The $\delta^{18}O$ value of Pyramid Lake water varies continuously, as isotopically depleted water evaporates from the epilimnion and isotopically depleted water enters the lake as streamflow discharge and on-lake precipitation. In a normal water year, the $\delta^{18}O$ of Pyramid Lake surface water varies seasonally because the components of the hydrologic balance in the Pyramid Lake system are seasonally shifted. The problem of isotopic and thermal heterogeneity can be minimized by analyzing carbonates that integrate the annual variance in $\delta^{18}O$, form in an environment in which temperature is relatively invariant, and are not subject to recrystallization after deposition.

Application of the oxygen-isotope compositions of authigenic lake carbonates to studies of past climates has become common practice since the seminal work of Stuiver and his colleagues (Stuiver 1968, 1970; Covich and Stuiver 1974). Variations in carbonate $\delta^{18}O$ generally have been attributed to several processes, including variation in the isotopic composition of precipitation reaching the lake as rain or streamflow discharge (Stuiver 1968, 1970; Fritz et al. 1975). Because the $\delta^{18}O$ of precipitation is highly correlated with condensation air temperature (Yurtsever 1975), change in the $\delta^{18}O$ of precipitation reaching a lake has often been ascribed to change in air temperature (Eicher and Siegenthaler 1976; Eicher 1980), i.e. $d\delta^{18}O/dT \sim 0.7$‰ per °C (Dansgaard 1964). Variation in carbonate $\delta^{18}O$ has been attributed to variation in the temperature of carbonate precipitation (Craig 1965), variation in the residence time of water in a lake basin (Craig 1961), and variation in the relative rates of evaporation and inflow in both closed- and open-basin lake systems (Craig 1961; Covich and Stuiver 1974; Fritz et al. 1975).

In many paleoclimatic studies, it has been assumed that the mean $\delta^{18}O$ of lake water is recorded within a carbonate precipitate. This assumption is valid only if a carbonate sample integrates temporal and spatial variations in water temperature and $\delta^{18}O$. For example, carbonates deposited in the sediments of closed-basin lakes from the Great Basin of the western U.S. often form over short timespans in lakes that are thermally and isotopically heterogeneous (e.g. Bischoff et al. 1991; Galat and Jacobsen 1985).

The purposes of this study are to document the behavior of $\delta^{18}O$ and $\delta^{2}H$ in the Truckee River surface-water system of California and Nevada (Fig. 1), determine how the history of stable isotope variation in lake water is transferred to the sedimentary record, and provide data to validate a numerical model used to simulate the temporal and spatial behavior of $\delta^{18}O$ and $\delta^{2}H$ (Hostetler and Benson 1994).

Climatology

Cool-season precipitation from North Pacific sources is dominant throughout the central Sierra Nevada (Fig. 2). From November to April, successive low-pressure centers follow the westerlies from the North Pacific across the northern Great Basin (Houghton et al. 1975). The development of the thermally induced Great Basin High throughout much of winter forces storm tracks north of 39°N over

Acknowledgments

My appreciation to those who assisted in sampling, including Harold Klieforth, James Ashby, Martin Lebo, Dan Mosley, Jim Davis, Nancy Vucinich, Carman Toby, Rick Warner, Doug Hutchinson, Ron Collins, and Robert Bostic. Jim Ashby and Harold Klieforth also contributed historical weather data for various sites in Nevada and California. Paul Wagner provided logistical support for much of the work in the Pyramid Lake subbasin. I also thank Richard Marzolf and Steve Hostetler for comments on earlier versions of the manuscript.

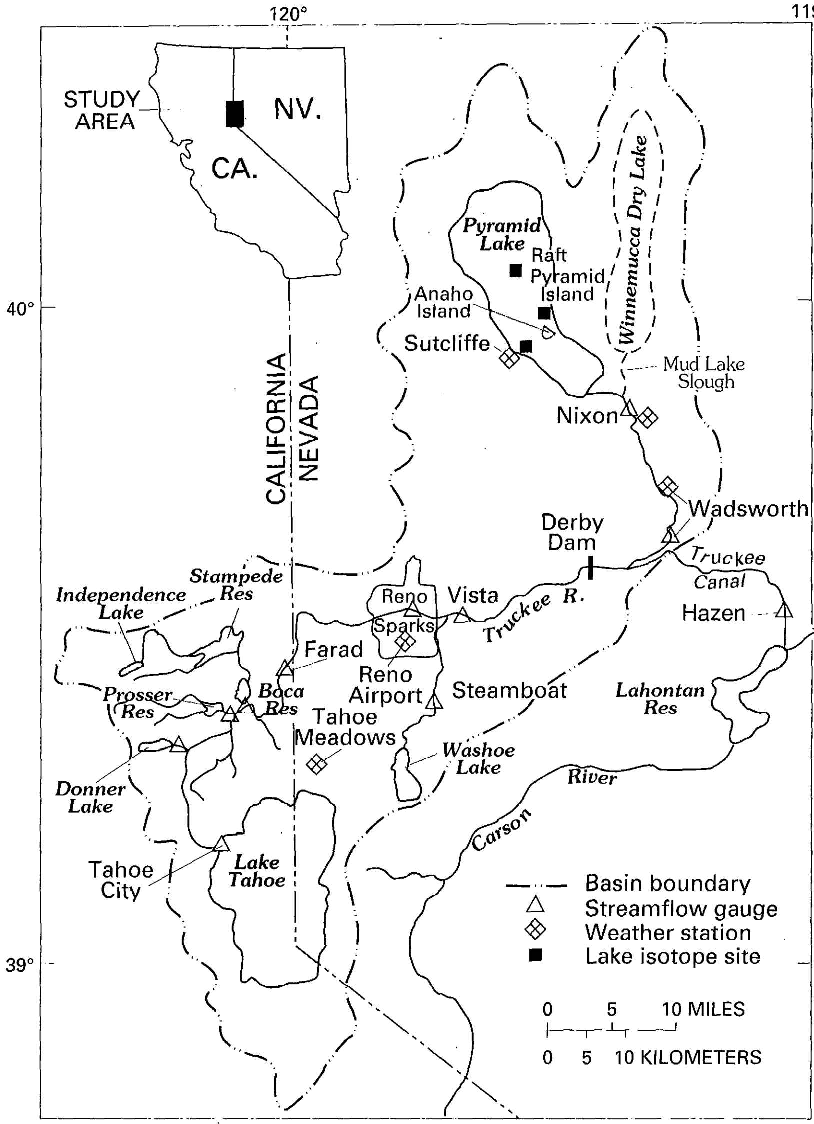

Fig. 1. Map of the Truckee River system, Pyramid Lake, and Winnemucca Dry Lake subbasins indicating locations of streamflow-gauges, weather stations, reservoirs, lakes, and isotope sampling sites.

the headwaters of the Truckee River. The air-parcel trajectories followed by these winter storms are referred to as Maritime Polar.

During July and August, the westerlies weaken and Pacific storm tracks move far north of the central Sierra Nevada. During this warm season, the study area receives only a small amount of moisture (Fig. 2) in the form of convective storms that bring moisture from the Gulf of California. These air-parcel trajectories are referred to as Maritime Tropical.

The study area lies in the rain shadow of the Sierra Nevada, and the amount of precipitation reaching the ground decreases eastwardly from the Sierran crest (Table 1). Inputs to the Truckee River from tributaries downstream from Farad, California, are small, and most of the water reaching Pyramid Lake is streamflow discharge that originally fell as cool-season precipitation in the Sierra Nevada.

Evaporation from Pyramid Lake lags ambient air-temperature variation by about a

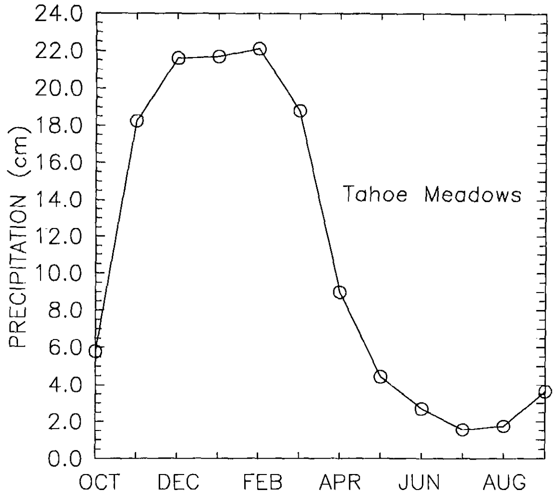

Fig. 2. Mean monthly precipitation at the Tahoe Meadows weather station between 1968 and 1991. Mean annual precipitation at this site is 131±52 cm.

month between March and June due to the thermal inertia (heat capacity) of water. Application of an improved form of a physically based thermal-evaporation model to Pyramid Lake (Hostetler and Benson 1990) has yielded evaporation rates of 1.29 and 1.18 m yr^{-1} for water years 1987 and 1988 (a water year begins on 1 October, i.e. water year 1987 began on 1 October 1986). Milne (1987) performed annual water-balance calculations, assuming that groundwater flux across the bottom of Pyramid Lake was zero, and determined that an evaporation rate of 1.25 m yr^{-1} provided the best match ($R^2 = 0.84$) between simulated and actual levels of the lake for the period 1917–1987. The similarity between Milne's (1987) and Hostetler and Benson's (1990) estimations of evaporation supports the assumption that groundwater recharge to Pyramid Lake is negligible.

Table 1. Precipitation data for selected weather stations in the Truckee River surface-water drainage. Location of stations shown in Fig. 1.

Weather station	Length of record (yr)	Precipitation (cm)
Tahoe Meadows	1968–1991	131.3±52.2
Reno airport	1937–1991	18.5±5.8
Sutcliffe	1967–1991	20.4±7.4
Nixon	1948–1974	16.5±6.0
Wadsworth	1974–1991	13.5±1.5

Table 2. Streamflow discharge at selected gauging stations on the Truckee River. Location of gauging stations shown in Fig. 1.

Gauging station	Period of record (water year)	Mean discharge (km^3 yr^{-1})
Tahoe City	1901–1990	0.228
Prosser	1943–1950, 1952–1990	0.078
Boca	1912–1915, 1940–1990	0.168
Farad	1900–1990	0.721
Steamboat	1961–1990	0.018
Vista	1901–1907, 1933–1954, 1959–1990	0.740
Wadsworth	1919–1957, 1959–1990	0.340
Nixon	1929–1990	0.380

Surface-water hydrology

The study area encompasses the Truckee River system, which drains the eastern slope of the Sierra Nevada and terminates in the Pyramid Lake subbasin (Fig. 1)—one of seven subbasins once occupied by Lake Lahontan (Fig. 3; Benson et al. 1990). Between 1882 and 1930, Pyramid Lake was hydrologically open when Mud Lake Slough spilled to the adjoining Winnemucca Dry Lake (Mud Lake) subbasin (Fig. 1) (Harding 1965; U.S. Geol. Surv. 1960). Pyramid Lake also spilled intermittently to Winnemucca Dry Lake during the past 3,000 yr (Benson et al. 1990). Since 1930, however, Pyramid Lake has existed in a hydrologically closed basin.

Discharge of the Truckee River is regulated by a system of reservoirs (Fig. 1); 31.6% of the long-term mean discharge upstream from Farad comes from Lake Tahoe, 4.4% comes from Donner Lake, 23.3% passes through Boca and Stampede Reservoirs, and 10.9% passes through Prosser Reservoir (Table 2). Thus, 70.2% of the discharge passing the Farad gauge has spent time in one or more reservoirs where it has undergone evaporation. The longest residence time (τ) of water in any reservoir except Lake Tahoe (τ = 235 yr) is calculated to be 1.7 yr (Stampede Reservoir; data for this calculation taken from U.S. Geol. Surv. 1961–1992).

Downstream from Farad, the Truckee River supplies water to Reno and Sparks, Nevada. In 1905, Derby Dam was completed, and diversion of water from the lower Truckee River to the Newlands Project via the Truckee Canal began in August 1906 (Townley 1980). Since

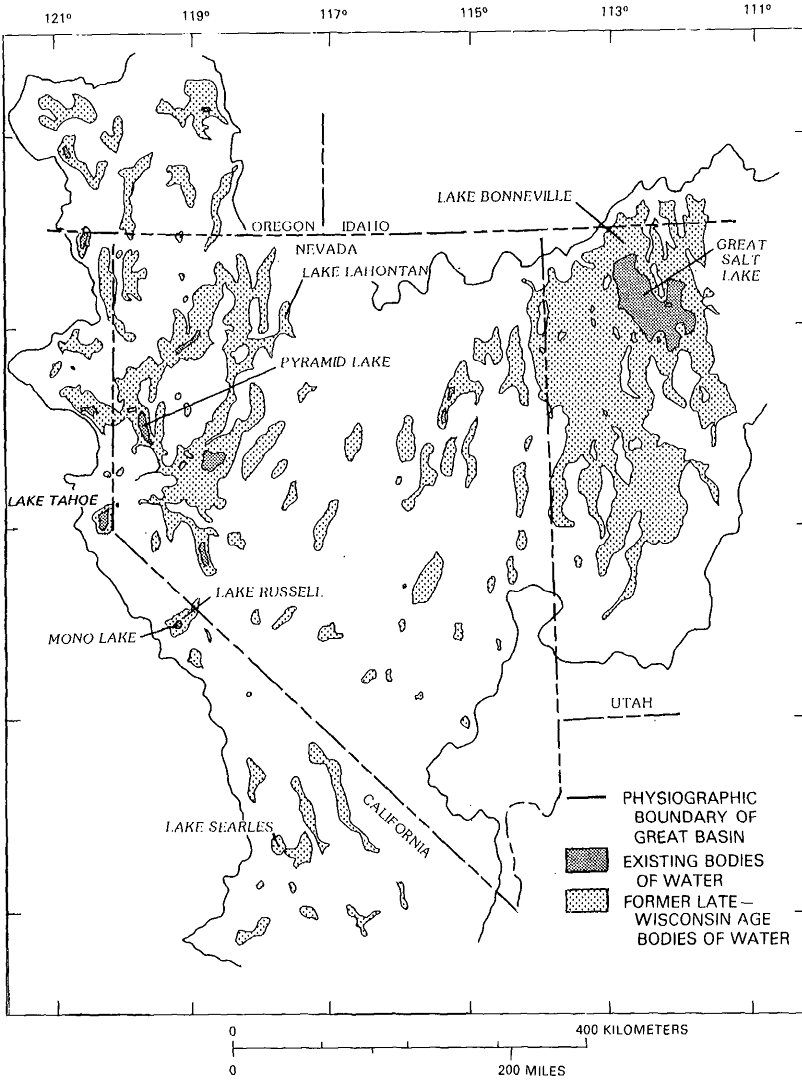

Fig. 3. Paleolakes of the Great Basin ~14,000 B.P. (from Benson et al. 1990).

1919, 54.1% of the discharge at Vista (Table 2) has been diverted from the Truckee River drainage (Table 2), causing the level of Pyramid Lake to fall ~20 m to its present (August 1992) elevation of 1,158 m (Fig. 4). Pyramid Lake would have maintained a surface elevation at or slightly below its spill point to Winnemucca subbasin (1,177 m) throughout the past 85 yr if water had not been diverted to the Newlands Project (Milne 1987).

Since this study began in May 1985, there has been 1 yr of above-normal precipitation in the northern Sierra Nevada followed by 6 yr of abnormally low precipitation. The six successive drought years caused Lake Tahoe to fall intermittently below its outlet to the Truckee River (Tahoe City) between October 1988 and September 1990; since September 1990, Lake Tahoe has remained below sill level and no discharge has passed Tahoe City (Fig. 5). The amount of discharge from Boca Reservoir and other tributaries to the upper Truckee River system (Fig. 1) also diminished during the drought years, and the discharge reaching Farad has continued to decline since winter 1986 (Fig. 5). Diversion of water via

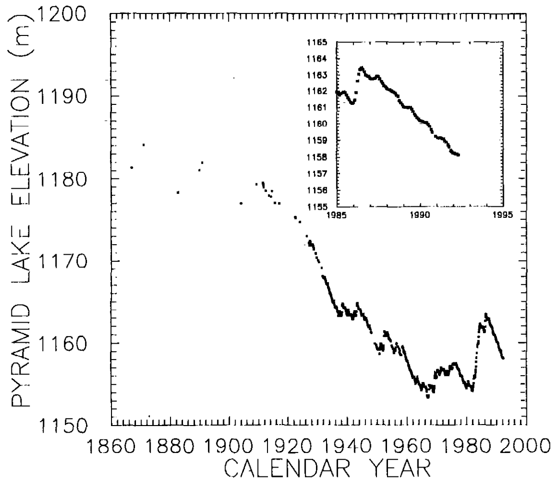

Fig. 4. Lake-surface elevation in the Pyramid Lake subbasin between 1867 and 1992 (data from U.S. Geol. Surv. 1960, 1963, 1961–1992).

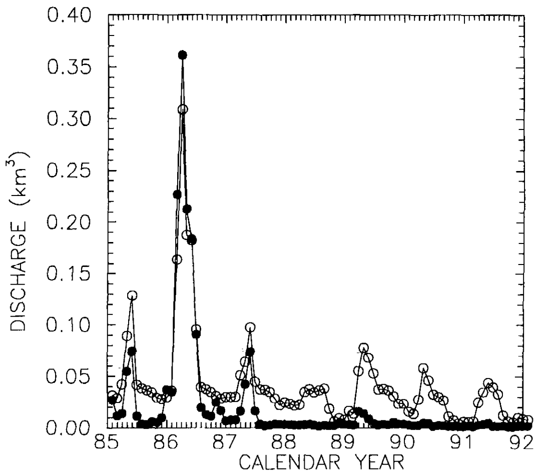

Fig. 6. As Fig. 5, but at the Farad (O) and Nixon (●) gauges.

the Truckee Canal at Derby Dam (Fig. 1) has further decreased discharge of the Truckee River throughout its lower reach; <0.0035 km^3 month^{-1} of discharge reached Nixon most of the time since July 1987 (Fig. 6), resulting in rapid decline in the level of Pyramid Lake (*see Fig. 4*).

Methods

Samples of precipitation, river, and lake water were collected between 29 May 1985 and 15 January 1992 (Fig. 1). Precipitation from individual storms was sampled at the Tahoe Meadows (elevation, 2,525 m) and Sutcliffe, Nevada, weather stations. Monthly samples of river discharge were taken at the Tahoe City, Farad, and Nixon gauging stations. Monthly surface-water samples were collected from Pyramid Lake in shallow water at one or more of three locations, including the west shore at Sutcliffe, the east shore due east of Pyramid Island, and a deep-water raft site (Fig. 1). Six sample profiles were collected from the deep-water raft between July 1985 and May 1987, and monthly profiles were obtained from the same site between February 1991 and January 1992.

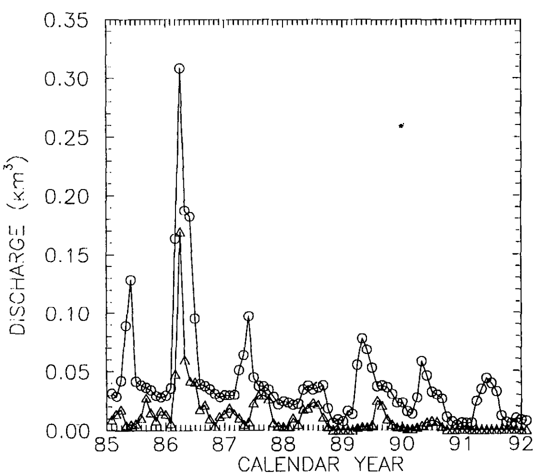

Fig. 5. Monthly Truckee River streamflow discharge at the Farad (O) and Tahoe City (△) gauges from January 1985 through February 1992 (data from U.S. Geol. Surv. 1986–1992).

Analyses of $\delta^{18}O$ and δ^2H were routinely performed on all river and lake samples, on the first 20 precipitation samples from Tahoe Meadows, and on 52 samples of precipitation collected from Sutcliffe. Most of the isotopic analyses were performed under the supervision of Tyler Coplen. The oxygen- and hydrogen-isotope results are reported in per mil (‰) relative to VSMOW (Vienna Standard Mean Ocean Water). The 2-σ precision of the oxygen and hydrogen results is 0.2 and 2‰.

Results

Stable isotopes in precipitation and groundwater in the Lake Tahoe basin—The $\delta^{18}O$ and δ^2H of Tahoe Meadows precipitation season-

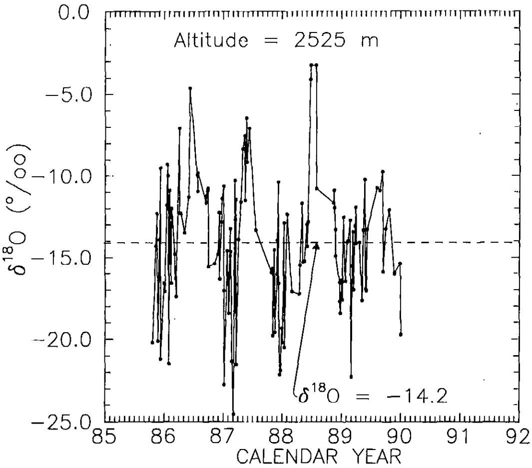

Fig. 7. Precipitation $\delta^{18}O$ at the Tahoe Meadows weather station from 21 October 1985 to 2 January 1990.

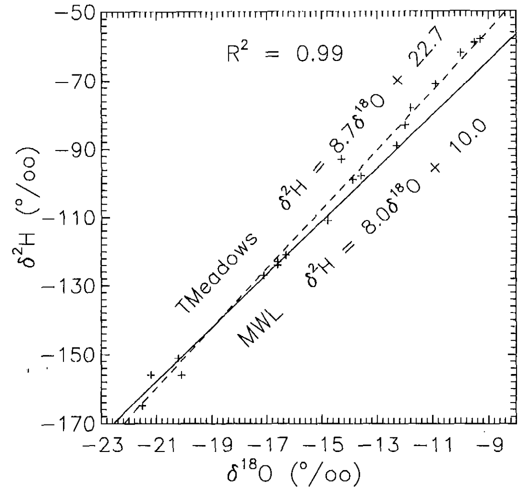

Fig. 8. The $\delta^{18}O$ and δ^2H of precipitation from 19 storms (+) reaching Tahoe Meadows between 21 October 1985 and 9 March 1986 compared to the meteoric waterline (MWL).

ally vary more than 20‰ and are most depleted (*depleted, lighter,* and *decrease in* are used to indicate that the $\delta^{18}O$ value is more negative and *enriched, heavier,* and *increase in* to indicate that the $\delta^{18}O$ value is more positive) in winter when precipitation falls as snow and most enriched in summer when precipitation falls as rain (Fig. 7). The seasonal differences can be attributed to several factors, including differences in temperature during evaporation of the source water, differences in temperature during precipitation (Friedman et al. 1964), and differences in storm-track trajectories (Benson and Klieforth 1989). Smith et al. (1979) showed that air masses that pass over the Sierra Nevada resulted in precipitation east of the divide depleted in δ^2H by amounts often exceeding 50‰ (equivalent to ~5‰ $\delta^{18}O$) compared to precipitation from air masses that reached the site by a low-altitude route.

The $\delta^{18}O$ and δ^2H of 19 samples of precipitation collected at 2,525 m at Tahoe Meadows (Fig. 8) plot close to the global mean meteoric waterline (MWL) (Craig 1961). The slope of the regression line is somewhat more positive than the slope of the MWL. The $\delta^{18}O$ and δ^2H of eight samples of groundwater (each of which represents an integration of many precipitation events) from the Lake Tahoe basin plot on the MWL (Fig. 9) and have a mean $\delta^{18}O$ of -14.6 ± 0.7‰—a value identical to the volume-weighted mean of 134 samples of Tahoe Meadows precipitation (Table 3). An additional 26 groundwater samples obtained from the Tahoe basin have a mean δ^2H of 107.7 ± 4.8‰ (S. Loeb pers. comm.). If we assume that these samples also lie on the MWL, a calculation indicates that the mean $\delta^{18}O$ value is -14.7 ± 0.7‰—nearly identical to the volume-weighted mean of Tahoe Meadows precipitation. It follows that the volume-weighted mean $\delta^{18}O$ of Tahoe Meadows precipitation and the mean $\delta^{18}O$ of Tahoe basin groundwater are identical.

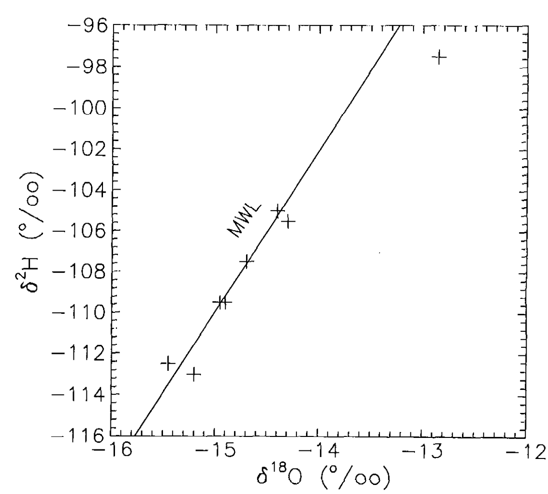

Fig. 9. The $\delta^{18}O$ and δ^2H of eight groundwater samples (+) from wells in the Tahoe basin compared to the meteoric waterline (MWL).

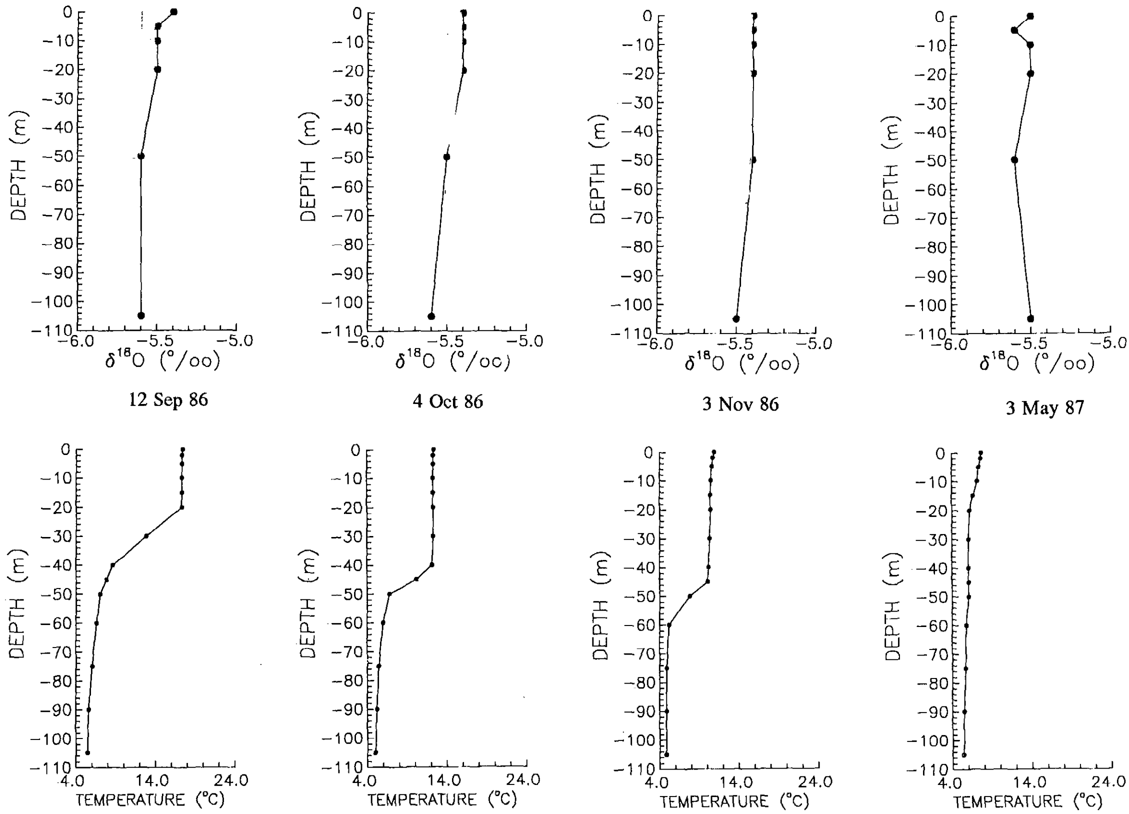

Fig. 10. Profiles of $\delta^{18}O$ and temperature in Lake Tahoe. Note the progressive enrichment of epilimnetic $\delta^{18}O$ that resulted from evaporation. Shaded areas indicate epilimnetic volumes.

Stable isotopes in Lake Tahoe and the Truckee River—Profiles of $\delta^{18}O$ in Lake Tahoe (Fig. 10) indicate that between 12 September 1986 and 3 May 1987 epilimnetic $\delta^{18}O$ increased as a result of isotopic fractionation that accompanied evaporation. This same phenomenon was detected in the outflow of Lake Tahoe at the Tahoe City gauge, where $\delta^{18}O$ increased between 3 June 1985 and 2 November 1987.

Table 3. Stable-isotope statistics for the Truckee River surface-water system. Tahoe Meadows and Sutcliffe precipitation sites are located at altitudes of 2,525 and 1,164 m. (X—mean; σ—standard deviation; $X_{vol.\,wt}$—weighted mean.)

Sampling site	$\delta^{18}O$ ($X \pm \sigma$, $X_{vol.\,wt}$)	No. samples
Tahoe Meadows precipitation	−14.2±4.0, −14.6	134
Wells around Lake Tahoe	−14.6±0.7	8
Truckee River (Tahoe City)	−5.5±0.2, −5.5	30
Truckee River (Farad)	−10.3±1.7, −10.1	55
Truckee River (Nixon)	−10.4±0.8, −9.9	61
Sutcliffe precipitation	−9.8±4.4, −10.6	52

Discharge reaching the Farad gauge is composed of precipitation that has undergone various degrees of isotopic fractionation in six upstream reservoirs and lakes (Fig. 1). As a result of the mixing of fractionated discharge from reservoirs with unfractionated runoff, the $\delta^{18}O$ of discharge reaching Farad is variable (Fig. 11), and the $\delta^{18}O$ and $\delta^{2}H$ values of the discharge do not plot on the MWL (Fig. 12). The residence time of water in lakes and reservoirs (except Lake Tahoe) is small, and the isotopic variability observed at Farad can be considered a mixture of isotopically enriched Lake Tahoe water and isotopically depleted precipitation (Fig. 13). By the time Truckee River discharge reaches Nixon, the variance

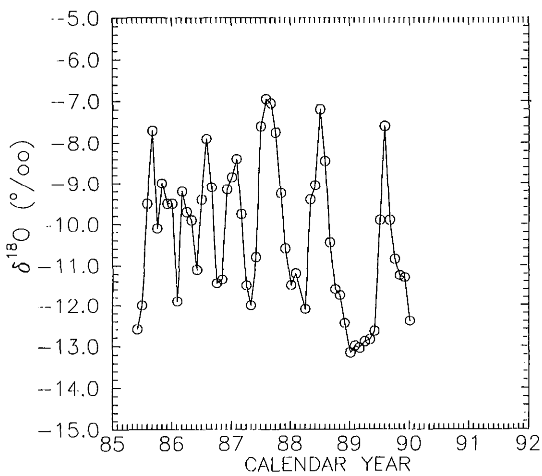

Fig. 11. The $\delta^{18}O$ in Truckee River discharge at Farad as a function of time.

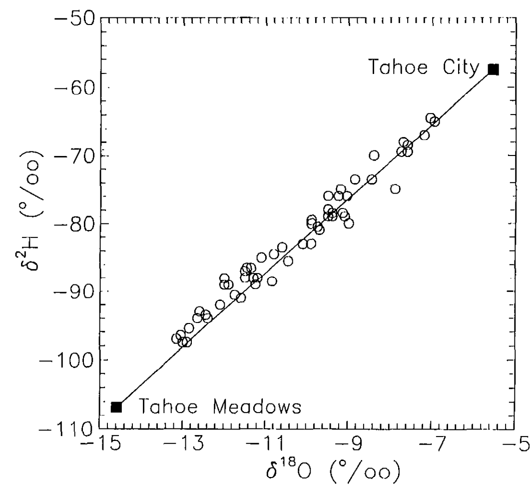

Fig. 13. The $\delta^{18}O$ and $\delta^{2}H$ of Truckee River discharge (O) at Farad plotted on a mixing diagram with Tahoe Meadows precipitation and Lake Tahoe discharge end-members.

of $\delta^{18}O$ is reduced, although the mean value of $\delta^{18}O$ is about the same (Table 3).

Stable isotopes in precipitation falling on Pyramid Lake—At the Sutcliffe weather station (elevation, 1,163 m), which borders the western shore of Pyramid Lake, the mean $\delta^{18}O$ of precipitation ($-9.8 \pm 4.4‰$) is enriched relative to Tahoe Meadows precipitation ($-14.2 \pm 4.0‰$) (Table 3). Most of the enrichment of Sutcliffe $\delta^{18}O$ is due to warmer condensation air temperatures, but some of the enrichment can be attributed to evaporation of falling raindrops, as indicated by samples that plot under the MWL (Fig. 14).

Stable isotopes in Pyramid Lake—The large quantity of isotopically depleted inflow to Pyramid Lake in spring 1986 and a moderate quantity of isotopically depleted inflow in spring 1987 (Fig. 8) caused the $\delta^{18}O$ of the surface water to decrease by 1.2 and 0.6‰ (Fig. 15). For the most part, subsequent minima in the $\delta^{18}O$ of the surface water resulted from mixing of evaporatively enriched epilimnetic water with a large volume of relatively de-

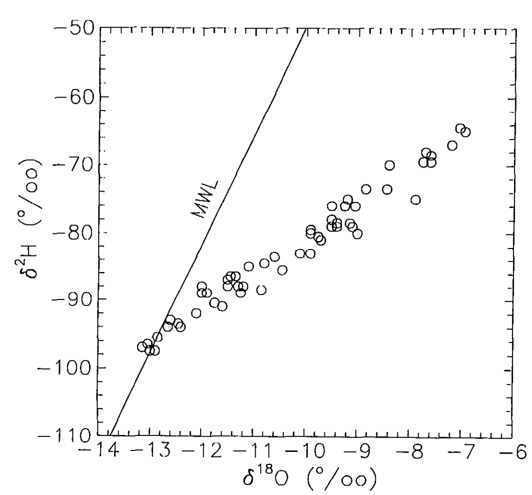

Fig. 12. The $\delta^{18}O$ and $\delta^{2}H$ of Truckee River discharge (O) at Farad compared to the meteoric waterline (MWL).

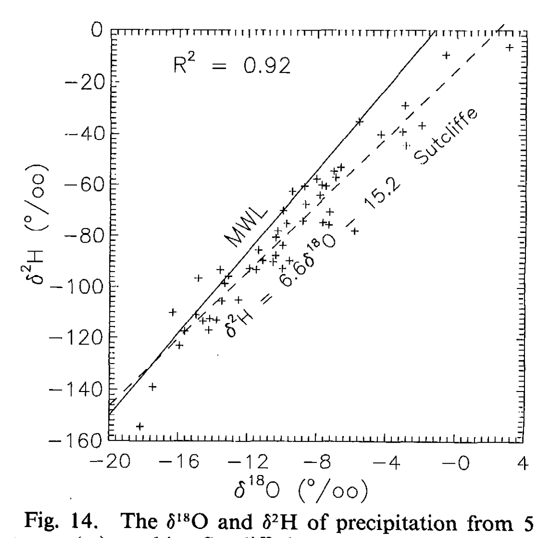

Fig. 14. The $\delta^{18}O$ and $\delta^{2}H$ of precipitation from 52 storms (+) reaching Sutcliffe between 21 June 1988 and 3 October compared to the meteoric water line (MWL).

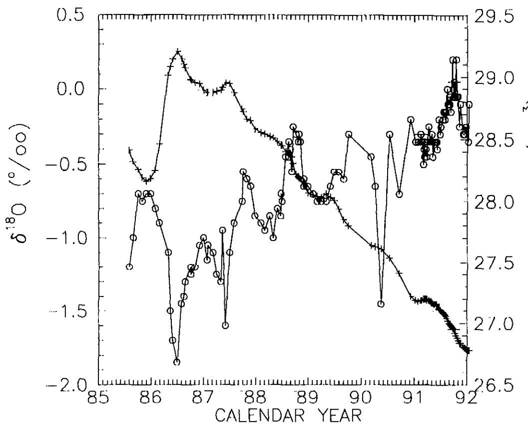

Fig. 15. Volume (+) of Pyramid Lake and $\delta^{18}O$ (O) of the surface water from 29 July 1985 to 14 January 1992.

pleted hypolimnetic water during turnover in January.

As a result of evaporation exceeding discharge plus on-lake precipitation since June 1986, the volume of the lake has decreased by 8.5%, and the $\delta^{18}O$ of lake water has correspondingly increased by ~2.0‰ (Fig. 15). Some of the observed first-order and much of the observed second-order variance in surface-water $\delta^{18}O$ is the result of transient processes confined to the upper few centimeters of the epilimnion. Dilute, isotopically depleted discharge has been observed to float on the relatively dense lake water. This phenomenon may explain the isotopically depleted sample that was taken in May 1990. In late summer and autumn, >1.0 cm of water evaporates from the surface layer of Pyramid Lake (Hostetler and Benson 1990), causing enrichment in $\delta^{18}O$. Profiles of $\delta^{18}O$ (Fig. 16) indicate that in spring 1986, epilimnetic $\delta^{18}O$ was depleted by input of Truckee River water. Evaporation in autumn 1991 caused enrichment of $\delta^{18}O$ throughout the entire epilimnion.

Conceptual model of $\delta^{18}O$ variation in Pyramid Lake

The $\delta^{18}O$ value of the water varies continuously as isotopically depleted water is evaporated from the epilimnion and isotopically depleted water is input to the lake as streamflow discharge and on-lake precipitation. The components of the hydrologic balance in the Pyramid Lake system are seasonally shifted; i.e. precipitation is greatest in winter, discharge is greatest in spring, and evaporation is greatest in autumn. It follows that in a normal water year the $\delta^{18}O$ of the surface water also varies seasonally. In winter, isotopically depleted precipitation falls on the lake surface; in spring, the Truckee River discharges relatively depleted $\delta^{18}O$ to the epilimnion; and in summer and autumn, epilimnetic $\delta^{18}O$ becomes more enriched as evaporation preferentially transfers the lighter isotopes of oxygen to the atmosphere. In January, when turnover occurs, water from all depths mixes and $\delta^{18}O$ becomes uniform throughout the lake, causing the $\delta^{18}O$ value in the hypolimnion to change in a stepwise fashion. At any particular time, the $\delta^{18}O$ value of the epilimnion is depleted or enriched relative to the volume-weighted mean $\delta^{18}O$ of the lake, depending on the amount of isotopically depleted surface-water input to the lake relative to the amount of isotopically depleted evaporation that leaves the lake. In the wet year of 1986, streamflow discharge depleted the epilimnion by ~1.0‰, and, in the dry year of 1991, evaporation enriched the epilimnion by ~0.5‰ (Fig. 15).

Because climate varies continuously, Pyramid Lake never achieves a hydrologic steady state or the isotopic steady state defined by Craig et al. (1963). Therefore, a return to a former set of environmental conditions does not imply a return to the former $\delta^{18}O$ value associated with the past environment (Lister et al. 1991). In addition, Pyramid Lake can change from a hydrologically closed to a hydrologically open system when it rises above 1,177 m and spills to the Winnemucca Lake subbasin (Fig. 1). The Pyramid Lake isotopic system is continually forced by changes in climate and the hydrologic balance. Thus, the historical and prehistorical $\delta^{18}O$ records of the lake must be modeled dynamically, whether the model is conceptual or numerical in nature.

Transfer of lake-water stable-isotope variation to the sedimentary record

Variability of water temperature in space and time—Pyramid Lake water temperature also varies in space and time. Surface-water temperature annually varies between 6 and 24°C, while deep-water temperature remains constant at ~6°C (Benson 1984). The incorporation of $\delta^{18}O$ in carbonates is a function of temperature (e.g. the $\delta^{18}O$ fractionation factor

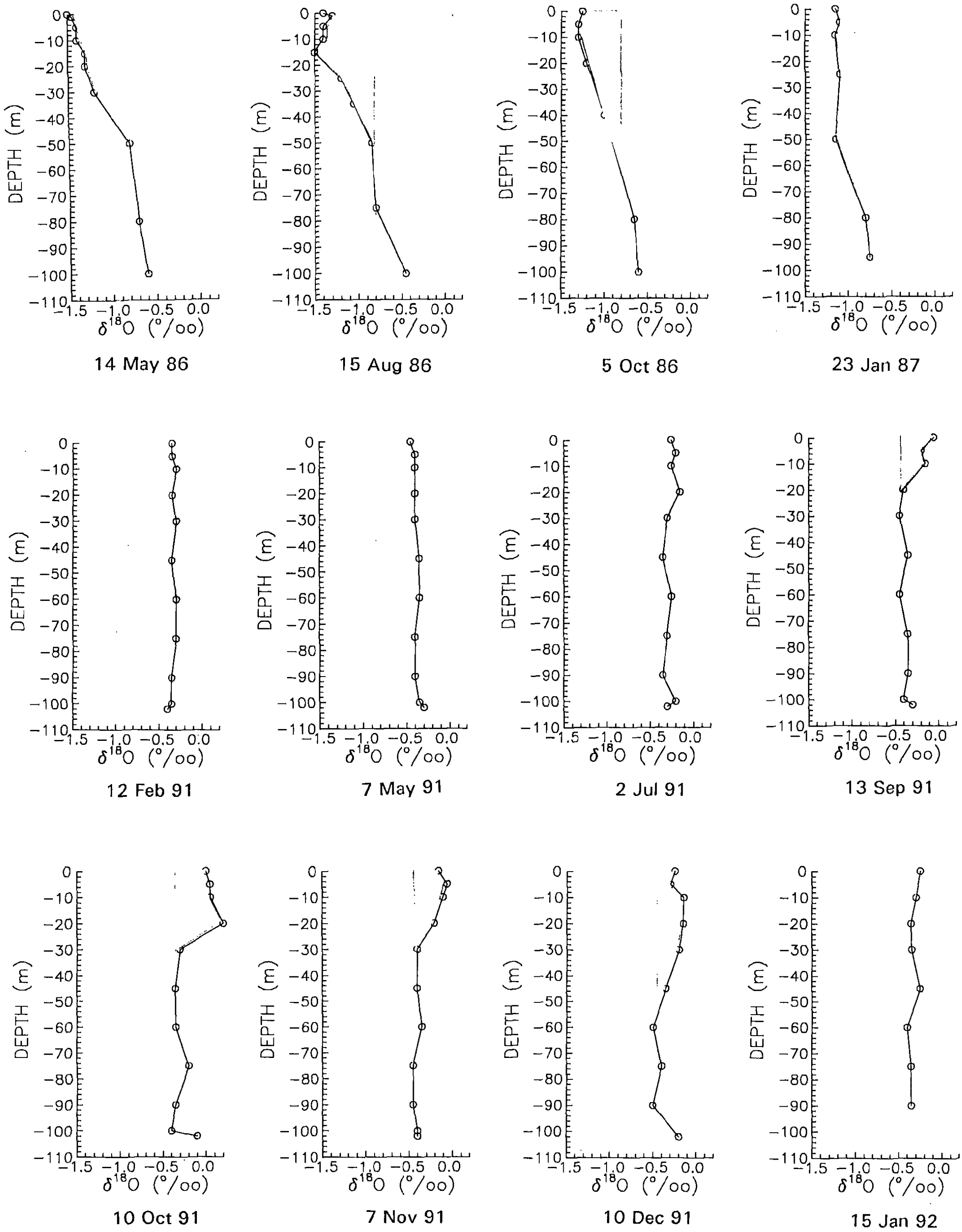

Fig. 16. Profiles of $\delta^{18}O$ in Pyramid Lake. Note that in 1986 and early 1987 the epilimnion became depleted when a large amount of isotopically depleted Truckee River water discharged to the lake (*see Fig. 8*). The enrichment of epilimnion water resulting from evaporation is shown for the period 13 September 1991 to 15 January 1992. Shaded areas indicate epilimnetic volumes.

between water and calcite decreases ~0.25‰ per °C, Craig 1965). Therefore, because of the temperature effect, carbonates formed in the epilimnion at various times during the annual cycle today may exhibit a ≤4.5‰ variance in $\delta^{18}O$. Epilimnetic carbonate samples that integrate the effects of the annual temperature cycle will be ~2.2‰ more depleted than samples formed in the hypolimnion.

In the past, deep lakes existed in the Pyramid Lake subbasin, and their thermal structure was different. For example, simulations of the effect of a jet-stream climatology on the hydrologic and thermal balance of Pyramid Lake indicate that when Pyramid Lake was part of Lake Lahontan surface-water temperature varied between 1 and 14°C, and deep-water temperature remained constant at ~2°C (Hostetler and Benson 1990). Carbonates precipitated in surface waters of this paleolake over the annual cycle would have exhibited a 3.2‰ variance in $\delta^{18}O$ and, due to colder water temperatures, the $\delta^{18}O$ value of paleolake epilimnetic carbonates would have been enriched ~2.0‰ relative to modern samples precipitated in the same environment. Today, epilimnetic and hypolimnetic carbonate samples that integrate the effects of the annual temperature cycle differ by ~2.2‰; during the existence of the colder paleolake such samples would have differed by 1.5‰.

Calcium carbonate formation in Pyramid Lake—Fractionation of the isotopes of oxygen between a carbonate precipitate and lake water is also a function of the mineralogy and chemistry of the precipitate (Friedman and O'Neil 1977; Tarutani et al. 1969). In Pyramid Lake, carbonate is presently produced by formation of low-Mg calcite ostracod valves and by precipitation of aragonite "whitings." Ostracods live at the sediment–water interface of the lake; they have calcite carapaces (two valves) that are shed at molting, generating sequentially larger valves up to nine times before reaching maturity (Chivas et al. 1986). The $\delta^{18}O$ incorporated in the calcite of these valves is acquired during molting—a process that takes only a few days (Turpen and Angell 1971). Due to vital effects, the fractionation of oxygen isotopes between ostracods and water may vary among species. Thus, it is important to analyze valves from species for which the departure from equilibrium isotopic fractionation is known. At this time (1993), experimental data that demonstrate the magnitude of vital effects on isotopic fractionation are not available.

Whitings in Pyramid Lake occur every 5–10 yr in autumn, when the epilimnion temperature reaches ~22°C and when aragonite saturation is ~10× (Galat and Jacobsen 1985). Aragonite whitings should be enriched relative to ostracods formed at the same time because aragonite is enriched ~0.6‰ relative to calcite formed under the same conditions (Tarutani et al. 1969).

Sampling strategy—As discussed above, fractionation of $\delta^{18}O$ between a carbonate precipitate and lake water is a function of the mineralogy of the precipitate, the chemistry of the precipitate, the temperature of precipitation, and the $\delta^{18}O$ content of the lake water. If we want to estimate the volume-weighted mean $\delta^{18}O$ of Pyramid Lake for some past time, it is preferable to analyze a thermodynamically stable form of carbonate that formed below the thermocline (Benson et al. 1991). This precludes analysis of carbonates that have recrystallized from metastable precursors such as aragonite whitings and high-Mg calcites. Fossil deep-water benthic ostracods are ideal for such an analysis because they are composed of a stable form of carbonate that was deposited in water whose temperature and $\delta^{18}O$ were relatively invariant (Benson et al. 1991; Lister et al. 1991).

I have discussed isotopic variability in the context of a closed-basin system. In the past, Pyramid Lake, as well as many of the lakes in the principal paleolake systems of the Great Basin, spilled to adjoining basins (Benson et al. 1990 and references therein). The transition from a closed- to an open-basin system should be readily identifiable in the isotopic record. Because the residence time of water in a closed-basin lake is greater than the residence time of water in an open-basin lake with a similar volume, the $\delta^{18}O$ of lake water will rapidly shift to more depleted values when spill commences.

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Submitted: 2 April 1993
Accepted: 12 July 1993
Amended: 19 August 1993