LATE-WISCONSIN PALEOHYDROLOGY OF THE WEST-CENTRAL AMARGOSA DESERT, NEVADA, U.S.A.

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Abstract

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Studies of isotopes in groundwater using ^{14}C, ^2H (deuterium) and ^{18}O have provided significant insight into the paleoclimate of middle and late Wisconsin age in a typical arid environment of southern Nevada, the west-central Amargosa Desert. Evidence indicates that recharge probably was through infiltration of runoff in paleostream channels and that this runoff was important only from ~ 17,000 to ~ 10,000 yr. B.P. Mean annual temperature at 17,000 yr. B.P. was ~ 8°C less than present mean annual temperature; some summer moisture was effective in recharge. Winter (October—May) temperature at ~ 10,000 yr. B.P. was ~ 1°C less than present; summer moisture did not contribute to recharge.

1. Introduction

A significant source of potable groundwater lies beneath the west-central Amargosa Desert, ~ 100 km northwest of Las Vegas, Nevada. That quality of water and the long growing season made the area attractive for homesteading during the Great Depression of the 1920's and 1930's. Current interest in the area is focused on the source of this groundwater, because the groundwater lies downgradient from a potential nuclear-waste isolation site.

1.1. Purpose and scope

The objective of this study is to obtain information pertaining to the source of the groundwater beneath the west-central Amargosa Desert and the climatic regime prevailing at the time of recharge. It is believed that recharge in the area is unlikely under present climatic conditions. The carbon-, hydrogen- and oxygen-isotope composition of groundwater was used in this investigation because it is especially useful in determining climatic conditions during recharge and variability of climate with time.

What then is the source of the groundwater beneath the west-central Amargosa Desert? Two hypotheses are considered: (1) infiltration of runoff that occurred in the Amargosa Desert during a climatic regime wetter and cooler than the present regime, when runoff was more frequent and runoff volumes were larger, or (2) subsurface movement of recharge that occurred in highland areas to the north. Isotope evidence presented here
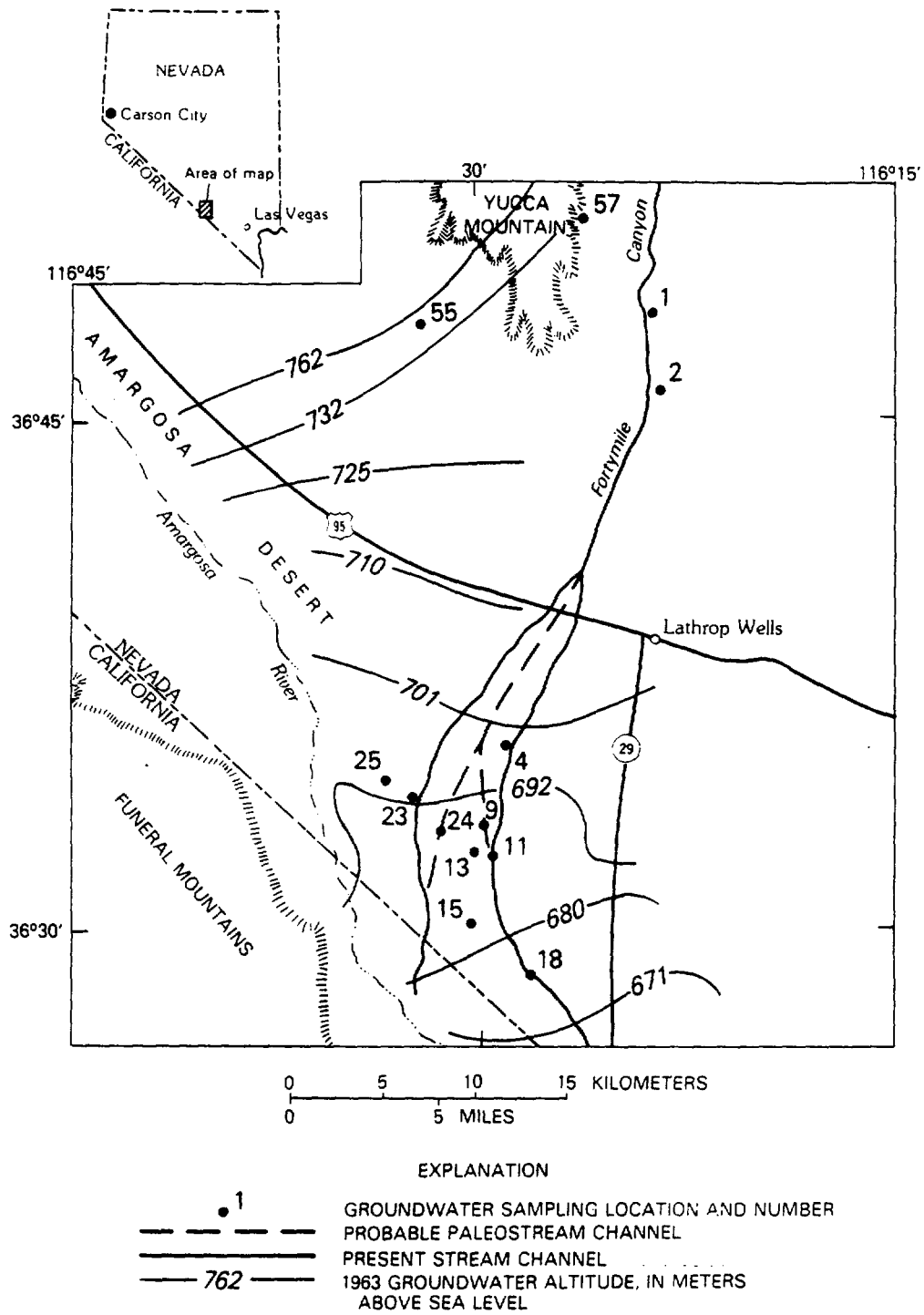
Fig. 1. Amargosa Desert and vicinity.

will be used to choose between these hypotheses, as well as to describe the climatic regime extant at the time of recharge.

1.2. Description of the study area

The west-central Amargosa Desert is a basin, at ~ 1000 m above mean sea level (a.s.l.), with a present-day climate classified as arid (Walker and Eakin, 1963; Winograd and Thordarson, 1975); the approximate study area is shown in Fig. 1. Hydrologic information for the area also is included in Fig. 1. A detailed discussion of the hydrology may be found in Walker and Eakin (1963); surface geology is described in Denney and Drewes (1965) and Swadley (1983).

Annual precipitation in the area is ~ 5—10 cm (Geraghty et al., 1973); mean annual temperature is ~ 20°C, a climate not conducive to groundwater recharge because of intensive evaporation. Occasional flooding occurs in the area; source of the flooding primarily is runoff from the uplands to the north and west. Although flood frequency is not recorded, present frequency appears to be about once a decade. Christensen and Spahr (1980) indicate that a 50-$m^3 \cdot s^{-1}$ flood in the major drainage way (Fortymile Canyon, Fig. 1) may be expected every 10 yr.

2. Theory and results

2.1. Carbon isotopes

Specific ^{14}C activities were determined for each groundwater sample. Ages may be computed from specific activities, if little or no carbon-isotope dilution occurs after recharge. Absence of isotope-dilution effects was determined from major-ion inorganic chemistry; this procedure is discussed in detail in Claassen (1985). Briefly, the procedure is: (1) a groundwater chemical evolution model is developed for tuffaceous aquifers; and (2) the chemistry of each

sample is compared to the family of allowable chemistries for the appropriate evolution path; if a sample chemistry is within 5% of allowance for select dissolved constituents, this sample is assumed not to be subject to carbon-isotope dilution in any substantive way. The method was further tested using water chemistries that were known to have significant isotope dilution and screened effectively against that dilution.

All samples collected in the Amargosa Desert that passed the screening process are included in Table I and the age dates are presumed correct. All samples were collected from wells in frequent use, or after extensive pumping, to assure that samples were representative of the aquifer. Samples for stable isotopes were collected in glass bottles that were completely filled and sealed to prevent evaporation. Radioactivity assay for carbon was performed on either $BaCO_3$ or $SrCO_3$, precipitated from 200-l barrels of sample, filled and sealed prior to chemical separation of inorganic carbon. All analyses were performed by U.S. Geological Survey laboratories in Denver, Colorado, or Reston, Virginia.

2.2. Hydrogen and oxygen isotopes

It is generally accepted that both hydrogen and oxygen isotopes fractionate during condensation of water vapor in a manner that causes the condensed phase to be more depleted in the heavier isotope as the temperature of condensation decreases. This suggests that isotope ratios can be used as measures of condensation temperature and, if sufficiently integrated over time, climate. Although the relationship between temperature and isotope depletion in precipitation has been described by Dansgaard (1964), Merlivat and Jouzel (1979), and Van der Straaten and Mook (1983), the more succinct treatment of the importance of the different processes given by Van der Straaten and Mook (1983) will be used as the basis for the following discussion.

TABLE I

Groundwater isotope data from Amargosa Desert and vicinity

Site[1]	Location	Well depth (m)	Depth to water (m)	^{14}C (yr. B.P.)	$\delta^2 H^{[2]}$	$\delta^{18}O^{[2]}$
1	14 km N of Lathrop Wells	1,060	283	9,900	−97.5	−13.0
2	11 km N of Lathrop Wells	350	226	9,100	−97.5	−12.8
4	16S/49E-5acc	90	21	13,200	−103	−13.2
9	16S/49E-18dc	110	33	10,100	−102	−12.6
11	16S/49E-19daa	90	30	12,600	−101	−13.1
13	16S/48E-25aa	50	26	13,200	−102	−13.0
15	17S/48E-1ab	60	16	13,600	−104	−13.0
18	17S/49E-8ddb	100	15	10,300	−102	−13.0
21	16S/49E-23add	—	—	10,400	−99	−13.2
23	16S/48E-15aaa	50	29	14,200	−103	−13.4
25	16S/48E-10cba	—	—	14,900	−102	−13.4
55	18 km NW of Lathrop Wells	760	184	17,000	−108	−14.2
57	18 km N of Lathrop Wells	1,220	471	14,100	−101	−13.4
64	35 km N of Lathrop Wells	1,680	328	11,000	−104	−13.0

All analyses by U.S. Geological Survey Laboratories.
[1] Shown in Fig. 1, except for sites 55, 57 and 64, which are north of map boundary, in Fortymile Canyon or tributaries.
[2] Parts per thousand deviation from standard mean ocean water (SMOW).

The Rayleigh condensation model is used for derivation of the isotopic composition of condensate from an air mass with given composition. This model is coupled with another model for the isotopic composition of an air mass over an evaporating water body (ocean) to provide the two-box model of Craig and Gordon (1965). The assumption is that the only important processes in determining condensate composition are initial evaporation and subsequent condensations. No re-evaporation occurs, and no new moisture is added to the air mass as it moves from the source to the area of interest. For coastal locations at southern latitudes, this assumption appears to give a very good approximation; however, deviations from theory are expected at northern latitudes because re-evaporation of previously precipitated moisture is added to the air mass, changing its composition for subsequent condensation steps. Van der Straaten and Mook (1983) observed this deviation for northern-latitude

(generally north of 30°) data from theory. Their equations for dependence of hydrogen- and oxygen-isotope shifts on condensation (or mean annual) temperature are:

$$d(\delta^2 H)/dT = 6.0 - 0.90T \qquad (1)$$

and

$$d(\delta^{18}O)/dT = 0.72 - 0.010T \qquad (2)$$

Integrating and determining the constant of integration from International Atomic Energy Agency (I.A.E.A.) data, they obtained:

$$\delta^2 H = 4.9T - 97 \qquad (3)$$

and

$$\delta^{18}O = 0.56T - 12.7 \qquad (4)$$

It remains to be proven that these equations are applicable for continental climates of the western U.S.A.

To attempt to show the foregoing, information on isotope composition in precipitation

TABLE II

Mean daily near-surface temperature, δ^2H and $\delta^{18}O$ of precipitation at a mountain site in Colorado (38°N, 107°W)

Mean daily temperature (°C)	$\delta^2H^{(*)}$	$\delta^{18}O^{(*)}$
11.6	−39.5	− 6.6
6.0	−115.0	−15.5
0.4	−107.0	−15.1
−4.8	−152.5	−20.4
−9.8	−142.5	−19.0

*Parts per thousand deviation from standard mean ocean water (SMOW).

at a southwestern Colorado mountain site (38°N, 107°W) is presented in Table II. Temperatures are near-surface daily means for the almost monthly periods during which the precipitation samples were integrated. Regression equations are:

$$\delta^2H = 4.58T - 113 \tag{5}$$

and

$$\delta^{18}O = 0.57T - 15.7, \quad r = 0.88 \tag{6}$$

The slopes are in excellent agreement with the equations derived from I.A.E.A. data, but the intercepts are more negative. If these equations are used to calculate temperature of condensation from isotope data, eqs. 3 and 4 would calculate temperatures 5.3°C lower than if eqs. 5 and 6 were used. The probable reason for the different intercepts can be more easily seen if the ^2H and ^{18}O equations are combined to yield the meteoric-water line:

$$\delta^2H = 7.97\delta^{18}O + 12.4 \tag{7}$$

and

$$\delta^2H = 8.15\delta^{18}O + 10.0 \tag{8}$$

Derivation of eq. 8 (Van der Straaten and Mook, 1983, equation 17) assumed an ocean-surface temperature of 25°C, a relative humidity of 85%, and a fractional initial condensation of 0.5. Deviation from these values with respect to evaporation temperature or with respect to fraction of moisture initially condensed has negligible effect on either the slope or intercept of the isotope—temperature curves, and, therefore, negligible effect on the meteoric-water line derived therefrom (Van der Straaten and Mook, 1983). In contrast, however, is the effect of the relative humidity of the atmosphere above the evaporating body. Although little effect on slope can be demonstrated, the effect on intercept is significant. For example, a decrease in relative humidity from 85% to 65% results in a change in intercept of eq. 8 from 10.0 to 19.0‰ (at evaporation temperature of 25°C).

Temporal changes in oceanic-source humidity would be expected to have a large effect on the intercept (but not slope) of the meteoric-water line; however, evaporated continental moisture at northern mid-latitudes would be expected to have a similar effect because the relative humidity at these latitudes is generally less than that at the primary oceanic source, and the fraction of the total vapor mass resulting from re-evaporated moisture may be significant.

Although the relationship between precipitation isotope-ratio changes and temperature is not significantly affected by other climatic variables, the absolute value of temperature calculated from a particular isotope ratio is affected by the exact nature of the evaporation or re-evaporation process. Therefore, extrapolation of data from one site to another requires acceptance of some assumptions, and additional assumptions need to be made, if extrapolation to paleoclimatic conditions is desired.

2.3. Extrapolation of isotope lapse rates to other sites

For continental locations, the major assumptions for extrapolation of data to other sites are that initial moisture sources for the two sites are similar (moisture derived by evaporation into atmosphere of similar

relative humidity), and that re-evaporation of precipitated moisture has a similar effect on the isotope ratio of precipitation at the two sites.

Although the calibration site in Colorado is located less than 1° further north than the Nevada site and both sites are affected by present-day air masses with similar trajectories, the weighting of these trajectories, according to the relative quantities of moisture reported at each site, may be different. The assumption will be made that present-day meteoric-water lines for the two sites will have the same slope and, more importantly, the same intercept (^{2}H excess).

2.4. Extrapolation of present-day isotope lapse rates to paleoclimatic scenarios

The validity of applying contemporary-precipitation isotope measurements to analysis of paleoclimatic conditions is a matter of concern. These concerns are grouped into two categories: (1) variations in moisture—source—evaporation conditions with time; and (2) changes in isotope ratios resulting from fractionation during groundwater recharge.

No successful attempts to respond to the first concern have been completed; it is accepted as a theorem without proof. A corollary concern is the possibility of a seasonal shift in effective moisture distribution from paleoclimatic conditions to the present. These concerns have not been addressed currently; usually it is assumed that no such shift has taken place. For example, the observation that a 3-yr. integration period occurs from recharge to cave seepage water is used to dispel concern about long-term variability in effective moisture (Schwarcz and Yonge, 1983). In this instance, short-term stability of a system becomes justification for long-term stability, a condition not likely for glacial to interglacial transitions.

Observations of changes in isotope composition during recharge have been made by many investigators (Gat and Tzur, 1967;

Bath, 1983). Gat and Tzur (1967) estimate that a maximum enrichment of < 1‰ in ^{18}O might be expected as a result of evapotranspiration during recharge. Unpublished results (1984) of H.C. Claassen are in agreement with that determination, indicating a shift of ~ +0.3‰. Therefore, it is necessary to correct the determined groundwater isotope ratio for changes resulting from the recharge process; however, because the slope of the δ^2H vs. δ^{18}O line describing the process is not known with certainty, either the δ^2H or the δ^{18}O can be corrected, but not both.

3. Discussion

3.1. Interpretation of carbon-isotope data

The water-level contours in Fig. 1 indicate the possibility of distant recharge and slow downgradient movement with significant lateral and vertical dispersion in the valley fill. Position along the downgradient vector would be expected to correlate to age, older samples being found further downgradient. The available data are inconsistent with this hypothesis.

Determinations of groundwater age vs. position in alluvial basins have suggested that, in some basins, younger water overlies older water (Tucson Basin, Arizona). This might be correct, if recharge were areally homogeneous — that is, infiltrating through many small channels that distribute runoff throughout a wide area. This process would result in a significant correlation between the depth of saturated zone sampled and the age of water from the well; wells completed in only the top of the saturated zone would yield more modern samples than wells completed at greater depths in the saturated valley fill. Data from sites 4—25 (Fig. 1) were analyzed to test this hypothesis. Samples from sites 1, 2, 55, 57 and 64 were excluded, because they were not directly within the west-central Amargosa Desert. Correlation between reach of saturated zone yielding water to a well and age of
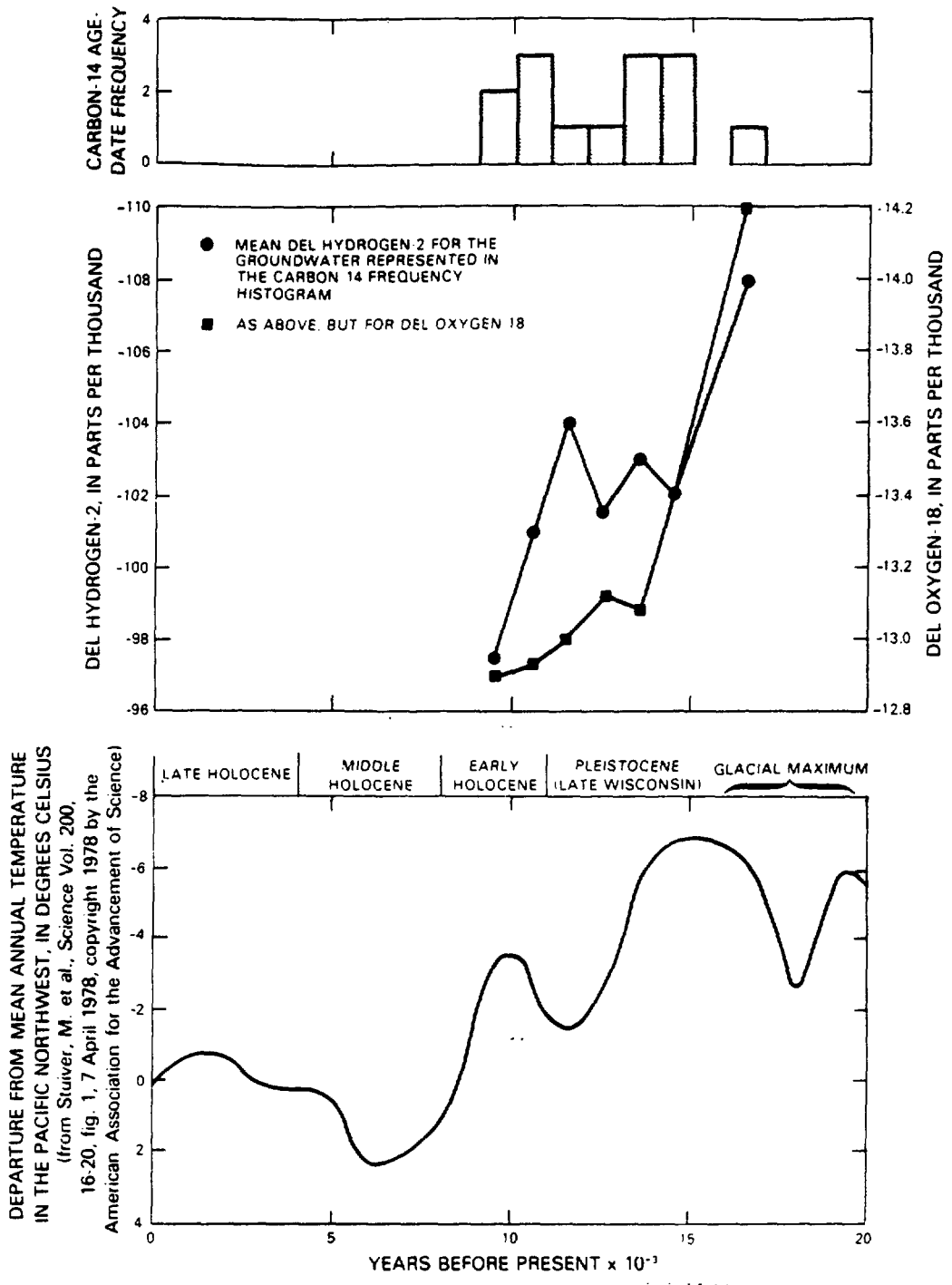
Fig. 2. Carbon-, hydrogen- and oxygen-isotope data for water in tuff and tuffaceous valley fill and their relationship to climatic change.

groundwater sampled is —0.82 ($N = 7$).

Testing a third hypothesis that recharge only occurs effectively via stream channels requires that the location of stream channels be known. A first approximation to the location of paleochannels may be made by observing present major channel locations (shown as solid lines in Fig. 1). A better approximation may be made by identifying the location of the youngest groundwater sampled. Probable paleochannels are indicated in Fig. 1 by the dashed lines that were sketched using the location of the youngest groundwater sampled. The estimated locations are similar. Using this paleochannel configuration, the relative distance of the other sampling points from this channel was determined. Where their location was between channel branches, the average relative distance was used. The correlation of age to relative distance from the proposed paleochannel is high: $r = 0.97$ ($N = 6$). It is not known whether this correlation indicates some regular progression of channel location with time or a fixed location, with time-progressive lateral movement of water recharged from the main channel.

Frequency distribution of ^{14}C ages is shown in Fig. 2. If random sampling of groundwater is accomplished by the data set in Table I, the approximate time distribution of recharge may be inferred, which may be compared to the temporal changes in mean annual temperature for the Pacific northwest U.S.A. (Stuiver et al., 1978). This reference contains the only quantitative estimate for the western U.S.A. Although the absolute magnitude of temperature changes in southern Nevada may be different, the trends appear to correlate well with the frequency distribution.

An additional observation is that no samples older than 17,000 yr. B.P. were obtained, although the Wisconsin glacial maximum occurred prior to that time and it is within ^{14}C dating range, which generally is defined to be ~ 30,000 yr. B.P. Recharge is determined by effective, not total, moisture; if evaporation is large, even large quantities of precipitation may result in no recharge. Low temperatures, such as those during the Wisconsin glacial maximum, generally indicate less evaporation, maximizing effective moisture and fostering recharge; therefore, the absence of any evidence for recharge during the Wisconsin glacial maximum is surprising. One possibility is that effective moisture was less during the Wisconsin glacial maximum than during 17,000 to 9,000 yr. B.P.; this is supported by evidence obtained from fossil packrat middens (Spaulding et al., 1984).

A second possible explanation for the absence of groundwater recharged during the glacial maximum is that such groundwater was not sampled by the methods used in this study. Although areal and depth coverage of sampling locations was complete, samples from wells drilled after the wells presented in this report (Benson et al., 1983) have not indicated older groundwater, although as much as 700 m of saturated zone was penetrated by drilling these wells. Therefore, it seems likely that a biased sample is not the explanation for the absence of groundwater recharged during the Wisconsin glacial maximum.

A third possible explanation is that groundwater velocities are such that virtually all groundwater recharged before 17,000 yr. B.P. has moved out of the region sampled; this alternative seems unlikely because of the large area sampled (a 43-km downgradient reach). With the exception of two perched-water samples, groundwater age dates from the northern (upgradient) part of the area (Benson et al., 1983) have a range similar to groundwater age dates from the southern (downgradient) part. If a simple subsurface flow system is postulated, younger groundwater would be expected in the northern part, and older groundwater would be expected in the southern part; this distribution was not indicated.

3.2. Interpretation of oxygen-isotope data

As previously discussed, numerous questions need to be resolved and assumptions need to be made to extract paleoclimatic information from groundwater data. Calibration of an isotope temperature scale, that is valid under paleoclimatic conditions, is the first step. Calibration data need to be obtained from a location with a similar moisture source and similar approximate latitude as the location of interest. Calibration data shown in Table II were obtained from a site at $37.8°N, 106.9°W$, at an elevation of ~ 3400 m a.s.l. Periods of precipitation-sample integration were approximately monthly; the temperatures represent mean daily high and low, averaged for the period of integration. Monthly sampling was chosen to allow definition of seasonal isotopic variability. Annual sampling is sufficient if recharge is annually distributed; however, if recharge is associated with seasonal precipitation, the isotope signature of recharge will reflect only that season, not annual.

Eq. 6 may be compared with eq. 4 developed by Van der Straaten and Mook (1983), using the I.A.E.A. data base. The different intercept value most likely reflects the effect of re-evaporated moisture; that is, the air mass contains less ^{18}O than an air mass whose isotopic composition was developed only by Rayleigh condensation. It is assumed here that eq. 6 is more appropriate, because the calibration site more nearly approximates Nevada conditions than the I.A.E.A. sites; further, it is assumed that eq. 6 would result from temperature and precipitation data collected during the late Pleistocene and early Holocene at the same location.

It needs to be emphasized that error in estimate of the intercept (in eq. 6) leads to error in estimate of absolute paleotemperature only, not in temperature shift with time, if the meteoric-water line does not change during the time period. The magnitude of the error can be illustrated using eqs. 4 and 6 by re-arranging and solving for temperature (T). For any given $\delta^{18}O$, temperature calculated from eq. 4 is less by $\sim 5°C$ than if eq. 6 is used. Absolute temperatures calculated from either eq. 4 or eq. 6, or any similar equation derived from independent data sets require independent verification; trends, however, require only reasonable constancy of the meteoric-water line during the period.

Average $\delta^{18}O$ and eq. 9, rearranged from eq. 6, are used to calculate temperature for each 1000-yr. interval; results are given in Table III:

$$T = 1.75\delta^{18}O + 27.5 \tag{9}$$

TABLE III

Paleotemperatures calculated from groundwater $\delta^{18}O$

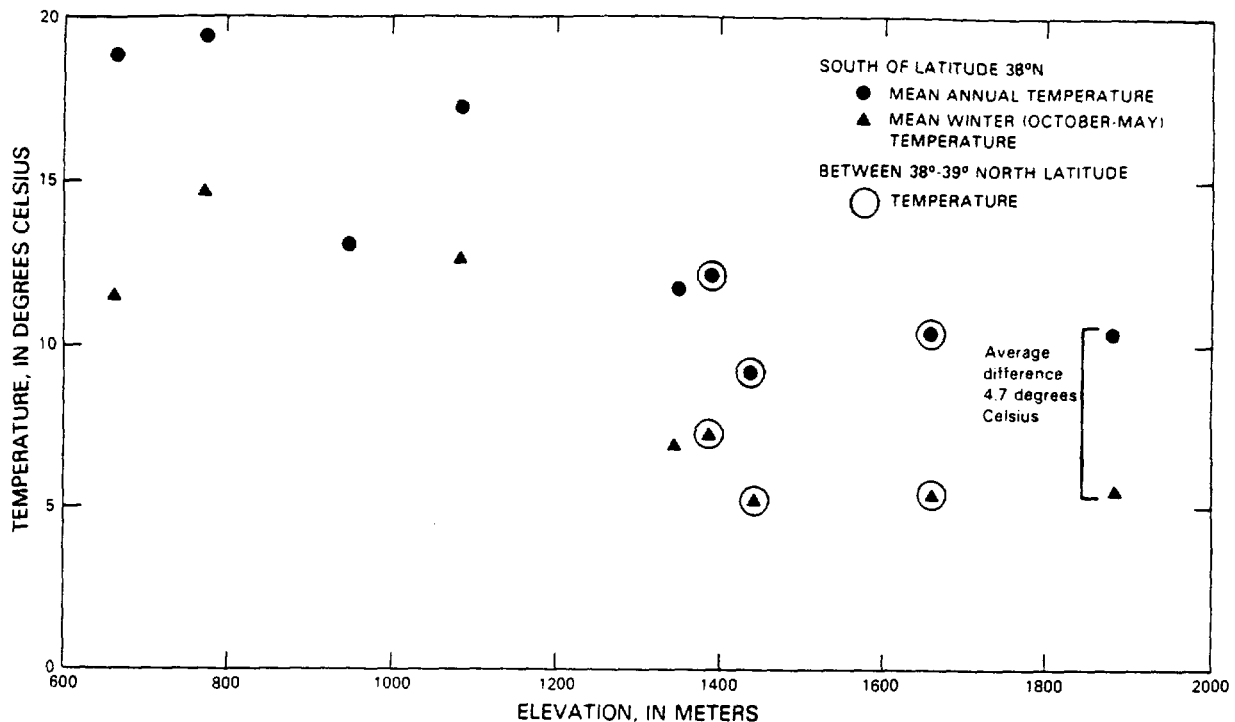| Time interval (10³ yr. B.P.)	Uncorrected for isotope shift during recharge		Corrected for isotope shift during recharge	
	average $\delta^{18}O$ (‰)	calculated temperature (°C)	average $\delta^{18}O$ (‰)	calculated temperature (°C)
9—10	−12.90	5.0	−13.20	4.4
10—11	−12.93	4.9	−13.23	4.4
11—12	−13.0	4.8	−13.3	4.2
12—13	−13.1	4.6	−13.4	4.0
13—14	−13.07	4.7	−13.37	4.1
14—15	−13.40	4.1	−13.70	3.5
16—17	−14.2	2.7	−14.50	2.1

Fig. 3. Temperature dependence on elevation in Nevada.

Temperatures are calculated from ground-water isotope data, both uncorrected and corrected, for the small positive isotope shift expected to occur during recharge. This isotope shift has the effect of ~ 0.5°C increase in the apparent temperature.

As groundwater represents only recharged water, the calculated temperature is representative of the recharge season only, not of the annual mean. For example, groundwater recharge at the calibration site yields a δ^{18}O that suggests a temperature of 0.7°C; whereas, recharging moisture known to contain only fall, winter and spring (hereinafter collectively called winter) precipitation suggests a temperature of —3.6°C. Annual mean temperature is ~ 0°C, whereas average winter temperature is ~ —3°C. Because the seasonal distribution of recharge in the past is unknown, two examples are used: annual average and winter average. Thus, the calculated temperature in Table III could represent annual or winter average.

The dependence of temperature on elevation for long-term weather stations south of 39°N latitude in Nevada is shown in Fig. 3. For the elevation range believed to be the source of the precipitation that resulted in recharge in the Amargosa Desert (1500—2100 m a.s.l.), little temperature dependence on elevation is apparent, although the data are admittedly sparse. The present mean annual and winter (October—May) temperatures are compared with the temperatures calculated from the groundwater δ^{18}O data in Table IV.

Assuming that the groundwater is an indicator of mean annual temperature, temperature vs. time is plotted in Fig. 4 (curve B) and compared to the mean annual temperatures predicted from data presented in Stuiver et al. (1978). These authors used ^{14}C dating of organic matter to predict deviations in mean annual temperatures from present-day values. The plotted values (curve A) were obtained by assuming the

TABLE IV

Comparison of calculated paleotemperatures with present-day temperatures in the precipitation source area

Time interval (10³ yr. B.P.)	Calculated paleotemperature (°C)	Annual temperature (°C)		"Winter" temperature of period October—May (°C)	
		present-day	difference	present-day	difference
9—10	4.4	10.0	−5.6	5.3	−0.9
10—11	4.4	10.0	−5.6	5.3	−0.9
11—12	4.2	10.0	−5.8	5.3	−1.1
12—13	4.0	10.0	−6.0	5.3	−1.3
13—14	4.1	10.0	−5.9	5.3	−1.2
14—15	3.5	10.0	−6.5	5.3	−1.8
16—17	2.1	10.0	−7.9	5.3	−3.2

same present-day mean annual temperature of 10.0°C used previously, then applying their computed deviations. The same general warming trend from 17,000 to 9,000 yr. B.P. is indicated by both curves; however, data for the Pacific northwest U.S.A. show a greater warming trend beginning ~ 12,500 yr. B.P. and ending ~ 10,500 yr. B.P., than that predicted from groundwater. These differences in predictions may be a result of the methods used: groundwater dispersion smooths or integrates short-term temperature variations; whereas, temperature inferred from vegetation integrates during the time represented by the life of the vegetation. When the plant dies, the integration period ends resulting in a shorter integration period than that for groundwater.

Because mean winter (October—May) temperature in southern Nevada is presently ~ 4.7°C less than mean annual temperature, less lowering of winter paleotemperature compared to present is predicted, if the δ¹⁸O in groundwater represents only winter precipitation.

To distinguish which of the two examples (shown in Fig. 4, curves B and C) may be more appropriate, paleoclimatic evidence for the area gleaned from fossil packrat middens was consulted. Spaulding et al. (1984) indicate that, during the late Wisconsin (12,000—10,000 yr. B.P.) precipitation during both the summer and winter

half-years was enhanced (compared to Wisconsin glacial maximum); whereas, during or shortly after the Wisconsin maximum (21,000—15,000 yr. B.P.), precipitation increase compared to present-day annual precipitation was concentrated during the winter months. Thus, in the unlikely event that increase in precipitation may be equated to increase in effective precipitation, the appropriate seasonal effective precipitation scenario changes from winter at 17,000 yr. B.P. to annual at 10,000 yr. B.P. This scenario decreases the computed mean annual temperature shift during 15,000—17,000 yr. B.P. from that computed, assuming annually effective moisture and allows the mean annual shifts for 9,000 to 10,000 yr. B.P. to remain the same. Specifically, under Spaulding et al.'s precipitation regime, the change in mean annual temperature from present would be ~ −3° to −4°C at 17,000 yr. B.P. to ~ −5° to −6°C at 10,000 yr. B.P., which contradicts the conventional view that the glacial—interglacial period represents a warming trend. This is also in disagreement with Spaulding et al.'s own estimate that mean annual temperature was −6° to −7°C below present at 18,000 yr. B.P. and −2°C below present at 12,000 yr. B.P. In fact, the groundwater isotope data are consistent with annual effective precipitation distribution at glacial maximum and a gradual shift to winter-effective precipita-
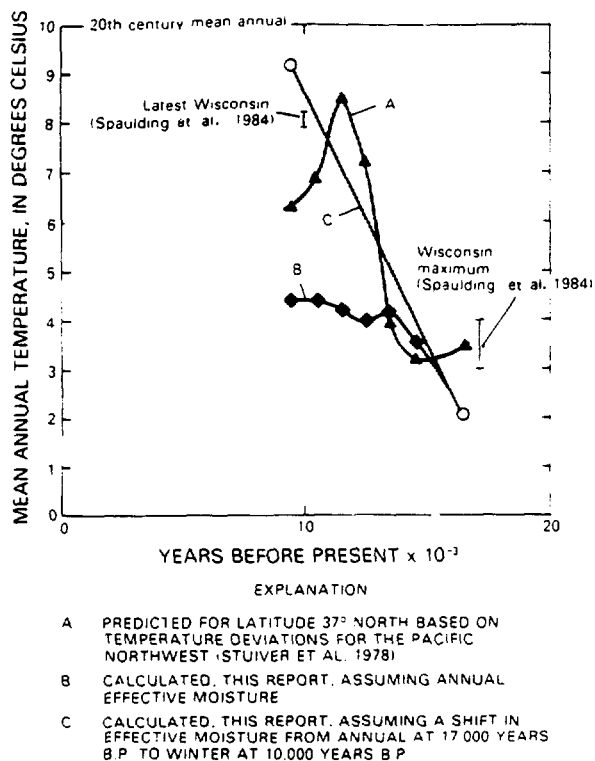
Fig. 4. Mean annual paleotemperatures near the west-central Amargosa Desert.

tion at 10,000 yr. B.P. This latter example is consistent with the anticipated effect on the hydrologic cycle of a glacial—inter-glacial transition: cooler temperatures decrease evaporation, resulting in summer moisture combining with winter moisture in recharge; warmer temperatures increase evaporation and effectively eliminate the summer moisture component of recharge. The effect of the aforementioned shift in effective moisture from annual at 17,000 yr. B.P. to winter at 10,000 yr. B.P. on mean annual temperature is illustrated in Fig. 4 (curve C).

4. Summary

Groundwater isotope data can provide information on the paleoclimatic conditions of recharge. Groundwater samples from the vicinity of the west-central Amargosa

Desert, Nevada, were analyzed to determine such paleoclimatic conditions. Chemical and ^{14}C isotopic results were used to determine the time distribution of recharge and to develop a hypothesis for the recharge mechanism. Dissolved inorganic ion composition was used to screen a large number of samples against inorganic carbon derived from calcium carbonate; this calcium carbonate would cause dilution of the radiocarbon, and would distort groundwater age. Areal distribution of groundwater ages indicated stream channels as line-sources for recharge; time distribution of groundwater ages showed that conditions were favorable for recharge only from 17,000 to 9,000 yr. B.P.

Oxygen-isotopic data were interpreted to derive the climatic regime under which recharge occurred. Slopes of $\delta^{18}O$ vs. temperature appear to be more or less invariant from place to place but intercepts appear to depend on location. A North American continent calibration site was chosen; this site was at the same latitude, but ~ 10° E of the study area. The temperature calibration developed at this site was used to determine climatic conditions of recharge for groundwater in the west-central Amargosa Desert.

Interpretation of computed temperatures requires a knowledge of the seasonal distribution of effective moisture. If winter moisture alone is responsible for recharge, the computed temperature represents a winter average temperature rather than an annual temperature. The most probable interpretation of recharge climate for west-central Amargosa Desert groundwater is that, compared to present, Wisconsin glacial maximum (~ 17,000 yr. B.P.) mean annual temperature was sufficiently decreased and summer precipitation sufficiently increased to result in a summer component of groundwater recharge. By late (terminal) Wisconsin (~ 10,000 yr. B.P.), annual temperature had increased enough to eliminate summer precipitation as a component of effective

moisture. This climate scenario results in annual-temperature departure from present of −8°C at 17,000 yr. B.P., and winter-temperature departure of −1°C at 10,000 yr. B.P., which is in agreement with the temperature evidence from fossil packrat midden studies.

Acknowledgment

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