Isotope hydrology of southern Nevada groundwater: Stable isotopes and radiocarbon
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Abstract. A new δ18O map of southern Nevada groundwater shows a systematic decrease in δ18O of ~5‰ from 36° to 39°N latitude. The variation is consistent with higher-latitude recharge following continuous flow paths along north-south trending graben valleys and systematically increasing in δ18O due to mixing with lower-latitude, higher-δ18O recharge. The data do not suggest that large masses of groundwater with unusually low δ18O values were recharged during the last pluvial period as suggested by previous workers. Many δ18O-δD pairs in groundwater indicate variable amounts of evaporation relative to the global meteoric water line. The precipitation rate and type (rain versus snow) for a given geographic area controls the extent of evaporation. A “model” δ18O value was calculated from evaporated groundwaters by subtracting postcloud evaporation. The distribution of these model δ18O values suggest new regional groundwater flow paths previously undocumented. Dissolved inorganic carbon in groundwaters collected from deep regional flow systems typically has low 14C concentrations (≤12% modern carbon). In contrast, groundwaters collected from the carbonate rock of the Spring Mountains and alluvium of Forty Mile Canyon have higher 14C contents, indicating more recent recharge. Because of nonlinear mixing of 14C, groundwater in the regional flow system likely acquired most of its observable 14C from mixtures of young, locally recharged groundwater.

1. Introduction

Isotope measurements complement hydrogeologic observations by providing direct evidence for groundwater sources and transport rates. This has been demonstrated for many simple groundwater flow systems [e.g., Stute et al., 1992; Ekwarzel et al., 1994]. However, as a flow system becomes more complex or more regional in extent, geochemical and isotopic data do not necessarily provide simple source and age interpretations. This is mainly due to (1) an increase in the number of recharge mixing-end members contributing to groundwater flow, which significantly complicates any groundwater age determination, (2) multiple water-rock reactions working simultaneously or at different times within the flow history, and (3) climate change and its effect on isotopic abundances. Interpreting isotopic data in regional groundwater flow systems requires an accountability of all atmospheric, surface, and subsurface (unsaturated and saturated zones) environmental effects controlling their abundance. The data interpretations also must be consistent with hydrogeologic data.

Regional groundwater flow in southern Nevada follows a complex system of interconnected, structurally controlled basins incorporating multiple lithologies with both fracture and matrix dominated flow [Blankenagel and Weir, 1973; Winograd and Thordarson, 1975; Thomas et al., 1996]. Groundwater can flow for distances in excess of a hundred miles before it discharges [Harrill et al., 1988], originating in alpine recharge areas and transcending extreme desert-basin environments. Southern Nevada has been the focal point of many detailed investigations related to potential risk of radionuclide migration from the Nevada Test Site (NTS; see work by Lacznik et al. [1996] for a review). The hydrogeologic framework established by Winograd and Thordarson [1975] designated northern boundaries of regional flow at ~38°N latitude. In contrast, other work [Eakin, 1966] recognized that recharge boundaries were further north for a similar system. However, later isotope hydrology and paleoclimate studies of groundwater [e.g., Claassen, 1986; White and Chuma, 1987; Winograd et al., 1988; Benson and Klieforth, 1989; Winograd et al., 1992; Thomas et al., 1996] mostly followed the hydrogeologic framework of Winograd and Thordarson [1975].

In this paper we have reevaluated these boundary conditions and some of the interpretations of interbasin flow. We present new isotopic data that extend north of 38°N latitude and combine them with existing data for groundwater in southern Nevada. Our analysis shows evidence for new flow paths previously undocumented. We furthermore suggest that almost all the groundwater isotopic variations observed can be explained by modern environmental conditions and groundwater mixing and that evidence of past climate preserved in the groundwater is limited only to areas of near closed-basin conditions. We suggest that these results have bearing on the interpretations for δ18O in the Devil’s Hole calcite. In addition, our analysis indicates that 14C concentrations in regional groundwater are predominantly controlled by mixing and water-rock interaction, which limits the usefulness of age calculations. Finally, our study serves to emphasize that consistent and defensible interpretations of isotopic data in regional flow systems such as southern Nevada require consideration of the entire flow system, not just parts of it.

2. Background

2.1. Hydrogeologic Framework

Structurally complex and geologically diverse rocks spanning Precambrian to Quaternary ages characterize the geology of
southern Nevada. Most of the sedimentary section is Precambrian- to Paleozoic-age marine rock of the North American western Cordillera miogeoclone [Stewart, 1980] (Figure 1a). Previous groundwater investigators have divided these rocks into hydrostratigraphic units [Winograd and Thordarson, 1975] (Figure 1b). The lower clastic aquitard forms the basal section comprising mostly Precambrian- to lower Paleozoic-age quartzite of low transmissibility. Overlying at nearly equivalent thickness, the lower carbonate aquifer consists of higher transmissibility limestones and dolomites of Paleozoic age. This aquifer is regional in extent and channels much of the regional groundwater flow in southern Nevada [Eakin, 1966; Winograd and Thordarson, 1975; Mifflin and Hess, 1979; Thomas et al., 1996; Dettinger et al., 1995].

Whereas Mesozoic-age rocks are poorly represented in southern Nevada, thick sequences of Tertiary-age volcanic deposits cover large areas of the lower carbonate aquifer. Late Tertiary, poorly sorted alluvial fan deposits of poor consolidation overlay the volcanic rocks in graben valleys. Matrix-dominated groundwater flow is probable in the alluvium but likely varies from matrix to fracture-dominated flow in the volcanic rocks. The lower carbonate aquifer transports groundwater by fracture or solution-channel flow paths [Winograd and Thordarson, 1975].

The topography in southern Nevada is marked by alternating basins and ranges oriented roughly north-south, which resulted from mid to late Tertiary crustal extension [Eaton, 1982; Hamilton, 1987]. This topography causes local groundwater flow systems to be superimposed onto regional groundwater flow. Local groundwater flow is common in horst-block ranges and is characterized by zones of perched groundwater and intermittent spring discharge [Winograd and Thordarson, 1975]. The regional groundwater flow system in southern Nevada has been divided into at least four independent units: (1) the White River, (2) the Ash Meadows, (3) the Oasis Valley, and (4) the Las Vegas Valley groundwater basins. Recently, workers [Laczniai et al., 1996] have subdivided these basins further, but we present our data in the context of the four original divisions for simplicity. Groundwater divides inferred from potentiometric data, lithostratigraphy, faults, and distribution of modern precipitation delineate these basins [Malmberg, 1965; Eakin, 1966; Blankennagel and Weir, 1973; Winograd and Thordarson, 1975].

The White River and Ash Meadows groundwater systems are underlain by continuous sections of the Lower Carbonate Aquifer (Figure 1a). According to Eakin [1966], the White River Basin is an interbasin flow system, whose inferred recharge origin is as far north as Long Valley at 40°N latitude. Flow is thought to be laterally continuous southward until it terminates at the Muddy River area north of Lake Mead. Of the 104,000 acre-feet (128,300,000 m³) of annual estimated recharge, nearly 78% originates north of the Pahranagat Valley. Most of this groundwater discharges as natural springs in the White River, Pahranagat, Kanes Springs, and Moapa valleys.

Regional groundwater flow in the Ash Meadows flow system is predominantly through the lower carbonate aquifer [Winograd and Thordarson, 1975]. Inferred recharge is from the Sheep Range to the east, the Spring Mountains to the south, and the Belted, Timpahute, and Pahranagat ranges in the north. Groundwater discharge in the Ash Meadows system occurs in part at the Ash Meadows spring discharge zone in the eastern Amargosa Desert (Figure 1a). The surface or near-surface exposure of the lower clastic aquitard has been used to denote basin boundaries in the Sheep Ranges, Spring Mountains, and Emigrant Valley [Winograd and Thordarson, 1975]. Its absence in the northeast part of the basin, combined with stable isotope evidence, lead to the interpretation that groundwater flows through the lower carbonate aquifer from the Pahranagat Valley into the Ash Meadows flow system [Winograd and Friedman, 1972]. The inferred western margin of the basin extends from the Belted Range southward to the Jackass Flats area.

To the west of this margin is the Oasis Valley (Forty Mile Canyon) groundwater basin, described by Blankennagel and Weir [1973], which consists largely of thick Tertiary volcanic sequences hosting caldera complexes (Figure 1a). Calderas are ringed fractures thought to downcut in excess of 13,000 feet (4000 m). The lower carbonate aquifer is not observed within the caldera areas but may lie deeply buried outside the ring fracture system. Southward flow through the caldera complexes is poorly understood. However, groundwater levels indicate that general subsurface flow is southwestward from Pahute Mesa toward the Amargosa Desert via Oasis Valley, Crater Flat, and western Jackass Flats. Up to 70% of the groundwater beneath Pahute Mesa is probably through-flow from Kawich Valley and Gold Flat [Blankennagel and Weir, 1973]. In these areas, regional groundwater flows through volcanic rocks, but possible recharge sources further north from the lower carbonate aquifer have not been investigated.

Malmberg [1965] characterized the 3000 square mile (7800 km²) Las Vegas Valley groundwater basin, which is bounded on the north by the southern Indian Springs and Three Lakes valleys and on the south by northern Ivanpah valley (Figure 1a). The aquifer system consists of a shallow, unconfined alluvial layer overlying a thicker artesian alluvial aquifer. Before groundwater overdrafting in the basin, natural artesian discharge occurred over a 75 square mile (194 km²) region that included the city of Las Vegas, Paradise Valley, and the Tule Springs areas. Predevelopment water levels indicated that groundwater recharges (~50,000 acre-feet (61,700,000 m³)) at the high-elevation (6000–12,000 feet (1800–3600 m)), high-precipitation (>20 inches/yr (51 cm/yr)) areas of the Spring Mountains and flows toward the eastern side of Las Vegas Valley [Malmberg, 1965].

2.2 Stable Isotope Studies

The variation of δ18O and δD values in natural waters is governed by the temperature of water evaporation and condensation, such that geographic variations are primarily controlled by differences in latitude and altitude [e.g., Epstein and Mayeda, 1953; Craig, 1961; Dansgaard, 1964]. For example, rain-out effects during continental migration of storm tracks depletes meteoric water of their heavy isotopes [e.g., Dansgaard, 1964; Sheppard, 1986]. This has been documented in some detail for California and Nevada [Ingraham and Taylor, 1991]. These stable isotope variations are preserved in shallow groundwater environments [Ingraham and Taylor, 1991] and have been used as tracers in regional groundwater flow systems [Gat, 1971], including areas of California [Davidson and Criss, 1993; Criss and Davison, 1996].

The global meteoric water line (MWL) of δD = 88.818O + 10 [Craig, 1961] reflects the isotopic abundance of water vapor condensed under isotopic equilibrium at different temperatures. The slope and the y intercept (i.e., deuterium excess) result from kinetic isotope fractionation between water and
water vapor over open ocean [Craig and Gordon, 1965]. Continental surface water and shallow groundwater also undergoes kinetic isotope enrichment during evaporation and enriches along trajectories with slopes between 2.0 and 6.0 on a $\delta$D-$\delta^{18}$O plot [Craig et al., 1963; Stewart, 1975; Barnes and Allison, 1988]. This is particularly relevant in desert areas where shifts in $\delta^{18}$O-$\delta$D pairs away from the global MWL result from evaporation of raindrops and shallow groundwater during recharge [Gat, 1971; Friedman et al., 1962; Stewart, 1975; Allison et al., 1984; Barnes and Allison, 1988].

Previous work in Nevada showed that rainfall collected in the Yucca Mountain area of southern Nevada defined a local MWL with the equation $\delta$D = 6.6$\delta^{18}$O − 7.2 [Benson and Kiefl, 1989]. However, when only snow samples were con-
sidered, the local MWL was δD = 8.0δ18O + 8.9, almost identical to the global MWL of Craig [1961]. Most of the precipitation events in southern Nevada that conform to the global MWL have mean isotopic values of less than −9.0‰ in δ18O and −60‰ in δD. Friedman et al. [1992] for desert regions of southeastern California observed a similar result.

Stable isotope values of modern precipitation in Nevada decrease with increasing latitude and altitude [Sheppard, 1986; Ingraham et al., 1991]. Similar geographic variations have been shown in δD values of hydrous minerals in weathered rocks [Taylor, 1974]. Different storm sources are also responsible for these regional-scale isotopic variations, since winter storms originating from the Pacific Ocean are isotopically depleted relative to isotopically heavier summer storms from the Gulf of California [Hale, 1974; Ingraham et al., 1991; Friedman et al., 1992].

Winograd and Friedman [1972] exploited the geographic variation of stable isotopes in precipitation as a means of quantifying groundwater recharge sources in the Ash Meadows flow system. They found that groundwater from the Pahranagat Valley in the north had lighter deuterium values than groundwater in the Sheep Range and Spring Mountains regions. From this observation they concluded that up to 35% of the groundwater discharge at Ash Meadows might be derived from the Pahranagat Valley region of the White River groundwater basin. Most of the remaining 65% was inferred to originate in the Spring Mountains and the Sheep Range (see Figure 1a). One of the assumptions in this interpretation was that stable isotope signatures in precipitation were constant over the timescales of the groundwater flow from recharge to discharge. The authors recognized the need to test this assumption, which raised subsequent questions regarding the isotopic record of Quaternary pluvial period recharge in the regional groundwater flow system [Winograd et al., 1988].

2.3. Radiocarbon Studies

Numerous studies using 14C to age-date groundwater have been published since the early 1960s (see work by Fontes and Gardnier [1979] and Mook [1980] for reviews). Most of these works focused on determining the sources and evolution of dissolved inorganic carbon (DIC) in the groundwater in order to correct apparent 14C ages. At least half of the carbon in groundwater originates as soil zone CO₂ produced by plant root respiration or from decomposition of humic material [Pearson and Hanshaw, 1970], while the remaining carbon is from dissolution of carbonate minerals in the unsaturated and saturated zones. Some exceptions have been reported [Mook, 1980; Andrews et al., 1994; Rose and Davison, 1996]. The distinctive 813C signatures between organic and inorganic carbon (−25 to −12‰ for plants, −2 to +2‰ for marine carbonates, and −12.0 to +2.0‰ for soil carbonates) have been used to characterize sources of DIC and estimate their relative contributions from water-rock reactions within the groundwater environment, particularly during recharge [e.g., Deines et al., 1974]. This isotopic correction method has led to the development of isotopic mass balance approaches [Pearson and Hanshaw, 1970; Mook, 1980], combined isotopic and aqueous chemical mass balance approaches [e.g., Fontes and Gardnier, 1979], and isotopic exchange models under open and closed conditions [e.g., Wigley, 1975]. These approaches are valid only where the carbon sources are uniform and the groundwater flow system is simple. A prerequisite for any groundwater age-dating study is to identify recharge zones, flow paths, and mixing zones before applying mass balance approaches for 14C corrections.

Previous studies of southern Nevada groundwater have shown that apparent 14C ages typically are >10 kyr [Grove et al., 1969]. In addition, δ13C measurements indicate significant water-rock interaction in the lower carbonate aquifer [Winograd and Pearson, 1976], and carbon isotope evolution along flow paths suggests both open and closed geochemical conditions [White and Chuma, 1987]. Furthermore, δ13C values of soil CO₂ and pedogenic carbonate decreased up to 8.0‰ with increasing pCO₂ content and elevation [Quade et al., 1989]. This reduces the utility of 14C corrections that depend on an isotopic mass balance approach since the source input is variable.

In this report we identify young recharge contributing to the regional flow in southern Nevada, using a limited radiocarbon database. We also estimate general flow rates where groundwater flow paths are well established.

2.4. Paleoclimate Studies

Considerable attention has been focused on climatic conditions during pluvial periods in the Great Basin of the southwestern United States [Smith and Street-Perrott, 1983; Spaulding et al., 1983]. The most striking evidence comes from the past occurrence of large closed-basin lakes, in particular Lake Bonneville and Lake Lahontan [Benson et al., 1990]. These lakes existed as far south as 37°N in Nevada and even further south in eastern California. Their existence required an increased flux of water into the Great Basin due to decreased rates of evapotranspiration, increased annual precipitation, and/or intensified snowmelt runoff.

Claassen [1986] observed that groundwaters in the alluvial and tuff aquifers of the eastern Amargosa Desert showed a decrease in δ18O with increasing model radiocarbon age. He interpreted the lower δ18O values to be pluvial period groundwater recharged ~17–10 kyr ago during a colder climate. White and Chuma [1987] observed similar stable isotope variations in the Oasis Valley but concluded that some groundwater flow originated from Pahute Mesa where elevation and temperatures effects could account for the lower stable isotope signatures. They further suggested that the persistent 5 per mil depletion in the deuterium excess of these groundwaters relative to the global MWL was due to increased marine humidity in storm source regions. Benson and Klegforth [1989] followed with a study of stable isotopes and uncorrected 14C ages of groundwater under Yucca Mountain. They concurred with the interpretations of Claassen [1986] and suggested that the recharge was synchronous with highstands of Lake Lahontan of northern Nevada. However, they also acknowledged that groundwater under Yucca Mt. could be a mixture of local recharge from Forty Mile Canyon and regional flow from as far away as Pahute Mesa. It should be noted that the study area of Claassen [1986] is directly downstream of Yucca Mountain, and that mixing of local and regional groundwater sources are equally valid for his area. In short, these paleoclimate studies assumed that groundwater with lower δ18O values that are depleted relative to modern local precipitation must have resulted from pluvial period recharge during cooler, climatic conditions.

Evidence for pluvial period vegetation patterns in the Great Basin have been gathered from pollen records and plant remains found in fossilized packrat middens [Martin and Mehlring, 1965; Spaulding et al., 1983]. The occurrence of differ-
ent types of plant remains coupled with radiocarbon dates has provided direct evidence for moisture and temperature regimes [Spaulding et al., 1983]. The Sonoran and Mojave deserts of southern California, Nevada, and Arizona experienced dramatic shifts in plant types during the Late Wisconsin period. At about 18 ka fossils suggest dry conditions and that winter precipitation was an important moisture source during the glacial maximum. From ~12–8 ka, fossils suggest increased temperatures and precipitation rates from southern summer storms [see Spaulding and Graumlich, 1986]. This paleoclimatic scenario is the opposite of those of Claassen [1986], White and Chuma [1987], and Benson and Klieforth [1989], suggesting instead low precipitation rates during the glacial maximum followed by higher rates in the Late Pleistocene to Early Holocene. If this latter period was warmer and wetter, then it should contribute to a higher $\delta^{18}O$ signature for any groundwater recharge.

Measurements of layered carbonate precipitated from groundwater in Devil’s Hole of Ash Meadows records variations in the $\delta^{18}O$ and $\delta^{13}C$ of groundwater over the past ~500 kyr [Winograd et al., 1988; Winograd et al., 1992; Coplen et al., 1994]. The last 50 kyr of the record are absent. Nevertheless, these stable isotope records vary approximately 2‰ in $\delta^{18}O$ and 1.5‰ in $\delta^{13}C$, with the highest $\delta^{18}O$ values correlating with the lowest $\delta^{13}C$. Winograd et al. [1988] correlated $\delta^{18}O$ maxima in the Devil’s Hole core with stage II and stage III maxima in the marine record and the Vostok Antarctic ice cores and concluded that the Devil’s Hole $\delta^{18}O$ variations were caused by changes in global temperature. Likewise, the covariances in $\delta^{13}C$ in the Devil’s Hole data were hypothesized to reflect changes in the density of vegetation in the groundwater recharge areas [Coplen et al., 1994]. The Devil’s Hole studies are set apart from the other groundwater paleoclimatic studies in that the carbonate layers record information on a much longer time scale and provide no direct evidence for climate in this region over the past 50 kyr.

3. Analytical Methods

This report combines existing isotopic data with new data generated at the Isotope Sciences Division of Lawrence Livermore National Laboratory (LLNL) and is compiled by Rose et al. [1997]. The existing data incorporates up to 30 years of stable isotope and radiocarbon analyses.

Almost half the LLNL samples were collected from pumped wells flushed for an appropriate length of time. Samples were collected from the main pump discharge or from water quality sampling taps. For the remaining well samples, groundwater were collected from open holes, equipped with a cover, using evacuated stainless steel bailers (~2 L total volume) vacuum sealed to a wireline tool. Additional samples were collected from flowing water at spring discharge areas. Field measurements for pH, temperature, and conductivity were measured with standard electronic instruments.

Samples collected for stable isotopes of $^{18}O$ and deuterium were collected untreated in 15–50 mL glass bottles with poly-seal caps to prevent evaporation and were analyzed using the CO$_2$ equilibration method [Epstein and Mayeda, 1953] and the zinc-reduction method [Coleman et al., 1982] followed by analysis on an isotopic ratio mass spectrometer. All stable isotope data are reported in the usual $\delta$ notation, where $\delta = (R/R_{STD} - 1)1000$, $R$ represents either the $^{18}O$/$^{16}O$, $D/H$, or $^{13}C$/$^{12}C$ ratio of the sample, and $R_{STD}$ is the isotope ratio of the SMOW (standard mean ocean water) or PDB (Pee Dee Belemnite) standard.

Samples for $^{13}C$ and $^{14}C$ were collected in a glass bottle (50–1000 mL) with either a polyseal cap or a Teflon-lined rubber septa cap to prevent exchange with atmospheric CO$_2$. Approximately four drops of a saturated HgCl$_2$ solution were added as a preservative, and most samples were kept refrigerated until analysis. The inorganic carbon was acid stripped under high vacuum and purged with an ultrapure carrier gas [Davisson and Velsko, 1994] in a manner similar to methods of McNichol et al. [1994]. The liberated CO$_2$ was then reduced to graphite on a separate vacuum line using a cobalt catalyst and hydrogen gas at a 570°C reaction temperature. All $^{14}C$ concentrations were determined on the accelerator mass spectrometer at LLNL. The $^{14}C$ results are reported as a percent of modern carbon (pmc) relative to a NBS oxalic acid standard [Stuiver and Polach, 1977]. Apparent ages are calculated using a half-life of 5730 years [Walker et al., 1989]. All data presented in subsequent figures and by Rose et al. [1997] are uncorrected pmc values.

4. Results and Discussion

4.1. Regional Groundwater $\delta^{18}O$ Variations

A $\delta^{18}O$ contour map of southern Nevada groundwater shows a 5.0‰ variation between latitudes 36° and 39°N (Figure 2). The $\delta^{18}O$ values average around ~16.0‰ east–west across Nevada at 39°N, although a single sample with this value was observed as far south as Cactus Flat (~37.5°N latitude). Groundwater between ~15.0 and ~16.0‰ was measured as far south as northern Pahranagat Valley (37°45’N) and Pahute Mesa (37°15’N); $\delta^{18}O$ values less than ~14.0‰ occur west of longitude ~116°W and north of 37°N. The ~14.0 contour at ~116°W roughly parallels the western terminus of the lower carbonate aquifer as delineated by Mifflin and Hess [1979]. In general, a good correlation exists between increasing latitude and decreasing $\delta^{18}O$ values in accord with previous studies in this region [e.g., Sheppard, 1986; Taylor, 1974].

In the Spring Mountains northwest of Las Vegas, $\delta^{18}O$ values decrease with increasing elevation, which is consistent with well-known altitude effects [e.g., Dansgaard, 1964; Smith et al., 1979; Ingraham and Taylor, 1991]. Springs and surface runoff from high elevations in the Spring Mountains are consistently ~14.0‰ and contrast higher $\delta^{18}O$ values observed at lower elevations in this region. The exception is ~14.0‰ groundwater that occurs in the natural artesian areas of Las Vegas Valley. The isotopic similarity between this artesian discharge area and springs in the Spring Mountains recharge area supports the earlier hydrologic interpretations of Malmberg [1965].

In the White Pine Range, west of the White River Valley, $\delta^{18}O$ values increase with elevation, in an apparent reversal of the altitude effect. A similar effect is observed in the Pintwater, Groom, and Sheep Ranges. Samples from higher altitudes are from shallow wells and springs and represent groundwater in local flow systems. Samples collected in the adjacent valleys are from alluvium or deeper aquifer units that are part of regional flow systems. This latter groundwater has a lower $\delta^{18}O$ and is likely transported from higher latitudes.

An isolated pocket of groundwater with $\delta^{18}O$ values ~13.0‰ occurs in Forty Mile Canyon (37°N, 116.3°W; Figure 1a). These values are higher than groundwater measured in the adjacent areas of Crater Flat and Jackass Flats and are similar to modern precipitation measured by Benson and Klie-
This suggests that groundwater in Forty Mile Canyon is locally recharged [Blankennagel and Weir, 1973; White and Chuma, 1987]. It has previously been suggested that the lower δ18O values in the Crater Flat and Jackass Flats areas represent Late Wisconsin–age paleowaters that were recharged during cooler climatic conditions [Claassen, 1986; White and Chuma, 1987]. However, it is equally plausible that they represent groundwater that has been transported along regional flow paths from recharge areas much further north. The high elevation mountains of central Nevada (~39°N latitude) receive relatively large amounts of annual precipitation during the cold winter months. Spring snowmelt in these areas provides a critical source of recharge to the regional flow system with the necessary hydrologic head to maintain south-
ward flow. Furthermore, with the exception of the White River flow system, the hydrologic connection between central Nevada recharge areas and the NTS has largely been unexplored. The geographic continuity of groundwater with low δ18O values from central Nevada to the NTS (see Figure 2) suggests a hydrologic interconnection that is consistent with interpretations from previous hydrogeologic work in this region [Blankenagel and Weir, 1973]. In particular, the contours have been drawn to suggest that low δ18O values are continuous between Railroad Valley groundwater and Padute Mesa. Although our data are limited in regions connecting these two areas, we justify the contouring by citing data from Roth and Campana [1989]. They report δD values to as low as −118‰ for groundwater south of Railroad Valley.

4.2. δD-δ18O Evaporation Effects

Many of the southern Nevada groundwaters have isotopic values that plot to the right of the global MWL on a δD-δ18O plot (Figures 3a and 3b). The main exceptions are groundwater in Las Vegas Valley, the southern Sheep Range, southern Three Lakes Valley, and the Spring Mountains (also see Figure 2), all areas which encompass the Las Vegas Valley groundwater basin. Modern precipitation in Las Vegas Valley and southeastern California that have δ18O values less than −9.0‰ generally adheres to the global MWL [Benson and Klieforth, 1989; Friedman et al., 1992]. In light of this fact, it is notable that many other groundwaters in southern Nevada with δ18O values < −9.0‰ plot to the right of the global MWL (Figures 3a and 3b), most notably groundwaters in the Ash Meadows, Oasis Valley, and White River groundwater basins.

Two explanations have been considered for the shifts away from the MWL. White and Chuma [1987] proposed that groundwater in Oasis Valley and Jackass Flats originated from pluvial period recharge events (i.e., 10–18 kyr), whose moisture sources were developed under different conditions (higher humidity) than present day precipitation. This interpretation implies that the local MWL for southern Nevada changed overtime. This interpretation originated from the model of Merlivat and Jouzel [1979], who suggested that the deuterium excess of the global MWL decreases during maximum glaciation, and the MWL described by Pleistocene-age groundwater should therefore reflect this shift. We rule this out as a major control because groundwaters that are isotopically shifted from the global MWL show a geographic dependence that does not correlate to age information gleaned from radiocarbon data presented later in this paper.

An alternative explanation for the shift is the process of evaporation. This includes evaporation of raindrops on their descent to the ground [Friedman et al., 1962], a phenomenon well documented in semiarid to arid regions [Gat and Dansgaard, 1972]. In addition, evaporation of surface runoff and shallow groundwater can also occur [e.g., Gat and Dansgaard, 1972; Allison et al., 1984; Ingraham and Taylor, 1991]. A less documented, but potentially significant process is the isotopic enrichment that occurs in aging winter snowpack during recrystallization and partial melting [Friedman et al., 1991].

Most precipitation derived from maritime sources has isotopic values that lie along the global MWL [Craig, 1961; Craig and Gordon, 1965; Gat and Carmi, 1970]. This has been observed in precipitation derived from the Pacific Ocean, the Gulf of Mexico, and the Gulf of California [Ingraham and Taylor, 1991; Friedman et al., 1992; Davison and Criss, 1993; Cortes and Farvolden, 1989; Benson and Klieforth, 1989], areas relevant to our discussion. On the basis of results from Benson and Klieforth [1989], it is reasonable to assume that snowfall in central Nevada also adheres to the global MWL, since vapor exchange has a negligible affect on snowfall. The first-order conclusion from Figures 3a and 3b is that the difference in the deuterium excess between groundwater and precipitation results from evaporation effects.

Recall that most stable isotope values of groundwater in the Las Vegas Valley basin adhere to the global MWL, while groundwaters in the Ash Meadows, Oasis Valley, and White River groundwater basins lie to the right of the MWL (Figures 3a and 3b). Fundamental differences between the Las Vegas Valley groundwater basin and the other basins include (1) the precipitation rates in the recharge areas, (2) the size of the recharge areas, and (3) the length of groundwater flow paths.
The main recharge area of the Las Vegas Valley basin is the Spring Mountains [Malmberg, 1965], which seasonally averages >20 inches (51 cm) of precipitation, mostly as snow. In the Ash Meadows, Oasis Valley, and White River basins recharge is derived over a broad area where precipitation rates typically average between 6 and 18 inches (15 and 45 cm) per year (Figure 4). The deuterium shift in these latter basins correlates to their lower precipitation rates and is similar to geographic/orographic variations in deuterium excess observed in Israel [Gat and Dansgaard, 1972]. Most groundwater in Las Vegas Valley flows under confined and artesian conditions along relatively short flow paths. In contrast, groundwater in the larger regional flow systems ranges from confined to unconfined, in many cases is not under artesian conditions, and has flow components that have been transported over distances in excess of 100 miles [Eakin, 1966; Blankenagel and Weir, 1973; Winograd and Thordarson, 1975]. Therefore, in the case of the Las Vegas Valley groundwaters, high precipitation rates, rapid recharge, and near confined aquifer conditions may limit the postsnowmelt evaporation.

### 4.3. Regional Groundwater Model \( \delta^{18}O \) Variations

The \( \delta^{18}O \) variation in Figure 2 results from two different effects: (1) geographic variations in \( \delta^{18}O \) of annual precipitation and (2) subsequent evaporation effects, whose extent also varies geographically. If we make the assumption that all groundwater in southern Nevada originated as precipitation that initially conformed to the global MWL, then we can correct the isotopic values for postcloud evaporation effects. We can accomplish this by projecting the evaporated data points back to the global meteoric water line along a constant slope of 3.5 on a \( \delta^{2}D-\delta^{18}O \) plot. The slope of 3.5 was selected to represent an approximate average value for evaporation trajectories in surface and groundwaters in arid environments [e.g., Craig et al., 1963; Barnes and Allison, 1988]. We acknowledge that this slope can vary depending on temperature, humidity, moisture content of the soil, and depth to water [Craig et al., 1963; Stewart, 1975; Barnes and Allison, 1988]. The basis for the correction is an attempt to benchmark geographic variations in the \( \delta^{18}O \) values as they reflect different moisture sources only. The benchmarking is achieved by assuming that all isotopic data originated on the MWL and meet the condition

\[
\delta D = 8\delta^{18}O + 10
\]

and that the applied isotopic correction is simply the difference in measured values and the pre-evaporated values,

\[
\delta_{in-3.5} = \delta_{in} - \Delta \delta,
\]
where $\delta_{m-3.5}$ is the model isotopic value corrected along a slope of 3.5 and $\delta_m$ is the measured value. Since the following condition must also be met,

$$\Delta \delta D = 3.5 \Delta \delta^{18}O$$

a correction formula can easily be derived for pre-evaporation $\delta^{18}O$ values:

$$\delta^{18}O_{m-3.5} = \delta^{18}O_m - [(8\delta^{18}O_m + 10) - \delta D_m]0.22$$

The geographic distribution of these model $\delta^{18}O$ values (shown on a contour map in Figure 5) differs from actual values of Figure 2 in minor ways. The model $\delta^{18}O$ values are more depleted than actual values, particularly in the northern region. These low $\delta^{18}O$ values reflect unevaporated mean annual precipitation, although many values have a high uncertainty since exact evaporation slopes are not known. The amount of uncertainty is a function of the length of the correction line for a particular point. Nevertheless, relative differences among the model $\delta^{18}O$ values are instructive and provide further insight.

The largest difference between Figures 2 and 5 is in the northwestern part of southern Nevada, where much lower model $\delta^{18}O$ values occur in the groundwater of Gold and
Cactus Flats. Note that the single sample of $-16.0^{\circ}{\text{o}}$ in Cactus Flat ($-37.5^{\circ}{\text{N}}$ latitude) from Figure 2 is now approximately $-18.0^{\circ}{\text{o}}$ but still does not conform spatially to adjacent groundwaters in Figure 5. This will be discussed further in a later section. The model values in groundwater beneath northwestern Pahute Mesa are similar to Gold and Cactus Flats, while actual $\delta^{18}{\text{O}}$ values (Figure 2) suggest a similarity to Kawich Valley. Groundwater flow from both these areas have been previously suggested [Blankennagel and Weir, 1973]. Note that a component of this low $\delta^{18}{\text{O}}$ groundwater plume continues southward into eastern Amargosa desert where it meets the higher $\delta^{18}{\text{O}}$ groundwaters of the Ash Meadows discharge area.

The $-17.0^{\circ}{\text{o}}$ groundwater in Figure 5 represents one mixing-end-member for groundwater recharge in the northern White River groundwater basin. Groundwater $\delta^{18}{\text{O}}$ values progressively increase toward the south, reaching values $>-14.0^{\circ}{\text{o}}$ because of mixing with local groundwater with higher $\delta^{18}{\text{O}}$ values. In the White River groundwater basin this pattern is confined to a strip less than 50 miles (80 km) wide that terminates in Pahranagat Valley. This distribution is consistent with conclusions of Eakin [1966], who showed that interbasin flow is confined to a narrow strip from the Long Valley in the north to the Moapa Valley in the south. A similar plume of low $\delta^{18}{\text{O}}$ groundwater stretches from north of Emigrant Valley to eastern Yucca and Frenchman Flats. These plumes are separated by a region of higher $\delta^{18}{\text{O}}$ values that extends from the southern Sheep Range north-northeast to the Grant Range at approximately 38°N. Note also that the reverse $\delta^{18}{\text{O}}$-elevation trend in the White Pine Range seen in Figure 2 no longer exists in Figure 5.

The model $\delta^{18}{\text{O}}$ values of Figure 5 tend to accentuate features observed in Figure 2. In particular, the model $\delta^{18}{\text{O}}$ values emphasize the fact that low-$\delta^{18}{\text{O}}$ groundwater migrates south from interbasin flow and strongly contrasts with higher-$\delta^{18}{\text{O}}$ values in groundwater derived from local precipitation such as the Sheep Ranges.

Wingrod and Friedman [1972] suggested that the stable isotope signatures in the Ash Meadows discharge area could be modeled as a mixture of groundwater from the Pahranagat Valley area with groundwater from the Spring Mountains. For this interpretation to be correct, it would require groundwater from the Pahranagat Valley to flow southwesterly beneath the western boundary of the White River basin. The ridge of higher $\delta^{18}{\text{O}}$ values west of the Pahranagat Valley, observed in Figures 2 and 5, do not support this hypothesis, although samples from prominent basins such as Tikaboo Valley are lacking.

On the basis of the available data, a more consistent hypothesis is that low-$\delta^{18}{\text{O}}$ groundwater from the Railroad Valley area flows south through Emigrant Valley and eastern Yucca and Frenchman Flats. This interpretation requires a continuous flow path from Railroad Valley through the lower elastic aquitard on the southern side of Emigrant Valley. Alternatively, low-$\delta^{18}{\text{O}}$ groundwater flowing south from the Pahute Mesa area through the caldera complexes could provide the lower-$\delta^{18}{\text{O}}$ component to the Ash Meadows springwaters. On the basis of the distribution of groundwater isotopic compositions in Figures 2 and 5, the two latter hypothesis for sources of lower-$\delta^{18}{\text{O}}$ groundwater in the Ash Meadow spring waters are more plausible than underflow from Pahranagat Valley.

Finally, it should be noted that zones of low annual precipitation dominate much of southern Nevada (Figure 4). Many of these areas have precipitation rates below 10 inches (25 cm) per year, and local recharge is therefore limited. As a result, most of the groundwater in these arid regions is probably derived from higher-precipitation areas in central Nevada where stable isotope values are more depleted.

4.4. Timing of Groundwater Flow

4.4.1. Distribution of $^{14}{\text{C}}$ data. A large radiocarbon data base exists for groundwaters in southern Nevada (Figure 6). However, the available data is densely clustered in localized areas around the Spring Mountains and Las Vegas Valley, the Ash Meadows discharge area, the NTS, and Pahranagat Valley. Data are limited in the White River flow system and absent in its recharge areas. There is essentially no data available from areas where most of the low-$\delta^{18}{\text{O}}$ recharge is inferred to occur. The majority of available data represent groundwater in the Lower Carbonate Aquifer with the exception of those samples from immediately north of the NTS, under Pahute Mesa and in the Oasis Valley. Contour lines in Figure 6 distinguish between $^{14}{\text{C}}$ values $>25$ pmc and those $<25$ pmc, except in the Spring Mountains and the Forty Mile Canyon region where contours for 50 and 75 pmc are also included. In the Spring Mountains the $^{14}{\text{C}}$ values systematically increase with elevation, showing the significant loading of modern recharge at high elevations in this area. A similar pattern occurs in the Forty Mile Canyon region. These data highlight the value of $^{14}{\text{C}}$ measurements for distinguishing recharge areas in southern Nevada. Note the general correlation between high-$^{14}{\text{C}}$ contents of groundwater in Figure 6 and high precipitation rates in Figure 4.

There are also isolated cases of high-$^{14}{\text{C}}$ groundwater in the Pahranagat Valley spring discharge area, in springs from eastern Cactus Flat, and in Oasis Valley (Figure 6). These particular cases are groundwater that is extensively evaporated (see Figures 2 and 5), and higher $^{14}{\text{C}}$ concentrations likely result from exchange with atmospheric CO$_2$ [e.g., White and Chuma, 1987]. One exception is a groundwater sample from a well (WT-3 of Rose et al. [1997]) northeast of the NTS that shows no evaporative signature in its stable isotopes but has a $^{14}{\text{C}}$ value of 27 pmc. Another exception is well U12s on Pahute Mesa (not shown on map), which is a shallow emplacement hole completed in granitic rock. It has a 100 pmc $^{14}{\text{C}}$ content and a $\delta^{18}{\text{O}}$ value of local precipitation ($-12.6^{\circ}{\text{o}}$). Most remaining groundwater samples have very low $^{14}{\text{C}}$ values that lie between <1 to 12 pmc, and most of these occur in the Lower Carbonate Aquifer.

Groundwater in the northern part of the White River groundwater basin has $^{14}{\text{C}}$ values between 2 and 14 pmc (Figure 6). Given that this area is presumably close to the recharge areas for this basin, these $^{14}{\text{C}}$ values appear to be unusually low. Groundwater further south in the White River basin has $^{14}{\text{C}}$ values between 6 and 11 pmc, with the exception of the highly evaporated waters. Therefore we do not observe a systematic decrease in the $^{14}{\text{C}}$ concentration of groundwater going from recharge to discharge areas of this basin, even though the $\delta^{18}{\text{O}}$ values clearly show southerly interbasin flow (e.g., Figure 5). The lack of a systematic decrease in the $^{14}{\text{C}}$ along these flow paths suggests an unaccounted for complexity, and two questions arise from this observation: (1) Why are the $^{14}{\text{C}}$ values low near the recharge areas of the White River flow system?, and (2) Why do the $^{14}{\text{C}}$ values remain essentially constant along its subsequent flow to the Pahranagat Valley springs discharge area?
4.4.2. Transport model of the White River flow system. To address the questions posed above, we propose a simple model for $^{14}$C evolution in the White River groundwater flow system. We infer that three separate processes govern the $^{14}$C evolution in this system: radiogenic decay, water-rock interaction, and groundwater mixing. Mixing rates are calibrated to changes in groundwater $\delta^{18}$O values from north to south in Figure 5, similar to methods of Kirk and Campana [1990]. Continuous mixing along flow paths is typically missing from water-rock reaction models that simulate carbon isotope mass transport. However, in the White River flow system, the persistence of low-$^{14}$C groundwater and the increase in $\delta^{18}$O along flow paths suggest that locally derived groundwater with a higher $^{14}$C content continually mixes with the low-$^{14}$C regional flow system. The $^{14}$C value of a mixed water depends on both the $^{14}$C content and DIC concentration of each mixing component, according to the expression

$$^{14}C_{\text{mix}} = \frac{X_1C_1^{14}C_1 + X_2C_2^{14}C_2 + \cdots}{\sum X_iC_i}$$

Figure 6. The higher $^{14}$C concentrations in groundwater of the Spring Mountains and the Forty Mile Canyon area delineate zones of young recharge. Most other groundwaters were collected from regional flow paths in the lower carbonate aquifer, which have lower $^{14}$C due to radioactive decay, water-rock interaction, and mixing.
where $C_i$ is the inorganic carbon concentration, and $X_i$ is the fraction of each mixing end-member. Mixing between waters with similar DIC concentrations and different $^{14}$C contents is nonlinear, resulting in calculated $^{14}$C ages biased toward younger end-members.

For our model we assume groundwater flow occurs under steady state conditions and has a uniform velocity along a flow path from Long Valley to the Pahrmanagat Valley Springs discharge area. The remainder of the flow system south of Pahrmanagat Valley is not included in this model. The total geographic area is estimated at 5900 square miles (15,300 km$^2$, ~78% of total basin area) with an average aquifer thickness of 3000 feet (900 m). Total annual groundwater flow is 81,120 acre-feet (100,100,000 m$^3$), or 78% of the total 104,000 acre-feet (128,300,000 m$^3$) as suggested by Eakin [1966]. We assume a 7% average effective porosity. Drill cores from the Ash Meadows flow system suggested a similar porosity for the lower carbonate aquifer, although effective fracture porosity may be a fraction of 1% [Winograd and Thordarson, 1975, pp. C17, C28]. Nevertheless, if 7% porosity is used as an upper limit, then the groundwater flow rate is estimated to be ~10 kyr along the path length of ~310 km, or 31 m/yr. This rate is consistent with previous estimates for the lower carbonate aquifer [Winograd and Thordarson, 1975].

The reaction rate of the groundwater DIC with the lower carbonate aquifer is a function of the surface area, grain size, and water chemistry. In a fracture flow system the surface area will be a function of fracture aperture. In simplifying this problem, we assume that fractures have the shape of a circular capillary whose surface area is a function of its radius. If the radius is 2.5 mm for instance, then the total surface area of the pore space would be $7.8 \times 10^{18}$ cm$^2$.

We assume that the initial $^{14}$C content in the upper most part of the flow system is 80 ppm. No data are available for the recharge areas, which likely comprise carbonate rocks surrounding Long Valley. However, the use of 80 ppm is consistent with observed $^{14}$C content of high-elevation recharge in the Spring Mountains (Figure 6) and with methods of Vogel [1970]. We also assume a constant DIC concentration along the flow path. Detailed analysis of the DIC and $^{3}$H along this groundwater flow may reveal variable calcite/dolomite saturation conditions that result in precipitation and dissolution reactions [e.g., Thomas et al., 1996]. Because of our limited data for this part of the flow system, we assume the entire flow path undergoes a simultaneous dissolution-precipitation reaction, resulting in no net gain or loss of DIC over the entire flow path length. Therefore a fixed amount of carbonate rock must dissolve into solution in order to reduce the dissolved $^{14}$C from 80 to 7 ppm between the recharge areas and the northern most data point in Figure 6 (~60 miles, or 97 km).

In order to introduce mixing, we divided the flow system into three equal lengths from the top of Long Valley to the Pahrmanagat Springs discharge area. Note that the $^{18}$O contours of ~17.0, ~16.0, and ~15.0 in Figure 5 approximately correspond to these three equal divisions. We assume that no change in $^{18}$O occurs in the first third, since much of the recharge areas between 39° and 40°N latitude have $^{18}$O values around ~17.0‰. However, we introduce a local mixing component of ~17.0 in the upper third of the flow system for purposes of modeling the $^{14}$C change. The change in the $^{18}$O values in the second and third part of the flow system suggests a constant mixing rate of local recharge. We assume that the local groundwater recharge has a $^{18}$O value of ~14.5‰ in the mid region of the flow path and ~13.5‰ in the lower portion. We assume that all local flow mixtures have a $^{14}$C content of 80 ppm. Therefore if we use a 10% addition of local groundwater for every 1000 years of flow, then by mass balance (i.e., 50% ~ 17.0, 25% ~ 14.5, and 25% ~ 13.5) we calculate the $^{18}$O value for final groundwater (~15.5‰), which is similar to that measured in the Pahrmanagat Springs area. Results using this mixing rate, combined with simultaneous radioactive decay and water-rock interaction, are shown in Figure 7. The final $^{14}$C values were obtained by calculating at each iteration radioactive decay, water-rock reaction, and mixing, respectively. Given the model assumptions, it is clear from Figure 7 that nearly 100% of the DIC must exchange with the carbonate rock within the first 30 km of flow (or ~100%/1000 years) in order to reproduce observed $^{14}$C of the White River flow system in Figure 6. Given the assumed porosity, surface area, and flow rate, a reaction rate of only $2 \times 10^{-14}$ mmol/cm$^{2}$ s is required. This reaction rate is far below experimental determinations at slightly lower $p$H values (e.g., $1 \times 10^{-8}$ mmol/cm$^{2}$ s at $p$H = 7.0; see work by Plummer et al. [1978]). This suggests that reaction rates in our model are a minimum. Therefore this model may be representative of the lowest flow rates and lowest reaction rates. A faster groundwater flow requires a lower porosity, a lower surface area, and a faster reaction rate.

Uncertainties in the boundary conditions should be addressed. For instance, if the effective porosity of the lower carbonate aquifer is lower than assumed, for example 1%, then shorter groundwater residence would be required. With this porosity a linear flow rate of ~200 m/yr, and a water-rock reaction rate of $1.4 \times 10^{-13}$ mmol/cm$^{2}$ are needed. Both are plausible values. Additionally, if initial $^{14}$C values for the regional and local flow components were lower than assumed, then slower water-rock reaction rates would be required.
4.4.3. Further effects of groundwater mixing. In southern Nevada, regional groundwater flow occurs through the fractured Lower Carbonate Aquifer or volcanic rocks and connections between topographically closed basins, as defined by hydraulic heads in wells [Winograd and Thordarson, 1975]. In the Ash Meadows basin, groundwater, migrating great distances from disparate sources, mixes and discharges into a single area. Groundwater flow beneath Pahute Mesa and into the Oasis Valley may originate further north than Railroad Valley. In all, the isotope contour maps presented here (Figures 2 and 5) suggest that groundwater in southern Nevada is highly mixed, integrating several flow paths from different aquifers over the entire flow system. The exception would be the Las Vegas Valley groundwater basin where groundwater mixing is minimized by the geographically limited flow. Longer flow paths with greater mixing can have a profound impact on the $^{18}O$ abundance in the groundwater (e.g., Figure 7).

In cases where groundwater of disparate $^{14}C$ contents mix in the subsurface, the resulting mixed $^{14}C$ age will in almost all cases be weighted toward the younger groundwater. For example, in the Ash Meadows discharge area, a portion of this groundwater is transported over a hundred miles from the north and northeast, while the remaining portion is recharged by the Spring Mountains [Winograd and Friedman, 1972]. Groundwater in the recharge area of the Spring Mountains has distinctly higher $^{14}C$ contents than regional groundwater flow (Figure 6). This component mixed into the Ash Meadows discharge will contribute the majority of the inorganic $^{14}C$ observed in these springs. It is therefore possible that much of the groundwater discharge in Ash Meadows has no $^{14}C$ and that the small quantities measured in the spring water are entirely from Spring Mountain contributions.

4.4.4. Paleoclimate and groundwater transport. On the basis of the hydrogeologic constraints of aquifer volume and discharge in the White River flow system, groundwater transport rates from upper Long Valley to the Pahranagat Springs region probably require <10 kyr. The coupled $^{14}C-\delta^{18}O$ model presented in this paper is consistent with this argument. This model is also applicable to and consistent with $^{14}C$ observations in the Las Vegas Valley flow system. On the basis of these results we conclude that both the White River and Las Vegas Valley systems are dominated by Holocene-age groundwater. Pleistocene groundwater with unusually low $\delta^{18}O$ values may occur locally in low-permeability zones, but it has not been observed in these basins. However, isolated pockets of Pleistocene age groundwaters with unusually low $\delta^{18}O$ have been observed in the Central Valley of California [Davisson and Criss, 1993] and in the Honey Lake Basin in Northern California and Nevada (T. P. Rose, Isotopic characterization of closed-basin groundwater flow, Honey Lake basin, California and Nevada, submitted to Water Resources Research, 1998, hereinafter referred to as submission, 1998). These observations suggest that late Pleistocene precipitation with depleted $\delta^{18}O$ values probably occurred in southern Nevada as well.

Large-scale groundwater systems such as the Oasis Valley and Ash Meadows basins have superimposed local and regional flow components that likely mix, forming a spectrum of groundwater ages and stable isotope values. A key question is whether recharge during the last glacial maximum contributed a sufficient recharge volume to cause large stable isotope perturbations in slower moving parts of the groundwater systems in southern Nevada. At this time we conclude that the groundwater $\delta^{18}O$ variations in Figures 2 and 5 primarily reflect regional patterns of $\delta^{18}O$ in modern precipitation. This pattern is reflected in $\delta^{18}O$ variations of Holocene groundwater (i.e., White River and Las Vegas Valley systems). However, at least one case was found where an unusually low $\delta^{18}O$ sample, collected from Cactus Flat (37.5°N latitude), indicated a remnant pocket of paleowater. In this example the groundwater $\delta^{18}O$ value is unusually low with respect to adjacent groundwaters in both actual and model maps (Figures 2 and 5). Further sampling and analyses in this region may elucidate the total geographic extent of this possible paleogroundwater. It is notable that similar pockets of groundwater were observed in Honey Lake Basin area in northeastern California (T. P. Rose et al., submission, 1998). Both the Cactus Flat and Honey Lake Basins are nearly closed basins. In the Honey Lake Basin paleogroundwater is found at depth, although mixing with Holocene water occurs in the discharge areas.

Rather than appealing to pluvial period recharge mechanisms to explain isotopic variations in the Oasis Valley groundwater system, we suggest that mixing between local recharge and regional flow is likely more important. In Figure 8 we have plotted $^{14}C$ against all the measured $\delta^{18}O$ values (with corresponding $^{14}C$ data) of the Oasis Valley groundwater system. In addition, we have computed and plotted mixing lines between groundwater from NTS wells J-12, J-13 and UE19c (mixing model 1), and U12s and UE19c (mixing model 2). These groundwater were chosen to represent mixing between a young, locally recharged end-member with an older, regionally recharged end-member. The data distribution in Figure 8 suggests that mixing combined with $^{14}C$ decay will explain these isotopic variations, including data originally interpreted to result from paleoclimate. Paleoclimate and mixing effects are two opposing hypotheses for explaining isotopic data. Given the regional pattern of $\delta^{18}O$ and $^{14}C$ values, we suggest that multiple recharge areas within a broad geographic area, followed by mixing during long range transport, can account for most of the isotopic variation observed in southern Nevada.

Figure 8. Groundwater $\delta^{18}O$ and $^{14}C$ in the Oasis Valley groundwater basin can be interpreted as resulting from mixtures of regional groundwater flow with depleted $\delta^{18}O$ values and local groundwater with higher $\delta^{18}O$. Model mixing curves are presented between wells (1) UE-19c and U-125s (solid line) and between well (2) UE-19c and the J wells in Forty Mile Canyon (dashed line). Abbreviations OVGMS, LCA AM GWS, and LCA WR GWS stand, respectively, for Oasis Valley groundwater system, Lower Carbonate Aquifer Ash Meadows groundwater system, and Lower Carbonate Aquifer White River groundwater system.
groundwater. The dynamic nature of these groundwater systems is not consistent with large pockets of static paleowater.

Mixing may also explain part of the long-term variations observed in $\delta^{18}O$ of the Devil’s Hole carbonate samples of Winograd et al. [1988, 1992]. Groundwater sources to the Ash Meadows discharge area are likely a mixture of Spring Mountain and Sheep Range recharge (modern $\delta^{18}O = -14.0\%$) with regional flow from the lower carbonate aquifer (modern $\delta^{18}O = -16.0\%$) in central Nevada [Winograd and Friedman, 1972]). Climate induced changes to the amount and isotopic composition of each mixing end-member would depend on local and regional climatic conditions. For example, almost no consideration has been given to possible impacts of summer versus winter precipitation sources, which have distinctly different $\delta^{18}O$ values [Ingraham et al., 1991]. However, the relative importance of each and its temporal variability are suggested in the fossil record [Spaulding and Graumlich, 1986]. In addition, changes in precipitation rates in central Nevada in the past 500,000 years could affect the hydraulic head and mixing proportion of different groundwater flow paths contributing to Devil’s Hole groundwater. Likewise, the extent of isotopic enrichment from evaporation during groundwater recharge may also change with climate. Therefore we suggest that more careful consideration is needed before interpreting the calcite $\delta^{18}O$ values strictly in terms of temperature variations without first considering the impact of groundwater mixing and changes in recharge sources.

5. Conclusions

In southern Nevada, interbasin flow is the dominant mechanism for regional groundwater transport. This mechanism is validated by the continuity of large-scale, low-$\delta^{18}O$ groundwater “plumes” that originate from high-latitude recharge areas and follow southward trends along graben valleys. Regional mixing with higher $\delta^{18}O$ groundwater at lower latitudes is clearly defined by the systematic increase in the $\delta^{18}O$ along these interbasin flow paths. These $\delta^{18}O$ values are independent of $^{14}C$ ages. Groundwater stable isotope values systematically plot to the right of the global meteoric water line and indicate postcoulod evaporation. This evaporation can occur in falling raindrops, melting and refreezing of snowpacks, during runoff, and by shallow groundwater evaporation. Evaporative signatures in groundwater are limited in the Las Vegas Valley groundwater basin.

Groundwater mixing biases the $^{14}C$ content of groundwater DIC towards the younger end-member. Ages that are calculated from mixed groundwaters are suspect unless mixing end-members are defined and their individual transport rates are known. Nevertheless, qualitative differences between the lower carbonate aquifer system of the White River basin and predominantly volcanic aquifers of the Oasis Valley system indicate that transport is faster through the White River basin. Our mass transport and mixing model for this basin is consistent with transport rates of $<10$ kyr from the Long Valley to the Pahranagat Valley. The inorganic $^{14}C$ has little bearing on groundwater age since water-rock interaction, coupled with mixing between local and regional flow, dominates the observed $^{14}C$ concentration. Paleoclimate evidence in groundwater is limited and may be found only in slower moving parts of localized flow systems.

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