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Evidence for interbasin flow through bedrock in the southeastern Sierra Nevada

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ABSTRACT

A combination of three independent lines of evidence (hydrogeological, geochemical, and isotopic) suggests that as much as 3.7×10^7 m^3 per year of interbasin flow may occur through faulted and fractured bedrock in the southern Sierra Nevada. The proposed interbasin flow path is through fractured Sierran bedrock from the Kern River drainage basin (which supplies recharge to the San Joaquin Valley to the west) into the southwestern Indian Wells Valley, a desert basin that is east of the Sierra. The interbasin flow produces an amount of recharge that significantly increases the ground-water budget for the desert valley.

The interbasin flow is identified by aquifer flux values much higher than local recharge and by ground water that has geochemical and isotopic signatures that differ from the local recharge. Interbasin flow that crosses topographic watershed divides has been identified in fractured carbonate aquifers in the southwestern United States. However, igneous and metamorphic rocks in the Basin and Range province are generally considered to be flow barriers, and watershed boundaries are usually delineated by topography. In this study we present an example in which recharge appears to be transported between adjacent watersheds through the fractured bedrock of an extensional area in the southern Sierra. If confirmed, this example shows that igneous and metamorphic rocks may not always act as barriers to ground-water flow in extensional tectonic regimes.

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INTRODUCTION

Ground water is the primary source of water in many arid and semiarid portions of the world. The demand on this critical resource will increase significantly in the next 30 yr as the largest population growth is expected to occur in these areas (Falkenmark and Lindh, 1993). For example, in the arid southwestern United States ground water currently supplies about half of total fresh-water use (Gleick, 1993). Most of this water is used for agriculture.

Alluvial deposits in arid regions can form prolific aquifers. The alluvial aquifers derive their recharge from streams and rivers that originate in the adjacent mountain ranges. There are few recharge measurements in these sparsely populated areas, so the usual procedure is to estimate recharge. The conventional methodology involves multiplying estimates and/or any measurements of precipitation at various elevation intervals in the mountains by the area of the basin's watershed within that elevation interval (Dettinger, 1989; Avon and Durbin, 1994). Recharge to the alluvial basin is considered to be the fraction of precipitation not lost to evapotranspiration that enters the basin as overland flow along streams draining the adjacent mountain ranges. The phreatic divide between ground-water basins is generally assumed to coincide with the topographic or surface drainage divides in the mountain watersheds (Mifflin, 1988).

However, some of the recharge areas are highly fractured. These fractures can impart a high degree of anisotropy to the direction of recharge. This flow may move in a different direction than the regional gradient if permeable fracture zones are not oriented parallel to the topographic gradient (Issar and Gilad, 1982). The recognition of fractures as a significant factor in fluid flow has prompted many studies on fractures on both the local and regional scale (Pollard and Aydin, 1988). Fractures are often the only pathways with significant permeability in hard rocks and limestones. Fracture patterns have been used to locate wells (Huntoon and Lundy, 1979; Gustafson, 1994) in order to develop effective remediation strategies (Banwart et al., 1994; Lachmar, 1994) and explain distribution of water quality on a regional scale (Weinberger and Rosenthal, 1994; Mayer, 1996).

Analysis of potential fracture-directed transport within the crystalline bedrock of the adjacent mountain ranges can be important in extensional geologic settings. The direction of subsurface flow in mountain aquifers may be controlled both by the orientation of the fractures and the head distribution within the fracture system, making the point of entry of this recharge into the ground-water system of the alluvial basin difficult to predict. Fracture-flow recharge may cross the topographic divides into an adjoining basin via subsurface flow and cause estimates based upon conventional means to be either too high or too low.

The Basin and Range province of the western United States, which consists of alternating basins and mountain ranges bounded by high-angle normal faults, is characterized by thin soils overlying crystalline bedrock in the mountain ranges and thick, alluvial deposits in the basins. The mountain bedrock is often fractured and faulted on multiple scales. The climate in the area is classified as midlatitude desert, and surface runoff from the mountains is the major source of recharge to the basins (Robertson, 1991). Precipitation in the mountains feeds mountain-front springs and streams in canyons that discharge into the basins, where surface water rapidly infiltrates the basin alluvium and recharges the alluvial aquifers (Heath, 1984). Accurate estimates of ground-water recharge in arid regions are essential in order to

Data Repository item 9989 contains additional material related to this article.

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develop the ground-water resources efficiently and to ensure that severe overdraft of the alluvial aquifers does not result. The purpose of this study is to verify and refine the existing recharge models for the Indian Wells Valley, California, using both hydrological and geochemical data.

DESCRIPTION OF THE STUDY AREA

The Indian Wells Valley is located in the southwestern part of the extensional Basin and Range province in eastern California, ~200 km north of Los Angeles (Fig. 1). The basin is bounded to the west by the Sierra Nevada Mountains, to the east by the Argus Range, to the north by the Coso Range and low-lying basaltic lava flows, and to the south by the El Paso Mountains. These mountain ranges are composed of tilted blocks of deformed sedimentary, igneous,

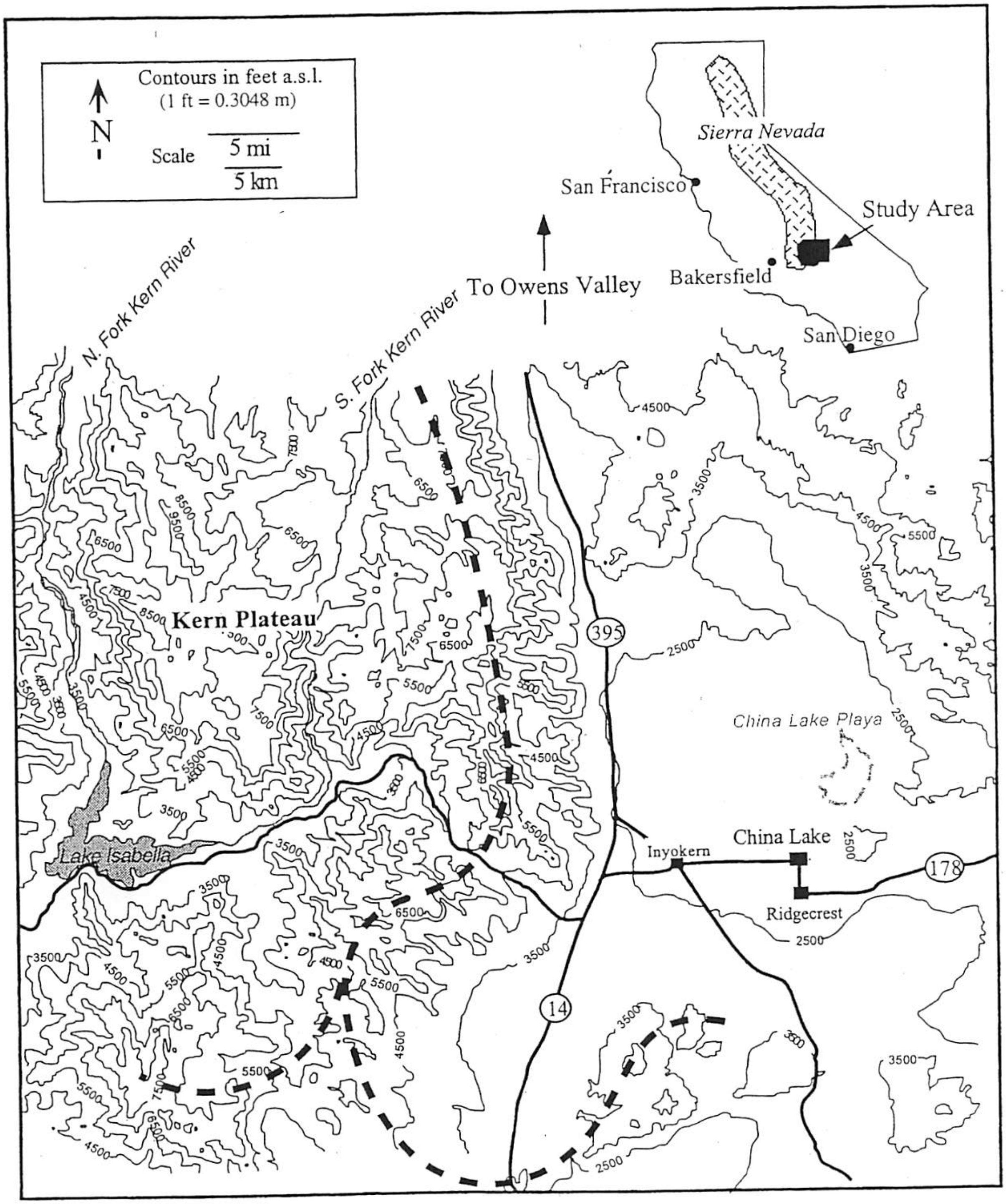

Figure 1. Topographic map of the Indian Wells Valley and Kern Plateau area encompassing lat 35°20′N to 36°10′N and long 117°20′W to 118°45′W. Note location of Lake Isabella in the west and the outline of the topographic drainage (heavy dashed line) for orientation. Modified from U.S. Geological Survey (1981).

and metamorphic rocks bounded on one or both sides by faults.

This region of the southern Sierra is very seismically active, with numerous low-magnitude earthquakes, ground deformation, and many recent faults in the valley (Roquemore, 1983; Zellmer, 1988). The Coso Range bordering the north end of the valley has fumaroles and intermittent thermal springs. Geothermal activity in the Coso Range has been sufficient to justify development of geothermal power facilities (Moore et al., 1982). Duffield et al. (1980) concluded the Coso Range is underlain by bodies of silicic magma that are associated with crustal extension. Saltus and Lachenbruch (1991) documented low reduced heat flows in the southeast Sierra west of the valley that are spatially correlated with a region of high seismicity including extensional earthquake swarms. They interpreted these data as evidence that extensional tectonics and associated magmatic processes are encroaching on the southeastern Sierra Nevada.

The mountain slopes adjacent to the valley are steep, particularly the eastern escarpment of the Sierras, which formed during Pliocene time by downfaulting the area to the east along the Sierra Nevada fault, a large, north-trending, normal fault (Christensen, 1966). The margins of the basin are characterized by coalescing alluvial fans derived from the adjacent mountains. The surfaces of these alluvial fans slope gently toward China Lake, a playa lake in the east-central part of the valley. During glacial periods, the valley was occupied by pluvial China lake, one of a system of pluvial lakes that received overflow from Owens and Mono Lakes to the north and, in turn, flowed into Searles Lake to the east (Grayson, 1993). The China Lake playa is all that remains of this large lake today.

The valley contains about 760 m of unconsolidated Pleistocene and Holocene alluvial-fluvial and lacustrine sediments that thin eastward to less than 305 m near the Argus Range (Zbur, 1963; Saint Armand, 1986). This unconsolidated basin fill overlies a basement of pre-Tertiary igneous and metamorphic rocks as well as consolidated sedimentary rocks of the Tertiary Ricardo Group (Loomis and Burbank, 1988). The alluvium, which consists of moderately to well-sorted gravel, sand, silt, and clay, forms the valley's aquifer (Kunkel and Chase, 1969; Dutcher and Moyle, 1973).

The current annual precipitation on the valley floor is ~10–15 cm and falls mainly during the winter and spring months (Berenbrock and Martin, 1991). Most of this rainfall is lost to evaporation that averages 200 cm per year from surface-water impoundments (Farnsworth et al., 1982). Most regional precipitation falls in the surrounding mountain ranges. The ranges surrounding the valley to the north, south, and east are lower in elevation and lie in the rain shadow of the Sierra Nevada. Therefore, the primary source of recharge for the valley is assumed to be precipitation on the east slope of the Sierras. The only significant perennial surface runoff is a small, influent stream draining the southern Rose Valley near Little Lake, a small perennial lake immediately north of the Indian Wells Valley.

Only the southern and western portions of the valley are populated and the remainder is used as part of a weapons test range. Ground water from the alluvial aquifer is the sole source of water for the city of Ridgecrest, the China Lake Naval Air Weapons Station, and other local industrial, agricultural, and domestic users (U.S. Bureau of Reclamation, 1993). Before 1953 (i.e., predevelopment), ground water in the alluvial aquifer moved eastward from the recharge area (Sierran watershed) across the valley to discharge by evaporation on the China Lake playa and by subsurface outflow into adjacent basins, such as Searles Lake basin to the east. Today, most of the ground-water discharge occurs by pumping from a series of public wells that are 100–300 m deep, located between Inyokern and Ridgecrest along Highway 178. There are also many shallow residential wells. Present total production is about 4.9×10^7 m^3 per year (U.S. Bureau of Reclamation, 1993). Historic water-level data show a slow decline in the water table near the city of Ridgecrest and the development of a water-table depression associated with the public well fields (Kern County Water Agency, 1992). However, other areas in the valley have not shown a significant decline in water levels over the past 40 yr (Bloyd and Robson, 1971, Fig 3; Kern County Water Agency, 1992).

Previous Work

Early works by Lee (1912) and Thompson (1929) laid the ground work for interpreting the aquifer architecture; they noted the varying water quality within the basin and estimated the amount of evapotranspiration from China Lake playa. These initial investigations were refined in later studies by Kunkel and Chase (1969) and Dutcher and Moyle (1973). In these later studies, the Indian Wells Valley alluvial aquifer on the western side of the basin was described as a single, deep unconfined aquifer with the majority of recharge supplied by the Sierran watershed. The aquifer becomes confined under lacustrine clays near the playa.

Recharge from Sierran mountain-front springs and streams flows down the canyons into Indian Wells Valley. The surface water rapidly infiltrates the basin alluvium and recharges the alluvial aquifers. Most streams are intermittent; there is significant flow only during the spring. Because stream gaging is of limited value in this setting, like most portions of the basin and range, recharge into Indian Wells Valley is estimated using alternative methods. The usual approach is to use estimates or measurements of precipitation for several elevation intervals multiplied by the area of the watershed (determined by topography) within an interval (Dettinger, 1989; Avon and Durbin, 1994). Recharge is that fraction of the precipitation not lost to evapotranspiration in the topographic watershed in the adjacent mountain ranges (i.e., the surface drainage area within the basin's topographic boundaries).

Most previous studies of the Indian Wells Valley aquifer have assumed that the basin is a closed hydrologic system, although some authors suggested that this assumption may not be valid (Austin and Moore, 1987; Erskine, 1989; Whelan et al., 1989). The closed-basin model implies that all of the basin's ground water is discharged by evaporation from China Lake playa and that there is no significant subsurface flow into or from the adjacent basins. Because there are no prior measurements of precipitation in the valleys' recharge area, recharge was estimated by assuming that the amount of recharge entering the valley was equal to the amount of evapotranspiration on the playa. Estimates of evapotranspiration in the playa have ranged from 3.9×10^7 m^3/yr (Lee, 1912) to 9.9×10^6 m^3/yr (Kunkel and Chase, 1969).

In 1991, 10 multiple piezometer wells were drilled to depths as great as 610 m in the western part of the valley (U.S. Bureau of Reclamation, 1993). Water-table elevations, hydraulic conductivity measurements, and samples for chemical analysis were taken in each piezometer. The multiple piezometer head data in the western Indian Wells Valley suggests that the deeper portion of the aquifer may be locally semiconfined, although no obvious confining layers are apparent in the well logs (Ostdick, 1997, Plates 1 and 2). The study results also indicate a fairly uniform value for hydraulic conductivity in the aquifer sands in the western valley, and downward vertical gradients along the basin edge indicate aquifer recharge.

Numerous shallow wells have been sampled in the valley for hydrochemical analysis (Kunkel and Chase, 1969; Warner, 1975; Whelan et al., 1989; Berenbrock and Schroeder, 1994; Houghton, 1994). Based on total dissolved solids (TDS), water quality ranges from very good to poor, degrading with depth. The best quality water is found in the southwestern corner of the valley, with TDS ranging from 150 to 300 mg/l. Ground-water quality in other parts of the valley ranged from 600 mg/l TDS in the northwestern part of the basin to <50 000 mg/l

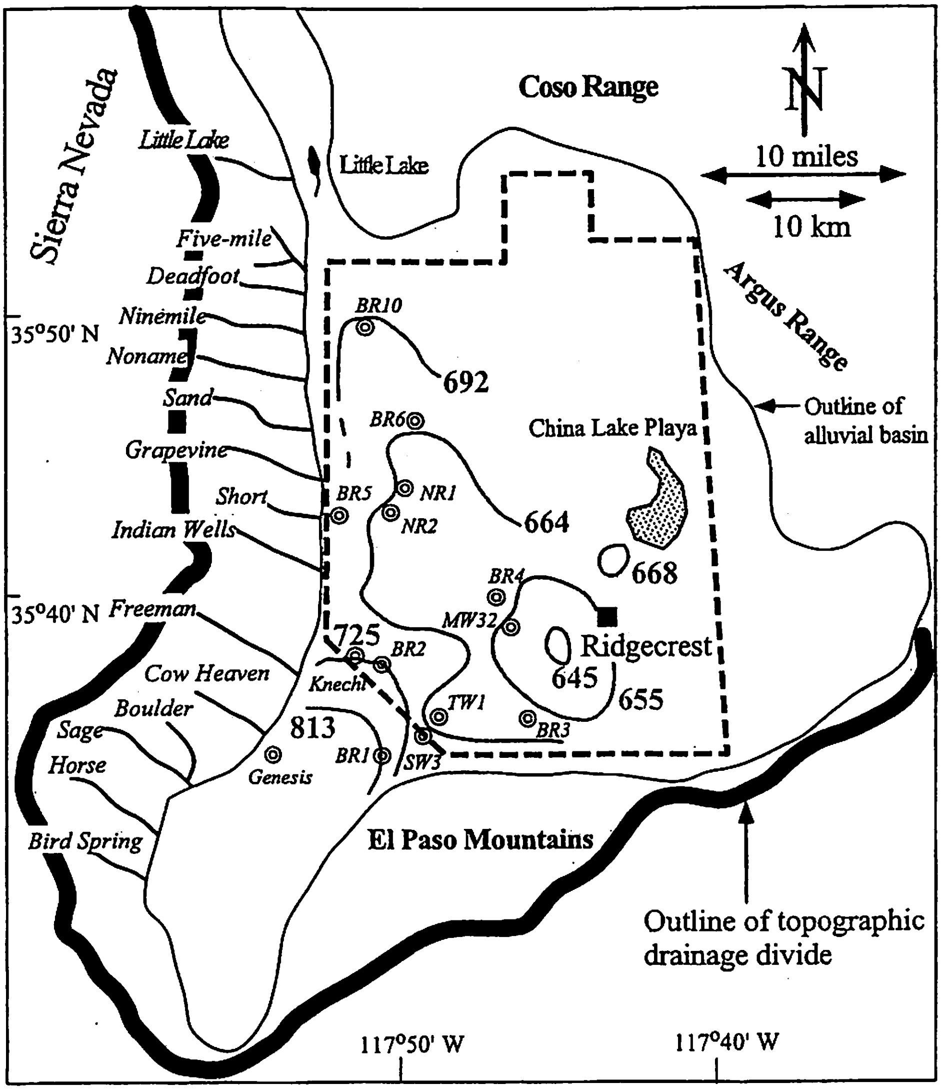

Figure 2. Schematic map view of Indian Wells Valley shows the locations of the canyons and wells (italics). Multipiezometer wells have the BR or NR prefix. The heavy dashed line shows the area included in prior models (Bloyd and Robson, 1971; Berenbrock and Martin, 1991). The potentiometric contours (in meters above sea level) are the heavy gray lines taken from 1991 head measurements (Kern County Water Agency, 1992). Note the irregular contour interval.

near the playa. Berenbrock and Schroeder (1994) suggested that water quality is related to aquifer sediment size, the worst quality water being associated with fine-grained lacustrine sediments. On the basis of stable isotopic data, they identified the low TDS ground water in the southwestern portion of the valley as relict Pleistocene recharge.

The demands of numerical models forced Bloyd and Robson (1971) and Berenbrock and Martin (1991) to quantify and distribute recharge to various parts of the basin's watershed. In their model, the recharge was considered to be equal to estimated evapotranspiration from the playa because they assumed a closed-basin system. Figure 2 shows the area modeled by prior workers together with the 1991 potentiometric surface for the valley. No-flow boundaries in the model area correspond to the aquifer limits, except in the southwestern part of the valley. In this area, the no-flow boundary was placed within the aquifer sediments and treated as a distant boundary by placing it at a sufficient distance from areas of interest so that it would have little effect on model simulations for critical areas (Berenbrock and Martin, 1991).

Bloyd and Robson (1971) distributed a total recharge for the valley of 1.2×10^7 m^3/yr by trial and error calibration of calculated water levels to measured values from equipotential maps of both predevelopment and postdevelopment conditions. The best-fit model assigned 64% of the recharge to the eastern Sierra Nevada, 32% to the Coso and Argus Ranges, and 4% to the El Paso Range. Berenbrock and Martin (1991) distributed the same total amount of recharge according to stream drainage areas above 1400 m in the Sierra Nevada and above 1500 m in the other ranges and did not adjust these values during calibration. In their model calibration, hydraulic conductivity, transmissivity, and leakance values were adjusted until the calculated head values reflected measured head values. In these modeling studies the distributed recharge models have a balanced water budget, but the recharge numbers are based on estimated discharge rather than direct measurements of either surface runoff or precipitation in the mountain ranges.

METHODS

We collected 167 samples for chemical and isotopic analyses over a 2 year period from Sierran springs and streams in the conventional watershed, wells in the valley, and springs and streams in the adjacent Kern River drainage basin to the west. Figure 3 shows the locations of water samples collected in this study. The wells include the 10 deep, nested piezometers installed by the U.S. Bureau of Reclamation in the spring of 1991. The results of the analyses are listed in the table in the Data Repository[1]. All the samples were measured on site for temperature, conductivity, pH, and alkalinity. The samples were filtered to 0.45 µm and cation samples were acidified to a pH of about 2 using one-half strength (8N) nitric acid. Chemical samples were stored in clean, rinsed HDLE (high density linear polyethylene) bottles and refrigerated until analysis at a laboratory (typically two to three days). Samples for stable isotope analyses were collected at the same time as the chemical samples. Samples were stored in 50 ml glass screw-cap vials sealed with Teflon tape until analysis. A few samples were collected for tritium analysis to determine the age of the ground water.

Like past investigators we assume the primary source of recharge to be derived from storms and snowmelt on the east slope of the Sierras that drain into the western valley. We measured the amount of precipitation in the topographic watershed of Indian Wells Valley and calculated how

[1]GSA Data Repository item 9989, water chemistry data, is available on the Web at http://www.geosociety.org/pubs/drpint.htm. Requests may also be sent to Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301; e-mail: editing@geosociety.org.

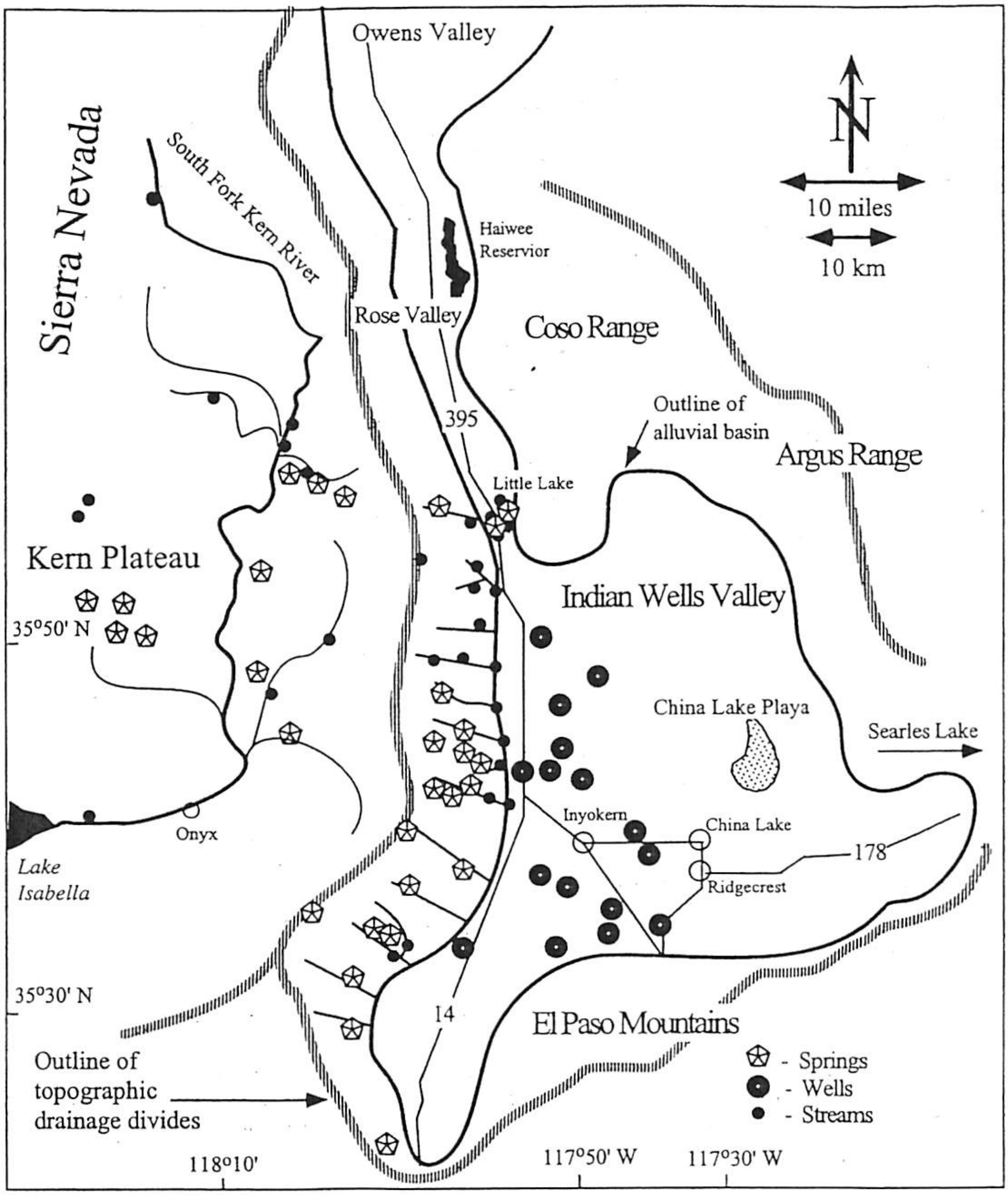

Figure 3. Schematic map view of study area showing the location of wells, springs, and streams sampled in this study. Chemical and isotopic data for the samples are in the Data Repository (see text footnote 1).

much of that precipitation actually recharges the aquifer. Accordingly, we divided the western part of the valley and adjacent topographic watershed into three sections: the southwest, west-central, and northwest (see Fig. 4).

In order to quantify Sierran recharge, precipitation gages were installed at 152 m elevation intervals in Freeman Canyon (Ostdick, 1997) and Ninemile Canyon. Precipitation samples for stable isotopic analyses were collected seasonally. The yearly total precipitation values were multiplied by the area of Sierran watershed within the elevation interval represented by each gage, as well as a factor to account for losses by evapotranspiration (Eakin et al., 1951). Surface outflow of the only perennial stream in the valley, which drains the Rose Valley to the north, was gauged on a regular basis. The average yearly flow was added to the total amount of recharge derived from the precipitation measurements to determine the total recharge for the western Indian Wells Valley.

Darcy's Law was used to calculate the amount of ground-water flux through the three sections of the aquifer. Water levels used to determine the gradient and values for hydraulic conductivity were taken from the U.S. Bureau of Reclamation (1993) data. Flux values were calculated assuming a single unconfined aquifer between pairs of U.S. Bureau of Reclamation project piezometers along a line perpendicular to the equipotential contours for each section. The gradient and hydraulic conductivity values in each nested piezometer were averaged for each piezometer pair used in the Darcy calculations to give a single value. The resulting average hydraulic conductivity and gradient values were used in the Darcy calculation. Hydraulic conductivity values for misrun slug tests were not included in the average. Cross-sectional areas were determined using an estimated 610 m depth in the west-central and northwest sections (depth component) and Dutcher and Moyles's (1973) map showing the limits of the Indian Wells Valley ground-water

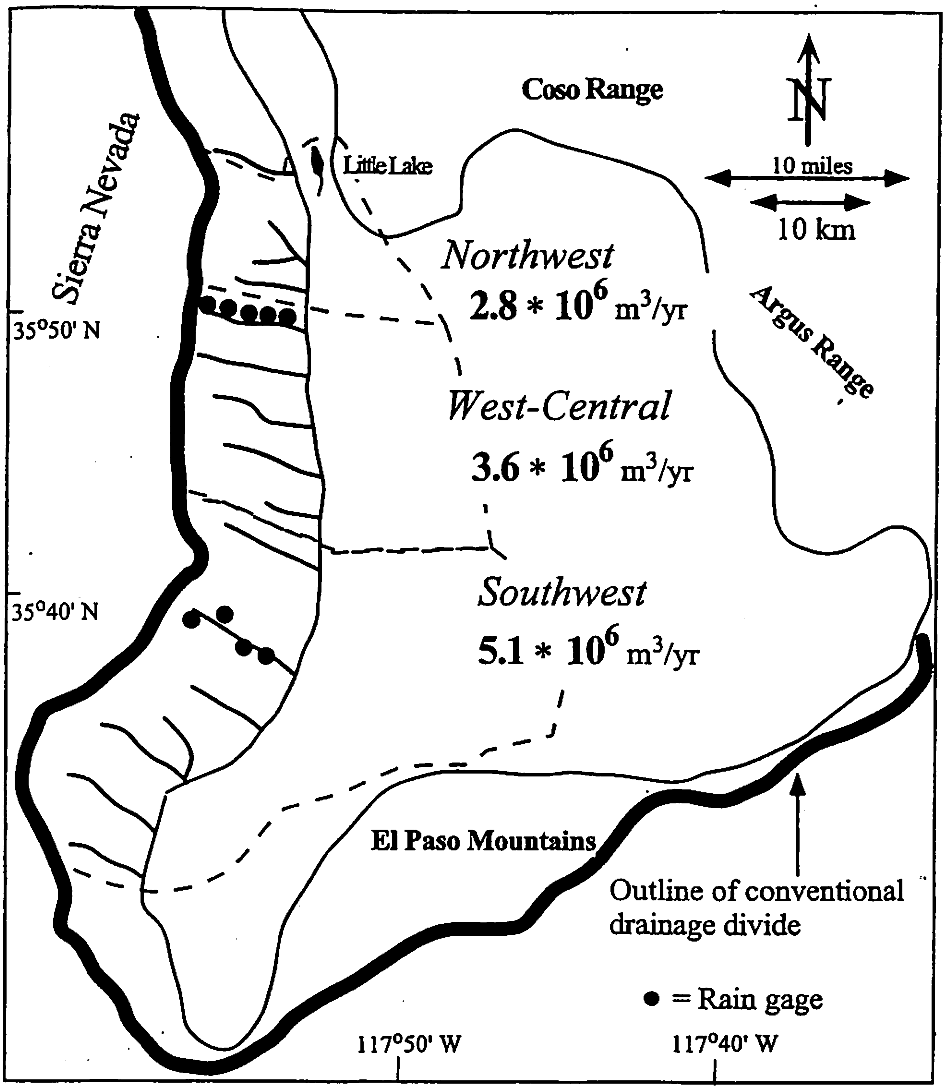

Figure 4. Results of recharge calculated by the precipitation-based method. The three sections of the study area and locations of the rain gages are also shown. The value for the northwest section includes the contribution from the Little Lake stream outflow (1.6×10^6 m^3/yr).

basin (width component). The cross-sectional area for the southwest was further refined by gravity surveys (Ostdick, 1997). The values used in the Darcy's calculations are listed in Table 1.

RESULTS

The results are divided into three parts; the recharge calculations, hydrochemistry, and isotopic data. The first section deals with our measurements of precipitation and calculations of recharge from the topographic watershed into the valley. The recharge calculations were independently verified by calculating the flux of ground water through the adjacent aquifer. Normally the amount of flux through the aquifer is similar to the amount of recharge into the aquifer. The second section deals with the chemical analyses of the water samples from surface and wells. The chemical data can be used to identify hydrochemical facies, groups of samples with similar chemical characteristics. Hydrochemical facies are used to delineate flow systems because samples from the same facies usually have similar flow paths and rock-water interaction. The last section considers the isotopic data. The hydrogen isotopic data (δD values) were used to supplement the hydrochemical data and to constrain the elevation range in the watershed from which each facies was derived. The tritium data were used to identify the location of recent (post-1953) water in the aquifer.

Recharge Calculations

Further details of methodology and the raw precipitation data are available elsewhere (Ostdick, 1997; Thyne and Gillespie, 1997). Figure 4 shows the locations of the rain gages, the boundaries of the three sections, and the calculated recharge values. The precipitation-based recharge values for each section of the western valley measured in this study are 2.8×10^6 m^3/yr for the northwest section, 3.6×10^6 m^3/yr for the west-central section, and 5.1×10^6 m^3/yr in the southwest section (Fig. 4). The total amount of recharge from the eastern Sierra Nevada watershed in the 1994–1995 season (a wet year by California standards) was 1.1×10^7 m^3/yr. Together with the Rose Valley stream flow, which averaged 1.6×10^6 m^3/yr (Gillespie et al., 1996), the total recharge to the valley is calculated to be 1.3×10^7 m^3/yr. This value is very close to the 1.2×10^7 m^3/yr estimated by Bloyd and Robson (1971) for the entire valley

The recharge values derived from the precipitation measurements were then compared to the calculated ground-water flux through each section of the aquifer (Fig. 5) as an independent check. The calculated aquifer flux for the northwest section of the aquifer is 5.9×10^6 m^3/yr based on the gradient between wells BR-10 to BR-6. The flux for the west-central section (wells BR-5 to NR-2) is 4.0×10^6 m^3/yr, and the flux for the southwestern section is 4.2×10^7 m^3/yr (wells BR-1 to MW-32).

There is good agreement between the calculated precipitation-based recharge and the ground-water flux calculations in the northwest and west-central sections of the aquifer (Figs. 4

TABLE 1. DATA USED FOR DARCY FLUX CALCULATIONS

Area	Gradient transect (see Fig. 3B)	K* (m/d)	Gradient†	XS area (m²)
Northwest	(BR10–BR6)	4.3	0.002	1.9×10^6
West-central	(BR5–NR2)	4.9	0.002	9.3×10^6
Southwest	(BR1–MW32)	4.0	0.013	2.2×10^6

*From slug test data (U.S. Bureau of Reclamation, 1993, Appendix 6, p. 16).
†From water level data (U.S. Bureau of Reclamation, 1993, Table 2, p. A49–A51).

and 5). The small differences may be due to the Maxey-Eakin recharge coefficients overestimating evapotranspiration (McDonald et al., 1997) or recharge from the Coso Range (not measured in this study). The calculated flux through the aquifer in the southwest section of the valley is 4.2×10^7 m^3/yr. This value is nearly an order of magnitude greater than the precipitation-based recharge available from the adjacent Sierran watershed.

The large aquifer flux is produced by the steep head gradient in the southwestern portion of the basin coupled with the hydraulic conductivity values (Fig. 6). This steep gradient is a long-term feature that is unchanged over at least the past 45 yr (Dutcher and Moyle, 1973; Kern County Water Agency, 1992). Kunkel and Chase (1969) inferred that a fault, which acts as a barrier to ground-water flow, is responsible for the steep gradient in this part of the basin. We investigated this possibility using field and aerial photo studies and gravity survey data (Ostdick, 1997), and were unable to locate a fault in the area. However, if future drilling in the area provides evidence for the presence of a flow barrier, the hydraulic conductivity values should be reevaluated and the flux values recalculated.

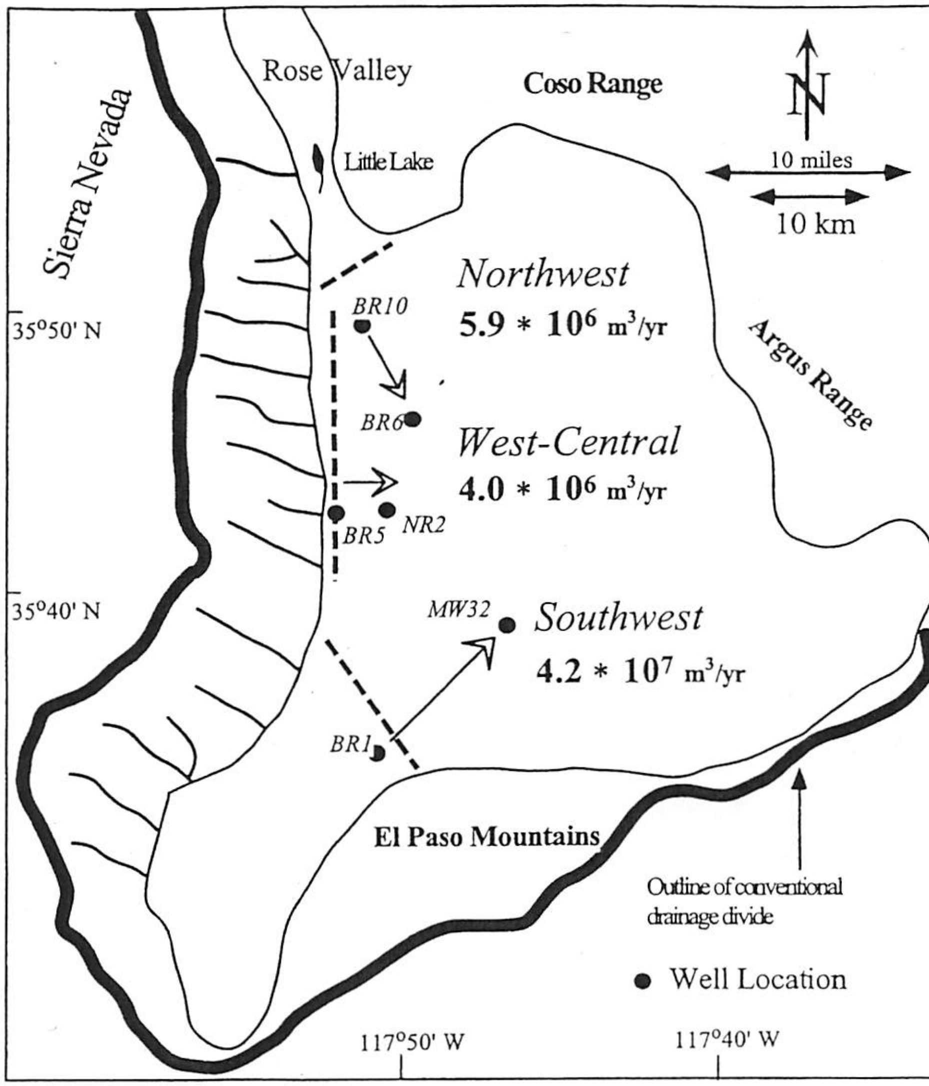

Figure 5. Results of the aquifer flux calculations using Darcy's Law. Data used in the calculations are listed in Table 1. The locations of the wells used to estimate the potentiometric gradient are shown with arrows to indicate the direction of flow. The heavy dashed lines indicate the cross-sectional area used to calculate the flux.

Hydrochemistry

The majority of the 169 samples collected from the valley and Sierras can be divided into 4 hydrochemical facies based on the major ion chemistry (Fig. 7). The facies are descriptively named: (1) the Sierran group, (2) the Indian Wells Valley group, (3) the Little Lake group, and (4) the Southwest Wells group. Table 2 shows the mean, average, maximum, minimum, and standard deviation of concentration for the major ions in each of the hydrochemical facies. The 12 samples that did not fit into the 4 facies are either from the deep piezometers of NR1, NR2, and BR3 that have chemical compositions influenced by geothermal brines, or the shallow BR2 piezometer that was excluded due to very alkaline pH values. The results of chemical and isotopic analyses are listed in the table in the Data Repository (see footnote 1). Samples with charge balances greater than ±10% were excluded from the statistical analysis, but are included in the same table.

The Sierran Group is composed of 40 samples from springs and streams in the topographic watershed of southwestern Indian Wells Valley (Bird to Indian Wells canyons) and from springs and streams that drain into the south fork of the Kern River (Fig. 8). The TDS value is low (<400 mg/l) with Ca > Na, low concentrations of K and Mg, and $SO_4 \approx Cl$. The Indian Wells Valley Group (57 samples) has a chemistry similar to the Sierran Group, but with a higher TDS (average = 793 mg/l), Na ≥ Ca, and $SO_4 > Cl$. The samples include streams and springs in the west-central and northwest sections (Short to Little Lake canyons) and some of the wells in those sections. The chemistry in the Sierran and Indian Wells Valley groups is typical for meteoric water weathering Sierran rocks (Garrels and Mackenzie, 1967) with the HCO_3 content positively correlated to the extent of water-rock reactions. The dissolved constituents are primarily derived from the dissolution of granodioritic minerals with increasing water-rock interaction generating higher TDS over the longer and/or slower flow paths as elevation decreases (Fig. 9).

The Little Lake Group includes 33 samples from Little Lake in the southern Rose Valley, some of the springs and wells around Little Lake, and wells BR5, BR6, and BR10 in Indian Wells Valley (see Fig. 2). The water chemistry is significantly different than the first two groups with higher TDS (average = 1177 mg/l), much higher Na contents, Mg > Ca, and $Cl \geq SO_4$. The chemistry is distinctive because few natural waters have more magnesium than calcium (Hem, 1985). The samples from this group have elevated lithium and boron contents that indicate a geothermal component. The abundance of dissolved magnesium is probably related to contact with igneous or alteration minerals (olivine and pyroxene, chlorite and serpentine) found in the basalt flows and mafic intrusions common to

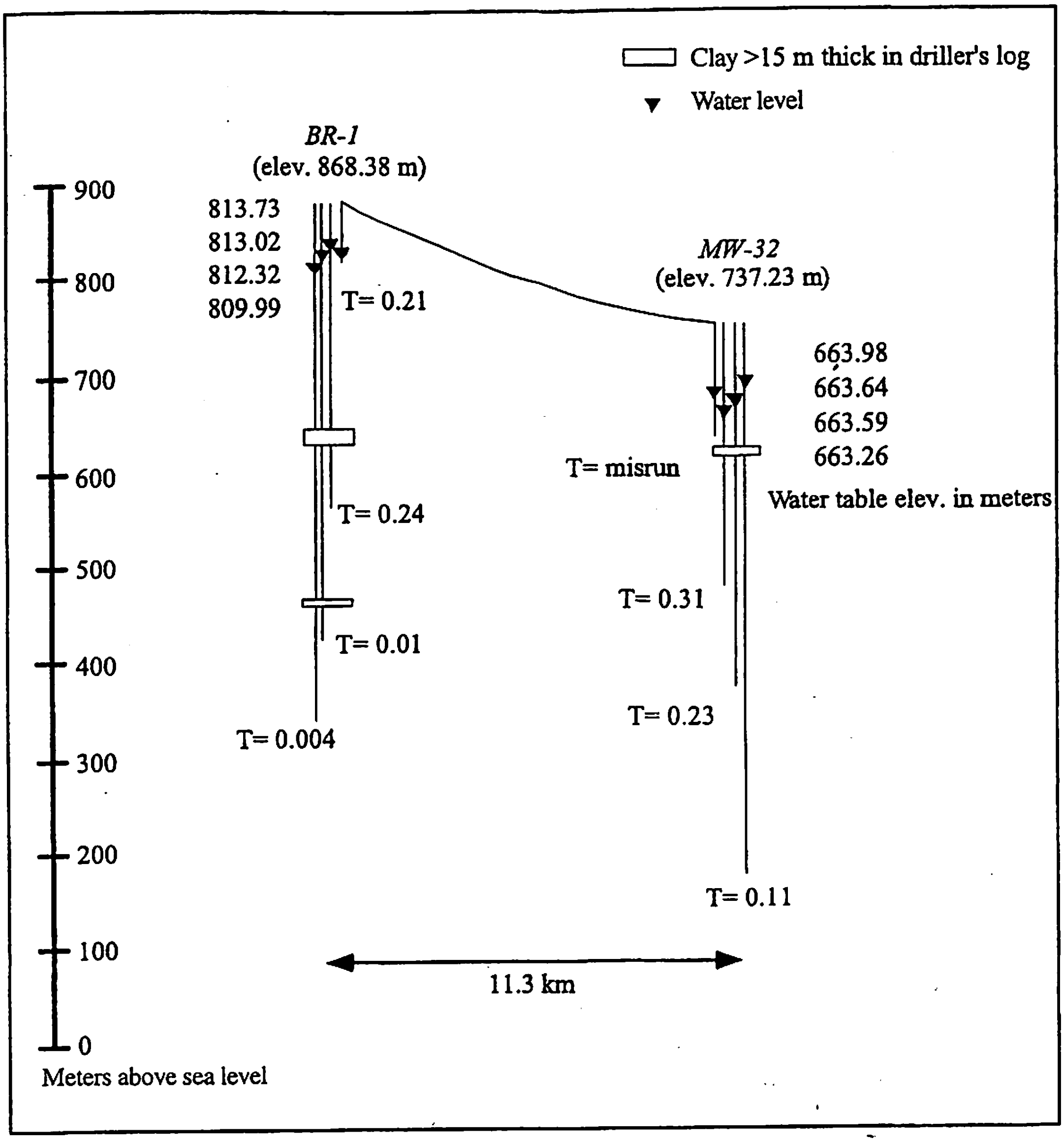

Figure 6. Schematic cross section of the multipiezometer wells and slug test hydraulic transmissivity values for the southwest section wells BR1 and MW32 used for the flux calculations.

northeastern Indian Wells and southern Rose Valleys and the Coso Range.

The 27 samples of the Southwest Wells group are from wells BR1, BR2, SW-3, Genesis, Knecht, and wells BR4 and MW32 (near the public well field west of Ridgecrest). The dissolved content of these samples is low (average TDS = 267 mg/l) with SO_4 + Cl < HCO_3 and cations dominated by Na (>90% of total cations). The water chemistry does not appear to be directly related to that of the local recharge.

The data show that the surface runoff from the topographic watershed in the eastern Sierra and outflow from Little Lake to the north supplies the ground water in the northwest and west-central sections of the valley, supporting the hydrogeologic data. The distinctive signature of the Little Lake group can be traced southward from Little Lake and the perennial stream toward the west-central portion of the study area. Upon recharge the ground water shows little change in TDS with increasing depth (Fig. 8) in the aquifer (except for the deep samples discussed previously), and the ground water remains in the same hydrochemical facies as the adjacent watershed, except for the Southwest Wells group.

The water in the southwestern wells does not belong to the same hydrochemical facies as recharge in the adjacent surface watershed, as is the case in the other sections. The Southwest Wells group samples have a lower TDS (average value = 262 mg/l) than any samples collected in Indian Wells Valley except for some low-volume springs in the southwest section (see Fig. 8). The local recharge is too saline (average TDS = 434 mg/l) to be the primary source of the Southwest ground water. Prior investigators have interpreted the low TDS ground water as being Pleistocene age recharge that has remained relatively unaltered by mixing with the modern recharge that has a higher TDS (Berenbrock and Schroeder, 1994).

Isotopic Data

The stable isotopic data can also be used to analyze the Indian Wells Valley flow system. Hydrogen isotope ratios (δD) are generally conservative (unaffected except by mixing) and are often used in ground-water studies (Mazor, 1991; Drever, 1997). Stream samples draining the topographic watershed in the northwest and west-central sections have δD values between −83‰ and −97‰, and samples of ground water from the same chemical group (Indian Wells Valley) have values between –92‰ and –97‰. Surface samples from the Little Lake Group have δD values between –95‰ and –108‰, similar to the ground-water samples from the same chemical group (–92‰ and –105‰). Most of the ground-water samples have isotope values that appear weighted toward the negative compared to the watershed samples. This difference reflects infiltration studies that suggest that recharge contribution to ground water increases with greater elevation and that in arid climates little of the precipitation that falls below 1370 m contributes to ground water (Eakin et al., 1951). However, ground-water samples from the Southwest Wells group have δD values between –96‰ and –115‰, significantly lighter than the recharge values in the adjacent watershed, which range from –88‰ to –100‰ (average δD = –92‰).

In the study area the major control on the range of stable isotope values is elevation. Precipitation becomes increasing depleted in the heavier isotopes (^{18}O and deuterium) as elevation increases. This produces a systematic trend that can be used as an indicator of recharge elevations for hydrochemical studies; however, the altitude effect is unique to each locale (Mazor, 1991). Based on the available data the relationship between δD and elevation in the study area is described by the equation $y = -25.8x - 204$, where x is the δD and y is elevation in meters (Fig. 10).

Using this relationship the ground water in the Little Lake Group (–92‰ to –105‰) is derived from precipitation that fell at elevations of 2170–2505 m, matching the elevation range found in the Sierran watershed of the Rose Valley. Samples from the Indian Wells Valley Group have values (–83‰ to –99‰) indicating an elevation range of 1937–2350 m found in the topographic drainage area (Five-mile to Bird Canyon). The Southwest Wells Group samples have an isotopic signature of –96‰ to –115‰ that corresponds to an elevation range of 2273–2763 m. However, only a few percent of the conventional watershed for the Indian Wells Valley is within the lower portion of this elevation range, much too small an area to supply the majority of the ground water in the southwestern valley.

In order to test the relict Pleistocene recharge hypothesis, samples were obtained for tritium analyses from streams in the southwestern topographic watershed, the South Fork of the Kern River, wells BR1 and SW3, and production wells of the Indian Wells Valley Water District. The results of the tritium analyses are reported in Tritium units (TU) and are shown in Table 3. Tritium levels in samples of present-day runoff (Kern River) show low values (~5 TU) similar to other modern samples in the southwestern United States (Vuataz and Goff, 1986). Wells adjacent to

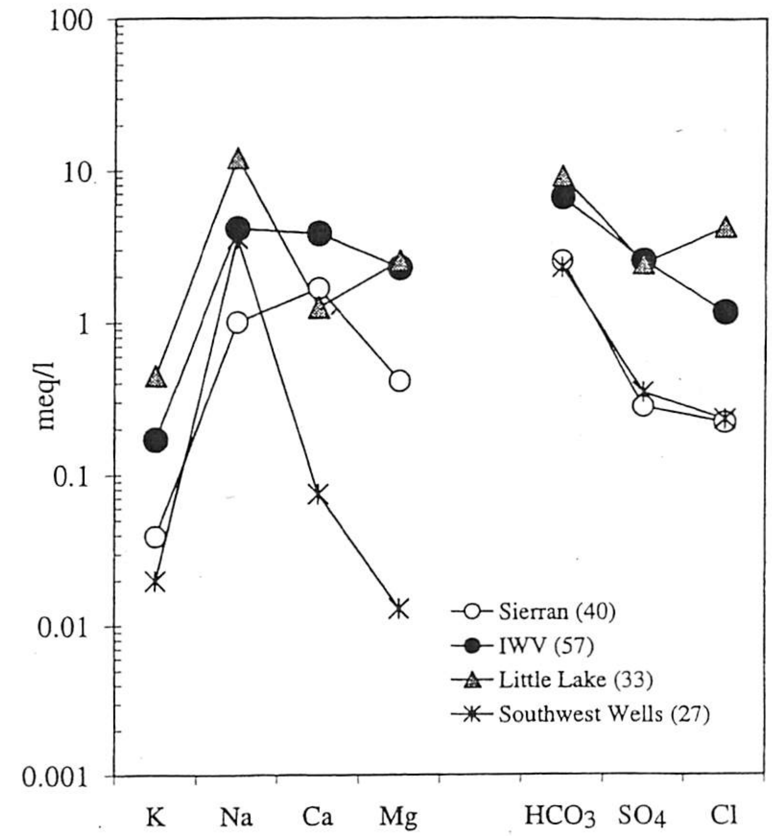

Figure 7. Schoeller diagram of the four hydrochemical facies (Sierran, Indian Wells Valley, Little Lake, and Southwest Wells) found in the study area. Mean values are plotted, other statistical data for each facies are listed in Table 2.

TABLE 2. STATISTICAL VALUES FOR HYDROCHEMICAL FACIES

Facies		K	Na	Ca	Mg	HCO3	SO4	Cl	TDS	δD	pH	Mg/Ca
Sierran (40)	Mean*	0.04	1.0	1.7	0.4	2.5	0.3	0.2	201	–95.0	7.5	0.25
	Average	0.05	1.2	2.0	0.6	2.9	0.6	0.3	234	–96.1	7.5	0.30
	Maximum	0.13	2.2	4.0	1.7	6.1	1.6	1.1	417	–85.0	8.1	0.51
	Minimum	0.00	0.1	0.3	0.0	0.5	0.0	0.0	46	–112.0	6.6	0.02
	Stdevp	0.03	0.6	0.9	0.4	1.4	0.5	0.3	105	7.0	0.4	0.14
IWV (57)	Mean*	0.17	4.1	3.8	2.3	6.6	2.5	1.2	691	–94.5	7.7	0.59
	Average	0.22	5.0	4.2	2.8	7.4	3.0	1.6	734	–94.8	7.8	0.66
	Maximum	0.87	20.0	11.0	8.2	34.7	9.6	6.2	2300	–83.0	8.8	1.86
	Minimum	0.02	1.3	0.1	0.0	2.4	0.2	0.3	421	–111.0	6.9	0.17
	Stdevp	0.18	3.5	1.7	1.7	4.6	1.8	1.6	301	6.5	0.4	0.33
Little Lake (33)	Mean*	0.45	12.0	1.2	2.5	9.1	2.4	4.2	1072	–95.0	8.5	2.00
	Average	0.53	12.8	2.0	3.7	10.7	3.2	4.7	1124	–88.5	8.5	2.45
	Maximum	1.46	25.7	9.1	9.6	31.2	10.0	10.4	2010	–23.0	11.0	9.32
	Minimum	0.06	4.9	0.1	0.0	1.9	0.0	0.8	484	–108.0	7.2	0.47
	Stdevp	0.27	4.5	1.9	2.2	5.1	2.0	2.0	331	22.5	0.8	1.81
Southwest Wells (27)	Mean*	0.02	3.6	0.07	0.01	2.3	0.3	0.23	241	–103.0	8.8	0.24
	Average	0.02	3.7	0.21	0.08	2.5	0.5	0.40	247	–102.3	8.8	0.43
	Maximum	0.06	5.1	1.75	0.56	4.0	2.1	1.61	370	–95.0	9.7	2.74
	Minimum	0.01	2.2	0.02	0.00	1.0	0.2	0.01	155	–111.0	6.8	0.07
	Stdevp	0.01	1.0	0.43	0.18	0.7	0.6	0.49	56	3.8	0.7	0.64

Notes: Major ions in meq/l; TDS in mg/l; Mg/Ca in meq/l; pH in standard units; δD in ‰ SMOW; number of samples in each group in parentheses.

*Geometric mean; median for isotopic data.

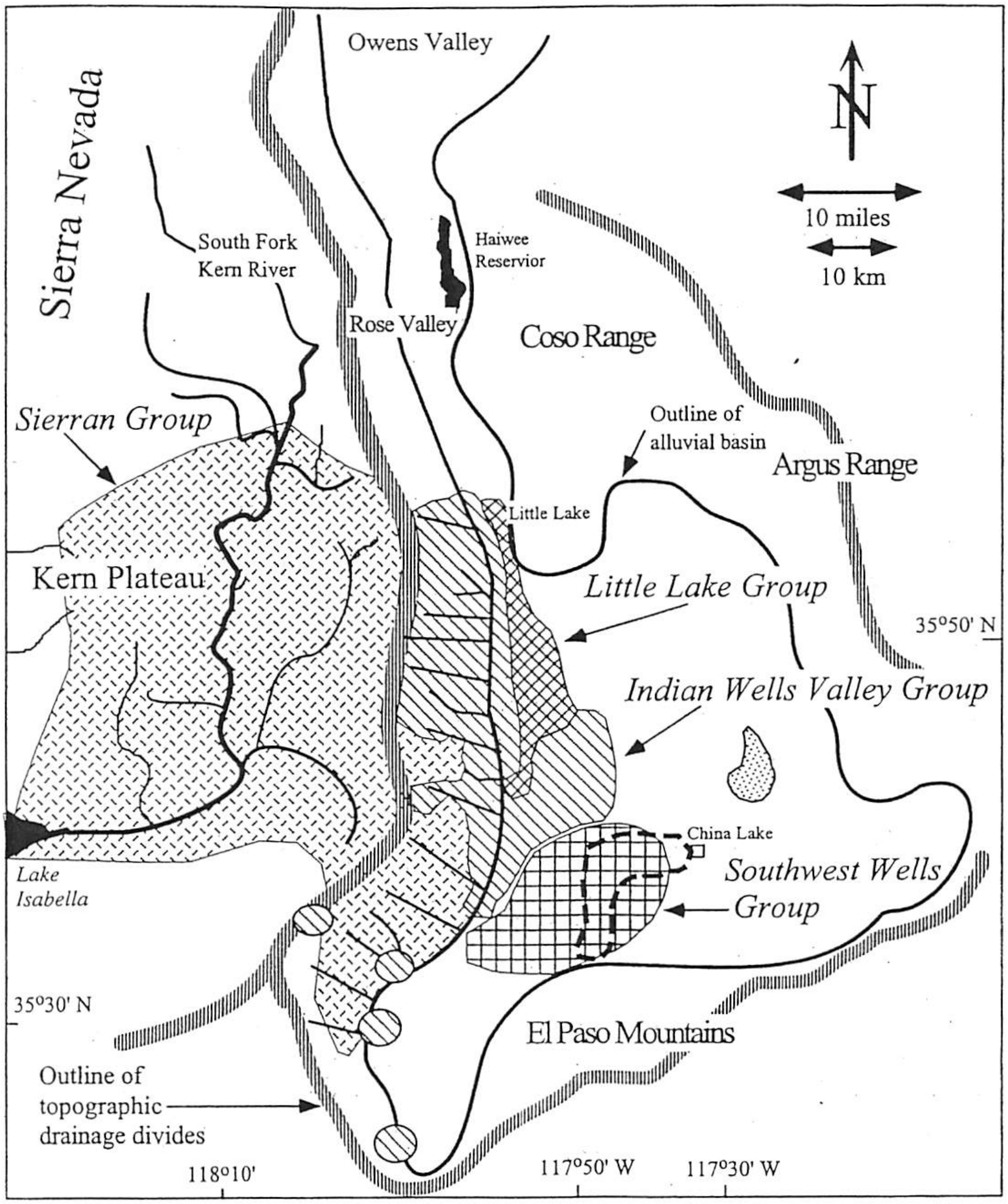

Figure 8. Map view of the hydrochemical facies (italics) in the study area. The heavy dashed line shows the location of wells where the tritium data show that a significant fraction of the ground water is post-1952 in age.

the basin margin show values <5 TU, indicating a pre-1952 age, with deeper wells having lower values. However, the tritium data from the Indian Wells Valley Water District production wells and well SW-3 (Fig. 8) have values >10 TU, indicating a post-1952 age (Mazor, 1991). The production intervals sampled (100–300 m) are not in direct contact with surface waters (see hydrographs for well MW-32; Ostdick, 1997), which eliminates surface contamination as a source for the tritium. Samples with post-1952 tritium values are interpreted as a mixture of older ground water with very low tritium content and more recent recharge. However, without further geochemical constraints we cannot determine the relative fractions (Mazor, 1991). In spite of this limitation, postbomb tritium in the ground water in the southwest section of the aquifer casts doubt on interpretations that the low TDS southwestern ground water is primarily relict Pleistocene recharge.

DISCUSSION

The integrated hydrologic, isotopic, and hydrochemical data show that ground water in the northwest and west-central portions of the valley is derived from the topographic watershed of the adjacent eastern Sierra, a conclusion similar to prior studies (Houghton, 1994; Berenbrock and Schroeder, 1994; Whelan et al., 1989). However, our interpretation for the source of ground water in the southwestern section of the valley differs from previous studies.

The precipitation-based recharge calculations in the southwest section show that the topographic watershed in the southwest can supply only 12% (5.1×10^6 m^3/yr) of the calculated flux through that section of the aquifer (4.2×10^7 m^3/yr). The flux calculations are somewhat simplistic, given evidence that the western portion of the valley may not be a single deep unconfined aquifer (U.S. Bureau of Reclamation, 1993). Darcy calculations are subject to uncertainty (mostly due to hydraulic conductivity values that may show variation of an order of magnitude), but the other sections of the valley showed good agreement between the calculated recharge and flux values. The hydrogeological data are not sufficient to constrain a more complex aquifer model, nor is it apparent that a more complex model would resolve the discrepancy.

The chemical and isotope ratios of the ground water in the northwest and west-central sections are clearly related to the recharge from the adjacent topographic watershed, but this is not the case in the southwest section. The calculated high flux rates for the southwest section generate faster travel times between wells BR1 and MW32 (~150 yr; Ostdick, 1997) than the prior models that generated travel times of thousands of years (Berenbrock and Martin, 1991). These higher ground-water velocities appear more reasonable given the tritium data.

Given the hydrogeologic and hydrochemical data, we believe that runoff from the adjacent topographic Sierran watershed cannot be the sole source of the southwestern ground-water samples. We hypothesize that the southwestern ground water is a mixture of two components, runoff from the topographic Sierran (local) watershed and water from outside the Indian Wells Valley drainage basin (interbasin flow). We can formulate three constraints to help determine the location of the extrabasinal recharge source: (1) water from this source must have chemical and isotopic contents that when mixed with the local recharge will produce the southwestern ground water; (2) the source area must receive sufficient precipitation to supply the additional recharge (as much as 3.7×10^7 m^3/yr); and (3) a viable flow path must connect the source to the southwest Indian Wells Valley.

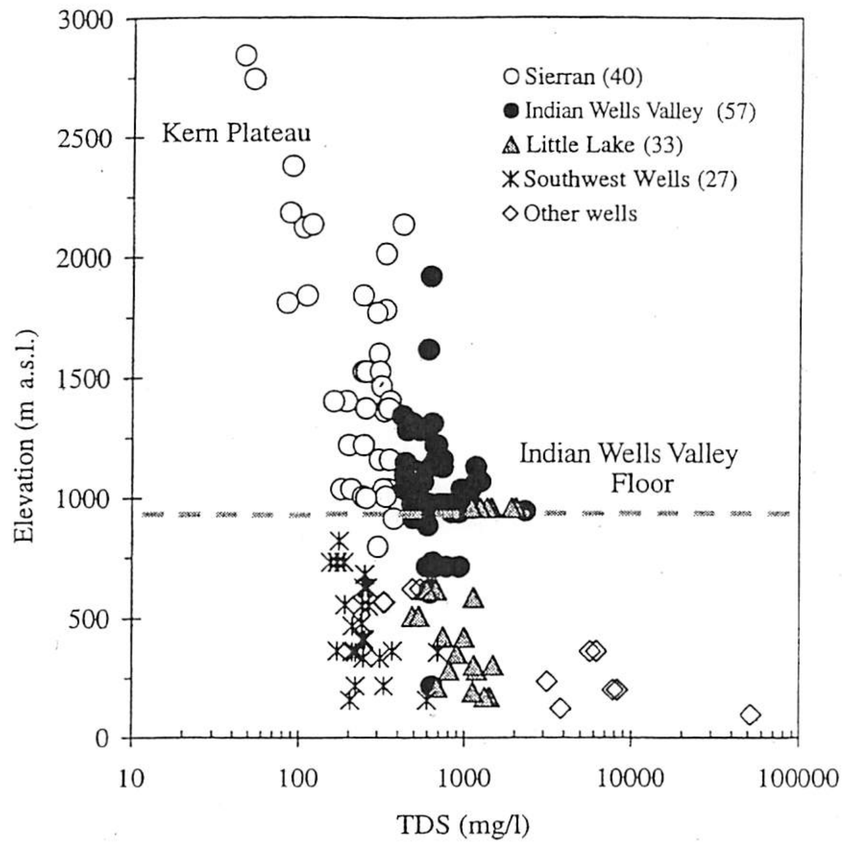

Figure 9. Plot of elevation (meters above sea level, m a.s.l.) versus total dissolved solids (TDS) for the hydrochemical facies and the deep wells (other wells) influenced by geothermal brine.

Geochemistry

The data from the local recharge in the southwest Indian Wells Valley and the southwestern ground water suggest that the source of additional recharge is located at greater than 2273 m elevation and has lower TDS than the local recharge. Samples with the appropriate isotopic signature and TDS are found in the Kern Plateau area, a source area for the south fork of the Kern River. The Kern Plateau samples have an average δD value of −103‰ (2453 m), and an average TDS of 135 mg/l.

To evaluate the Kern Plateau water as the source of extrabasinal recharge we formulated a mixing model using the chemical and isotopic data. The results can be compared to the hydrogeologic evidence that 88% of the recharge in the southwest section is derived from outside the basin. Mixing fractions calculated this way are likely to be as accurate as the Darcy calculations given the degree of uncertainty in both methods (Dettinger, 1989). The accuracy of mixing models is dependent on end members with well-defined compositions of conservative parameters (unchanged except by mixing) in which the values are significantly different. The end members in our system (Kern Plateau, Southwest Wells, and local recharge) have a range of values that are a result of factors such as seasonal variations in chemistry and mixing in the aquifer from dispersion during ground-water flow. This means the calculations can generate many valid solutions. To minimize this limitation we use average values in the calculation.

The Southwest Wells samples have a cation chemistry dominated (>90%) by sodium; however, there is no surface water in the area with the same chemistry. This water chemistry is similar to ground water from other alluvial aquifers in the region (Dale et al., 1966; Swartz et al., 1996) that has been affected by rock-water interactions such as cation exchange (Robertson, 1991). Given the possibility that the cation chemistry in the Southwest Wells group has been affected by cation exchange, we must eliminate the cations from the model.

This leaves chloride, sulfate, and δD as likely conservative parameters. We also include TDS, which is not a single parameter, but may behave somewhat conservatively in the system. Based on mixing calculations, the Kern Plateau water mixing with local recharge can form the Southwest Wells water. The values for the Kern Plateau fraction are 97% using δD, 90% using sulfate, 76% using TDS, and 60% using Cl. The four parameters produce results that bracket the flux calculations (88%) and show that the Kern Plateau is a possible source of extrabasinal recharge.

Kern Plateau Precipitation

The precipitation on the Kern Plateau drains south and eastward into the south fork of the Kern River, which discharges westward into Lake Isabella. The annual inflow (surface drainage from the Kern River) into Lake Isabella

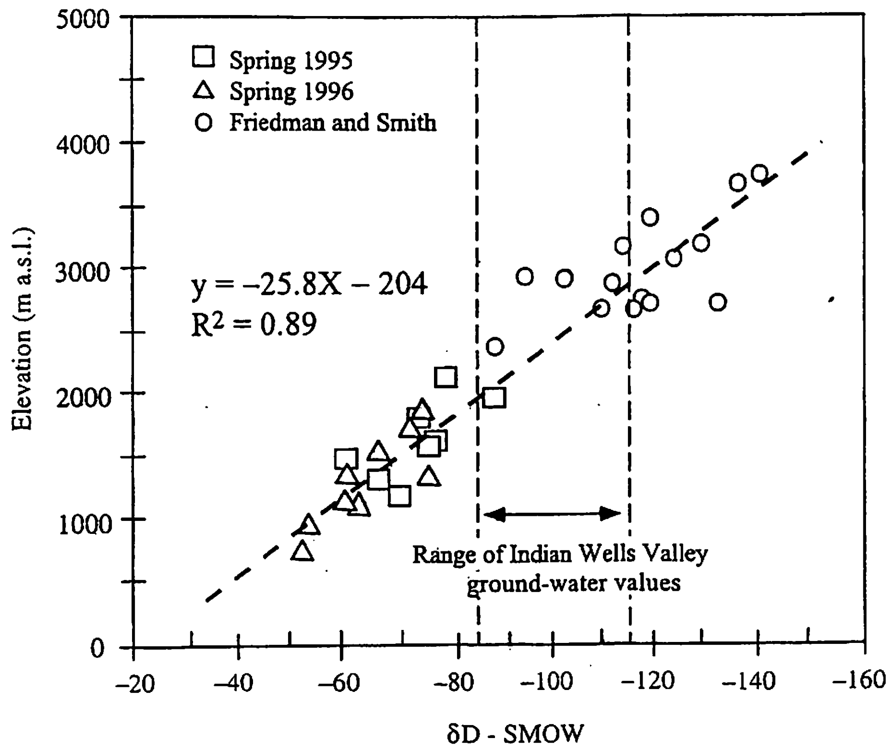

Figure 10. Plot of hydrogen isotope rations (δD) (SMOW, standard mean ocean water) versus elevation for the study area. Dashed line shows the linear regression for the data (equation shown). The data are from the rain gauge samples and published values (Friedman and Smith, 1972).

TABLE 3. TRITIUM DATA FROM SURFACE AND WELLS

Sample location	TU	eTU
Surface samples		
Kern River (South Fork)	5.37	0.15
Kennedy Meadows (Government Spring)	2.33	0.09
Big Spring	0.03	0.09
Indian Wells Canyon Mine spring	1.52	0.09
Walker Pass Spring	1.33	0.09
McIvers Spring	5.52	0.18
Bird Spring	0.38	0.09
Well samples		
BR-1MS	5.00	3.00
BR1 MD	2.00	2.00
BR2 M	1.00	2.00
BR2 D	0.00	2.00
Knecht well	0.11	0.09
Well SW-3*	73	37
Well 31*	98	25
Well 30*	35	24
Well 19 (Navy)*	79	25
Well 17 (CLA17)*	29	24
Well 16*	21	24
Well 13*	92	25
Well 12*	58	25
Well 11*	67	25
Well 10*	82	25
Well 9*	59	25
Well 8*	87	25
Well 7*	28	25

*Indian Wells Valley Water District samples courtesy of R. Tucker, analyzed by Davi Laboratory, Environmental Associates. Other samples analyzed by University of Miami. eTU is 1σ standard deviation.

averages 8.8×10^8 m^3/yr over about the past 45 yr (U.S. Geological Survey, 1994). The 3.7×10^7 m^3/yr hypothesized influx to the Indian Wells Valley would be <5% of this amount. The head gradient between the Kern Plateau and Indian Wells Valley is steep (0.05). In addition, a map of the water table elevation within the southeastern Sierra derived from contouring the elevations of springs and perennial streams (Fig. 11) indicates that the main phreatic divide is beneath the Kern Plateau with elevations as high as 2800 m. The surface drainage divide between the Kern River and Indian Wells Valley is east of the main phreatic divide and is underlain by a secondary phreatic divide with maximum elevations of only 2180 m. Because the main phreatic divide under the Kern Plateau is to the west of the surface drainage divide, deep ground water in fractures could cross under the south fork of the Kern River, bypassing the secondary phreatic divide driven by head elevations as high as 2800 m. Therefore, deeper, regional groundwater flow paths originating on the Kern Plateau may cross beneath the secondary phreatic divide in the manner described by Toth (1963). Most of the water from the Kern Plateau likely discharges into Lake Isabella because the gradient between the plateau and the lake is steeper than that between the plateau and Indian Wells Valley. However, the ultimate head drop between the plateau and Indian Wells Valley is 150 m greater than the head drop to Lake Isabella.

Flow Paths

Enhanced permeability and flow along faults is common in extensional areas where the stresses tend to pull brittle rocks apart and widen the fracture apertures (Huntoon, 1986; Barton et al., 1995). The ability of fractures in the crystalline rocks of this area to conduct fluid flow is supported by data from a recent deep test well drilled into crystalline basement in western Indian Wells Valley (Monastero et al., 1995). This well lost circulation during drilling within an interval of highly fractured crystalline rock at 1850 m depth that may represent a major fault zone in the crystalline basement. North of our study area a deep flow path in the fractured crystalline rocks between the Sierra Nevada and the Coso geothermal field has been suggested to explain stable isotopic data from deep geothermal wells in the Coso geothermal field (Fournier and Thompson, 1980).

Analysis of aerial and satellite photos reveals the presence of numerous lineaments between the Kern Plateau and southwestern Indian Wells Valley (Fig. 12), some of which were confirmed to be faults and fractures by field studies (Ostdick, 1997; Howard et al., 1997b). The

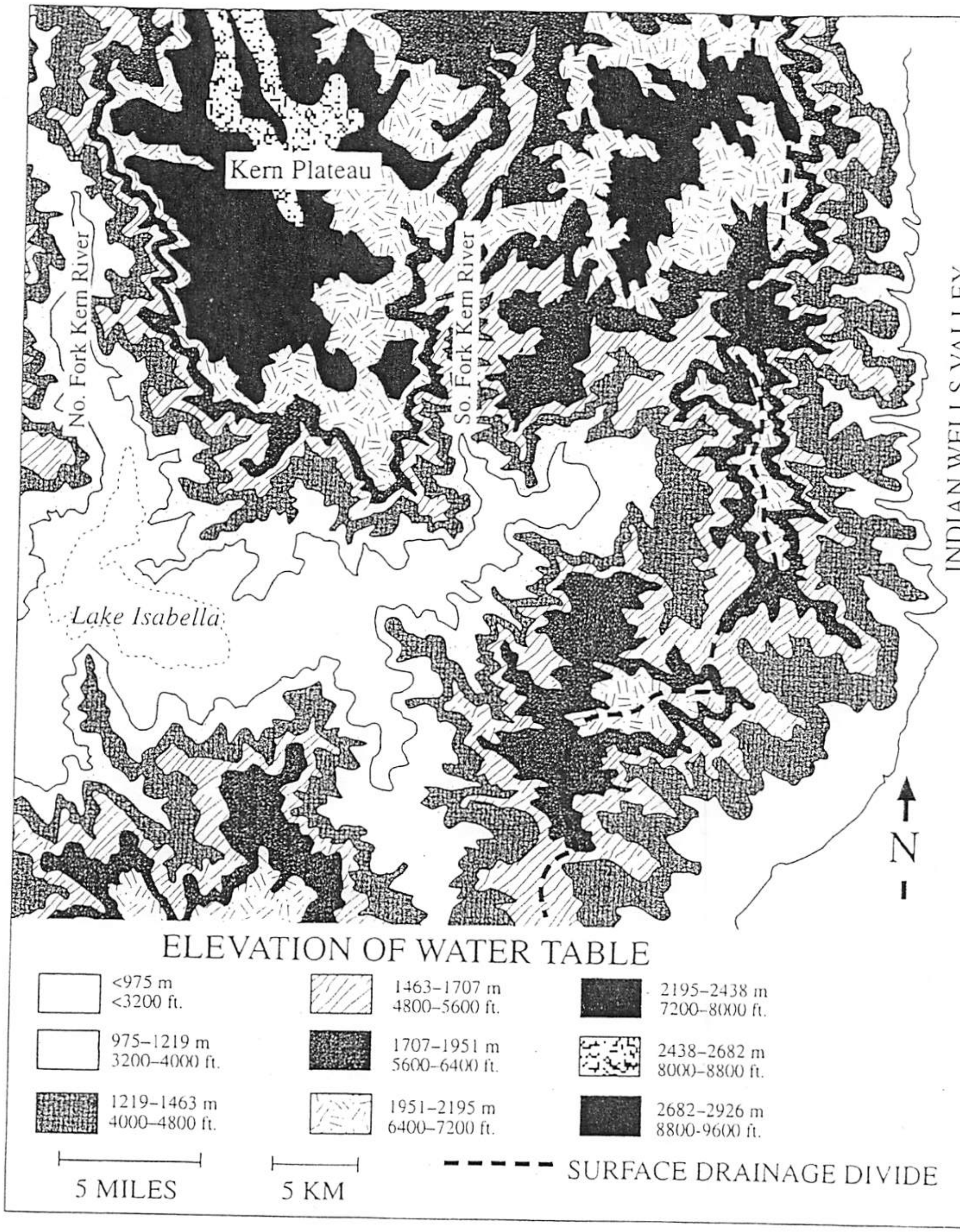

Figure 11. Map of potentiometric surface contours in the eastern Sierra and Kern Plateau areas. Map is modified from Howard et al. (1997a).

majority of the mapped faults in the area trend north and northwest (Jennings et al., 1962; Smith, 1964; Diggles et al., 1989). Other lineaments noted on aerial photos also have predominantly northwest trends, except for a small area in the northeastern part of the Kern Plateau, where northeast-trending fracture sets predominate. The fracture network associated with the sets of interconnected faults and fractures may have sufficient permeability to transport water from the Kern Plateau to the subsurface of Indian Wells Valley. Fractures in crystalline bedrock can carry the majority of recharge in some watersheds (Tiedeman et al., 1997). Because the mapped faults and fractures in Sierran watershed nearest the southwest Indian Wells Valley trend predominantly west and northwest (Fig. 12) and the water table dips predominantly east and south, Kern Plateau water moving through fractures would tend to flow southeast toward southwest Indian Wells Valley.

Field studies in the eastern Sierras show that some of the mapped faults and fractures are capable of acting as ground-water conduits (Howard et al., 1997a, 1997b). A northwest-trending fault zone in Bird Springs Canyon (Fig. 2) controls the location of a series of springs and seeps, which produces low TDS water (230 mg/l) all year from fractured granite along the trace of the fault (Ostdick, 1997). A nearly vertical, north-trending, mapped fault, which intersects Five-mile, Deadfoot, Ninemile, and Noname Canyons, controls the location of springs and seeps in these northern canyons. In Noname Canyon, the fault was dry at the surface but was observed to divert the canyon's stream into the subsurface, where it

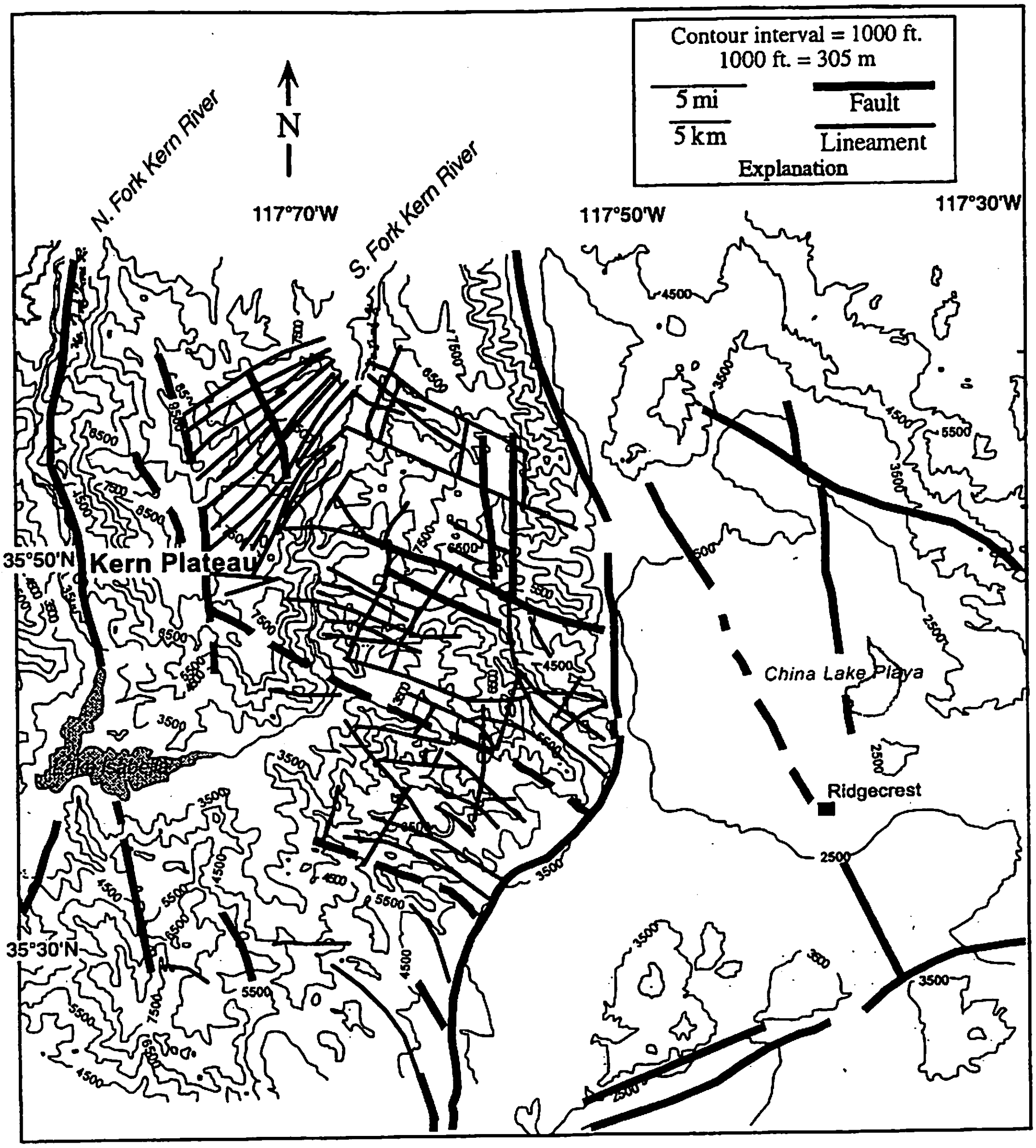

Figure 12. Map view of selected faults and lineaments in the study area. Modified from Ostdick (1997) and Figure 1 (this study).

later reemerged at the surface along subhorizontal fracture zones (Howard et al., 1997b). The gradient in this fracture was 0.03 and the water table dipped to the south.

Velocity in fractures is a function of permeability and head gradient (Domenico and Schwartz, 1998). Values for fracture permeability range from 10^{-10} to 10^{-4} m/s with watershed-scale values near the upper end of the range (Hsieh, 1995). In our hypothesis a significant fraction of the ground water in the southwest is derived from the Kern Plateau. The tritium data show that a portion of this water is post-1952, implying a fairly rapid rate for the fracture flow and travel times of <50 yr. The gradient of 0.05 and a permeability value near the high end of range (6×10^{-4} m/s) generate a velocity of 10 km/yr, more than sufficient for the proposed flow path.

Interbasin Flow

If proven, the diversion of water from the topographic watershed of the south fork of the Kern River into Indian Wells Valley would constitute interbasin flow through Sierran crystalline rocks. Eakin (1966) stated that significant interbasin flow can occur when rocks separating two watersheds with differing hydraulic heads are permeable enough to transmit fluid. Previous studies in the Basin and Range province based on geochemical and hydrologic data show that ground water travels via deep circulation in permeable carbonate bedrock between alluvial basins crossing local topographic divides (Eakin, 1966; Maxey and Mifflin, 1966; Harrill et al., 1988; Gates and Harrill, 1997; Johannesson et al., 1997).

In western Nevada, hydrologic budgets, potentiometric distributions, and water chemistry have been used to deduce interbasin flow by linking recharge from remote drainage areas to local springs with anomalous high discharge rates (Maxey, 1968). Interbasin flow in this region may amount to 28% of the total groundwater flow (Prudric et al., 1993), and forms a significant fraction of the local water budgets in some basin and range valleys (Miller, 1977; Bolke and Sumsion, 1978). This type of flow can influence geothermal features (Mifflin, 1988) and alter water table contours so that they do not conform to local topography (Farvolden, 1969). We believe that the case of southwestern Indian Wells Valley may meet the criteria for interbasin flow, specifically a high discharge rate (aquifer flux), elevated water tables (high head gradient in the southwest), and geochemical evidence of extrabasinal recharge.

In summary, we believe that the three independent lines of evidence, from the hydrogeologic, isotopic, and water chemistry data presented here, support an alternative interpretation of the hydrogeology of Indian Wells Valley. Figure 13 shows our conceptual model for the interbasin flow. If confirmed, the alternative model provides evidence that interbasin flow occurs through crystalline rocks, most likely via fractures. This would be particularly likely in the basin and range areas where crystalline rocks are affected by extensional tectonics and the adjacent mountain ranges have a main phreatic divide that does not coincide with the surface drainage divide. Therefore, if recharge estimates for alluvial aquifers in these areas are based upon models that assume that the amount of recharge to the aquifer is related to the amount of precipitation in the drainage area of the topographic watershed, they may overestimate or underestimate ground-water resources and safe yield.

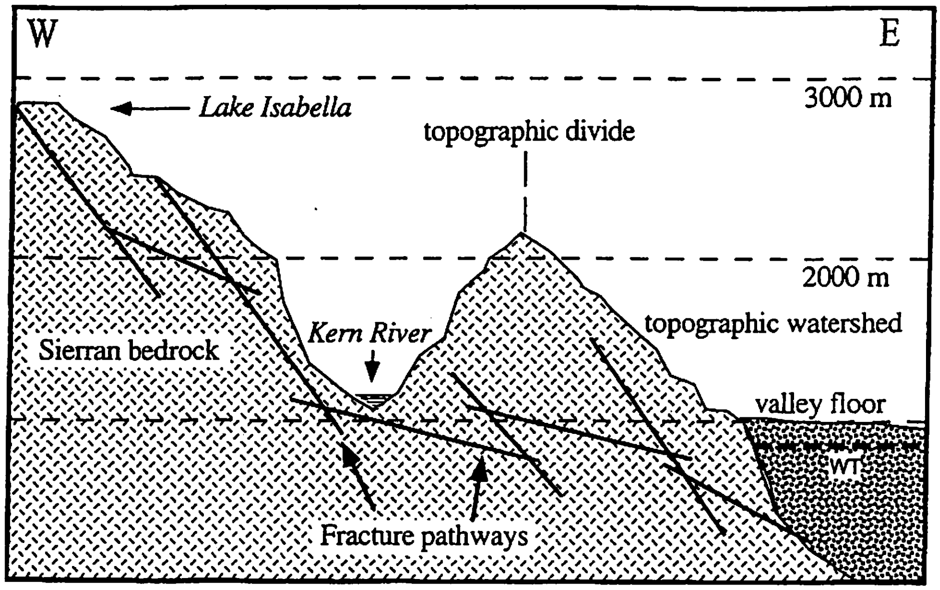

Figure 13. Conceptual model of fracture flow from the Kern Plateau into the Indian Wells Valley (not to scale). Flow crosses the local phreatic divide (Kern River) to discharge into the alluvial aquifer. WT—water table.

CONCLUSIONS

The calculated recharge values (measured precipitation values corrected for evapotranspiration) and the ground-water flux calculations suggest that the ground water in the northwest and west-central sections of Indian Wells Valley is derived from Little Lake outflow and surface runoff down the Sierran canyons in the adjacent topographic watershed. The isotopic and hydrochemical data from these sections of the valley support the hydrogeologic data.

In the southwest portion of the valley precipitation-based recharge estimates do not agree with the ground-water flux calculations because of the steep head gradient. Here, the ground-water flux through the aquifer may be as much as 10 times greater than the amount of available recharge (measured precipitation corrected for evapotranspiration) in the adjacent east Sierran watershed. The ground water in the southwestern Indian Wells Valley has a different chemical and isotopic signature than surface waters of the topographic watershed (the eastern slope of the Sierra).

Ground-water dating by tritium confirms that much of this ground water is of recent origin. The isotopic and chemical data identify the adjacent Kern River watershed, especially the higher elevation Kern Plateau area, as a possible source. Taking into account the effects of cation exchange, geochemical mixing models show that the ground water in the southwest Indian Wells Valley can be derived from mixing Kern Plateau water with local recharge.

We hypothesize that significant recharge to the southwestern Indian Wells Valley aquifer is derived from the Kern Plateau area as a result of interbasin flow along fractures in crystalline bedrock. The recharge travels from the main phreatic divide below the plateau and crosses the topographic (surface) watershed divide to recharge the aquifer in southwestern Indian Wells Valley. The Darcy calculations of flux and geochemical mixing model indicate that as much as 3.7×10^7 m^3/yr may be entering the Indian Wells Valley alluvial aquifer via interbasin flow. This represents an increase of 300% compared to most previous water budgets.

This hypothesis suggests the following.

1. Water budgets for alluvial aquifers in arid valleys separated by mountain ranges may be significantly in error if recharge estimates for these aquifers are based solely upon models that assume that the amount of recharge to an aquifer is related to the amount of precipitation in the drainage area of the topographic watershed. This would be particularly true in mountain watersheds in which the main phreatic divide does not coincide with the watershed divide.

2. Crystalline rocks, particularly those in areas affected by extensional tectonics, may transport significant quantities of water if the fractures and faults have sufficient permeability and are interconnected.

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Many of the data in this study were collected

from wells constructed by the Indian Wells Valley drilling and sampling program of the U.S. Bureau of Reclamation (1993). These wells allowed us to formulate a three-dimensional hydrological and hydrochemical picture of the aquifer. We recommend adoption of similar drilling and testing programs in alluvial aquifers scheduled for intensive ground-water development in other arid, alluvial basins.

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