GROUND-WATER FLOW AND QUALITY, AND GEOCHEMICAL PROCESSES, IN INDIAN WELLS VALLEY, KERN, INYO, AND SAN BERNARDINO COUNTIES, CALIFORNIA, 1987-88

By Charles Berenbrock *and* Roy A. Schroeder

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Conversion Factors, Vertical Datum, Water-Quality Information, Abbreviations and Acronyms, and Well-Numbering System

Conversion Factors

Multiply	By	To obtain
acre	0.405	square hectometer
acre-foot (acre-ft)	0.001233	cubic hectometer
acre-foot per year (acre-ft/yr)	0.001233	cubic hectometer per year
foot (ft)	0.3048	meter
foot per year (ft/yr)	0.3048	meter per year
gallon per minute (gal/min)	0.207	liter per second
inch (in.)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

Temperature is given in degrees Celsius (°C), which can be converted to degrees Fahrenheit (°F) by using the following equation:

$$°F=1.8(°C)+32.$$

Vertical Datum

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Water-Quality Information

Chemical concentration (solute per unit volume of water) is given in grams per liter (g/L), milligrams per liter (mg/L), or micrograms per liter (µg/L). For concentrations less than 7,000 mg/L, the numerical value in milligrams per liter is about the same as for concentrations in parts per million (ppm), and the numerical value in micrograms per liter is about the same as for concentrations in parts per billion (ppb).

Chemical concentration in terms of ionic interacting values is given in milliequivalents per liter (meq/L). Milliequivalents per liter values, which are numerically equal to equivalents per million, may be converted to milligrams per liter by multiplying milliequivalents per liter by the equivalent weight (weight of the ion divided by the ionic charge).

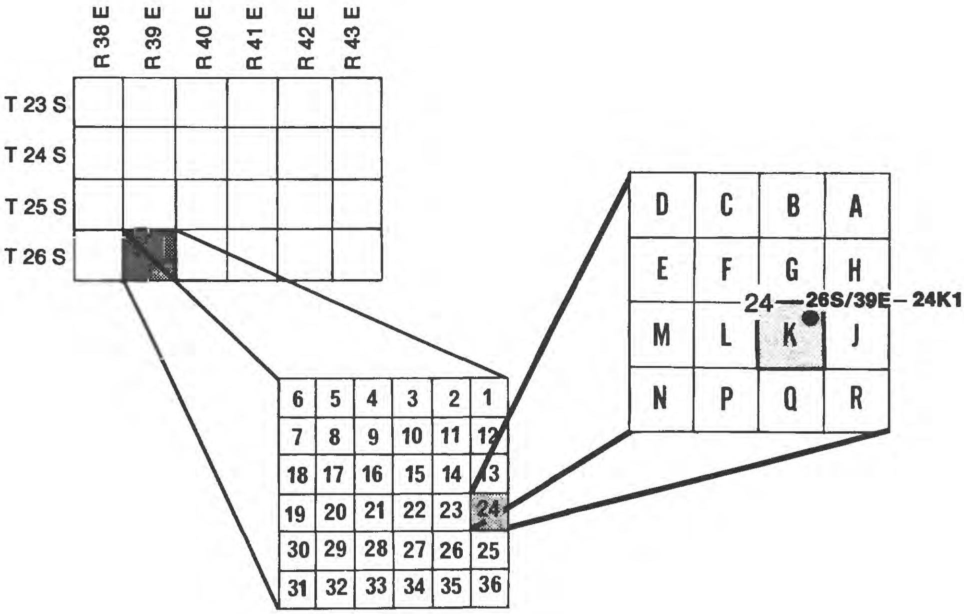
Specific conductance is given in microsiemens per centimeter (µS/cm) at 25°C. Microsiemens per centimeter is numerically equal to micromhos per centimeter.

Abbreviations and Acronyms

DSC	Dissolved-solids concentration
MCL	Maximum contaminant level
MODPATH	A computer model used to simulate advective transport
MODFLOW	A modular three-dimensional finite-difference flow model
Permil	Parts per thousand
SLAP	Standard Light Antarctic Precipitation
V-SMOW	Vienna Standard Mean Ocean Water

Well-Numbering System

Wells are identified and numbered according to their location in the rectangular system for the subdivision of public lands. Identification consists of the township number, north or south; the range number, east or west; and the section number. Each section is divided into sixteen 40-acre tracts lettered consecutively (except I and O), beginning with "A" in the northeast corner of the section and progressing in a sinusoidal manner to "R" in the southeast corner. Within the 40-acre tract, wells are sequentially numbered in the order they are inventoried. The final letter refers to the base line and meridian. In California, there are three base lines and meridians; Humboldt (H), Mount Diablo (M), and San Bernardino (S). All wells in the study area are referenced to the Mount Diablo base line and meridian (M). Well numbers consist of 14 characters and follow the format 026S039E24K01M. In this report, well numbers are abbreviated and written 26S/39E-24K1. Wells in the same township and range are referred to only by their section designation, 24K1. The following diagram shows how the number for well 26S/39E-24K1 is derived.



GROUND-WATER FLOW AND QUALITY, AND GEOCHEMICAL PROCESSES, IN INDIAN WELLS VALLEY, KERN, INYO, AND SAN BERNARDINO COUNTIES, CALIFORNIA, 1987-88

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Abstract

Aqueous chemical concentrations and hydrogen and oxygen stable-isotope ratios were measured to update ground-water quality information for Indian Wells Valley, California. Indian Wells Valley contains the China Lake playa, which is part of a chain of remnant Pleistocene lakes in the rain shadow east of the Sierra Nevada. Natural recharge to the valley originates predominantly as surface-water runoff from the Sierra Nevada, and natural discharge is by evapotranspiration at China Lake on the east-central side of the valley. Prior to development of ground-water resources, recharge was approximately balanced by discharge.

As ground water moves through alluvium near the mountains into lacustrine deposits of an ancestral lake, dissolved-solids concentration increases from about 200 to 1,000 milligrams per liter, with further increases to several thousand milligrams per liter in discharge areas (the playa) on the valley floor. Several geochemical processes alter the relative concentrations of major ions in ground water during its transit through the lacustrine deposits: alkalinity generated by sulfate reduction, precipitation of alkaline-earth carbonates, removal of aqueous alkaline-earth ions by exchange for sodium on clays, and dissolution of evaporites such as sodium chloride.

Stable oxygen- and hydrogen-isotope ratios in ground water display a range of values determined by location and timing of recharge and by evaporation history. Delta deuterium in ground water from alluvium near the Sierra Nevada ranges from about -90 permil (parts per thousand) in the south to about -100 permil in the north. This trend probably is caused by an increase northward in altitude of the mountains. Evaporation from a shallow water table on the valley floor enriches deuterium to about -80 permil in shallow ground water. Comparison of isotope ratios in current precipitation and recharge to ratios in deep ground water and in ground water that has undergone evaporation suggests that some of the water present was recharged during an earlier (late Pleistocene) pluvial period. Delta deuterium in recharge for this old ground water is estimated to have been 15 to 35 permil more negative than in current recharge.

Increased pumping since 1920 in the most highly developed part of the valley at Ridgecrest, and in the area (referred to as the "intermediate area") between Ridgecrest and Inyokern, has resulted in a 100-foot-deep cone of depression and has induced movement of comparatively saline ground water toward the depression from playas to the east (Mirror and Satellite Lakes) and north (China Lake) and of fresh ground water from Little Dixie Wash to the west (from which flow occurred naturally prior to pumping). Data document a trend of increasing dissolved solids in water from several wells, since at least the early 1970's, in the areas of heaviest pumping. Induced flow of saline ground water from deposits in the Mirror and Satellite Lakes area is likely to be high in sulfate relative to chloride; flow from the China Lake area is high in chloride relative to sulfate. The difference in relative abundance of sulfate and chloride between these playas reflects differences in the intensity and extent of sulfate reduction that has occurred beneath each area. A potential exists for intensified degradation in ground-water quality in the Ridgecrest and intermediate areas as a result of movement of saline water from areas near the playas if pumping continues at current or increased rates.

Although there are insufficient data to construct a solute-transport simulation model, a simplified advective-transport model was used to qualitatively evaluate the effects of pumping on ground-water flow paths and traveltimes. Several simulations were completed using MODPATH, a particle-tracking program. First, the potential for saline ground water to migrate from China, Mirror, and Satellite Lakes into the Ridgecrest and intermediate areas was assessed. Second, the effect of pumping on natural recharge from Little Dixie Wash was assessed. Results suggest that movement of ground water from Little Dixie Wash to the natural discharge areas takes several thousand years. Such long traveltimes are consistent with the probable presence of late Pleistocene ground water in deep zones and in discharge areas of the valley.

INTRODUCTION

Ground water is virtually the sole source of water for residents of Indian Wells Valley (fig.1). Since 1959, annual pumpage from the aquifer has exceeded estimates of mean annual recharge (Berenbrock and Martin, 1991, p. 39), and continued increases in pumpage are expected. In areas of greatest pumpage, water levels in the aquifer have declined about 30 ft since 1960. Previous investigators (Dutcher and Moyle, 1973; Warner, 1975; Mallory, 1979; Berenbrock and Martin, 1991) have indicated that water-level declines could result in degradation of the ground-water quality in the areas of heaviest pumpage.

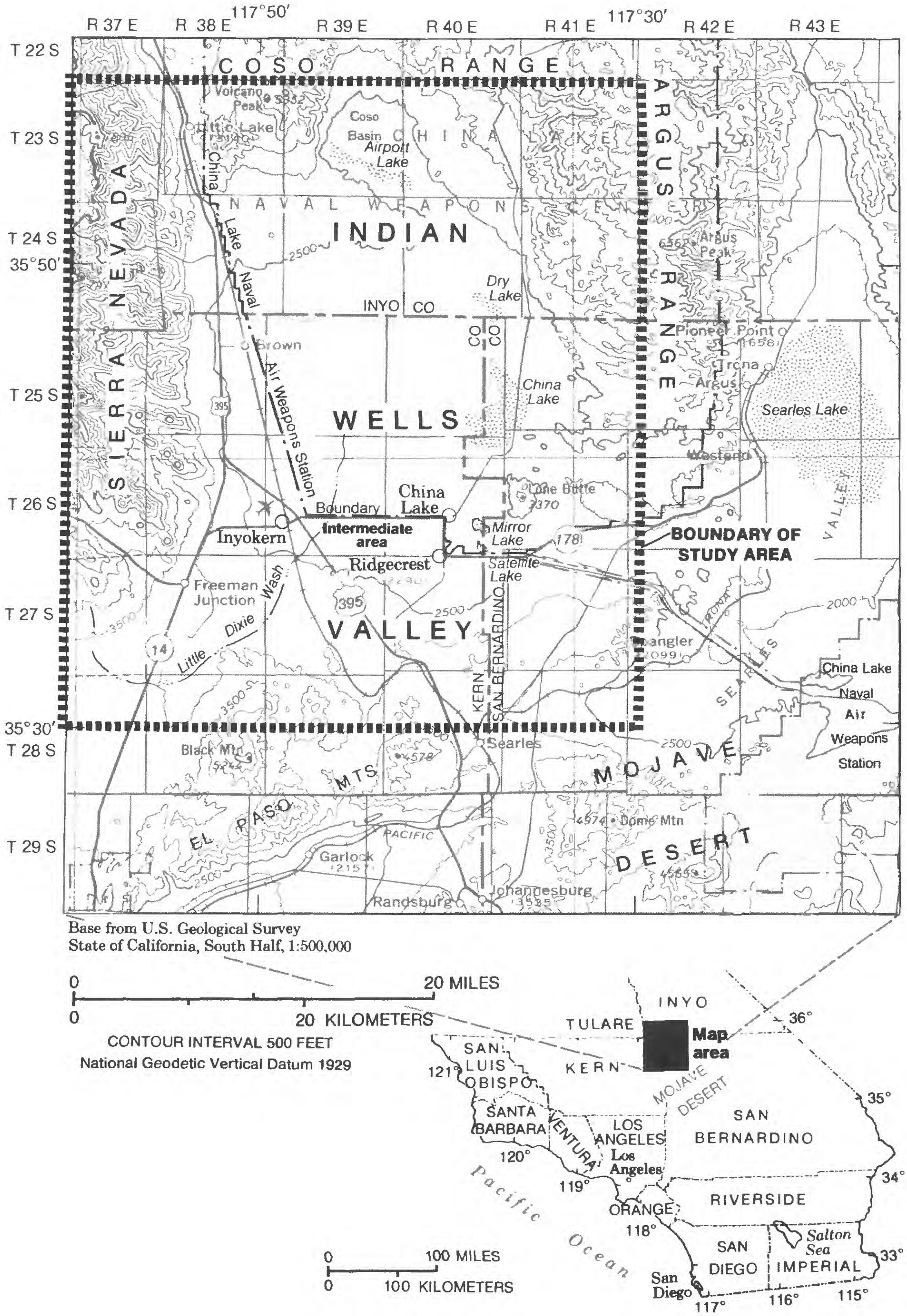
In 1988, a three-dimensional mathematical ground-water flow model was used by the U.S. Geological Survey (USGS) to simulate flow through two aquifers: a shallow aquifer that occupies the east-central part of the valley where it overlies a deep aquifer, and a deep aquifer which extends throughout the unconsolidated deposits of the valley and which is the source of pumpage (Berenbrock and Martin, 1991). The model has proven to be a useful tool to simulate water-level changes in the two aquifers. Berenbrock and Martin (1991) indicated that ground-water pumping had caused water levels in the deep aquifer to decline almost 100 ft from 1920 to 1985 in the intermediate area (defined in this and previous reports as the area between Ridgecrest and Inyokern) and had reversed the direction of ground-water movement in the deep aquifer in the area southwest of China Lake. Ground water of poor quality (that is, high dissolved-solids concentration) underlies China Lake; therefore, water quality in the deep aquifer south and west of China Lake, where pumpage is greatest, is subject to degradation.

A comprehensive ground-water-quality study completed by the USGS in 1972 (Warner, 1975) indicated that pumping had not affected water quality. Data collected since that study, however, indicate that ground-water quality has degraded significantly in the areas of greatest pumpage in the southeastern part of the valley near Ridgecrest. Dissolved-solids concentrations of ground water throughout most of the Ridgecrest area currently exceed the secondary maximum contaminant level of 500 milligrams per liter (mg/L) for drinking water (U.S. Environmental Protection Agency, 1979). Therefore, current ground-water-quality conditions need to be compared with historical data to determine the degree of change that has occurred; to determine water quality in areas not currently developed for public supply; and to determine the potential for further degradation. An assessment of water quality in the valley provides local water districts and the military with information that will assist them in providing needed water and maintaining adequate quality. The U.S. Geological Survey, in cooperation with the Indian Wells Valley Water District and the U.S. Department of the Navy, China Lake Naval Air Weapons Station (formerly China Lake Naval Weapons Center), began a ground-water-quality assessment of Indian Wells Valley in 1987. Insight gained from this study can be used to design water-level and water-quality monitoring networks, collect additional geohydrologic and geochemical data where needed, and design more sophisticated models (flow and solute-transport) for future management of ground-water resources in Indian Wells Valley.

PURPOSE AND SCOPE

The purpose of this report is to present an appraisal of ground-water flow conditions and current (1987-88) ground-water quality, compare current ground-water quality with historical conditions, and assess the potential for future degradation of water quality. The report includes a discussion of how various geochemical processes determine the chemical composition and how recharge and evaporation determine the isotopic composition of ground water in Indian Wells Valley.

The study included: (1) collecting and analyzing ground-water samples from wells for major ions, selected minor constituents, and stable hydrogen and oxygen isotopes; (2) describing areal and vertical variations of ground-water quality in the valley, with emphasis on dissolved solids; (3) comparing ground-water quality in 1987-88 with historical ground-water quality; (4) using a particle-tracking model to develop a geochemical and physical framework in order to evaluate present knowledge and concepts of the aquifer system and processes of solute transport; and (5) estimating the isotopic composition of recharge during the late Pleistocene.



Base from U.S. Geological Survey
State of California, South Half, 1:500,000

CONTOUR INTERVAL 500 FEET
National Geodetic Vertical Datum 1929

Figure 1. Location of study area.

DESCRIPTION OF STUDY AREA

Indian Wells Valley (fig. 1) is in the western part of the Mojave Desert in southern California, about 125 mi north of Los Angeles. The valley is bounded on the west by the Sierra Nevada, on the north by a low ridge of volcanic rocks and the Coso Range, on the east by the Argus Range, and on the south by the

El Paso Mountains. The surrounding mountains and hills slope steeply to the broad valley floor, which slopes gently toward China Lake, a large playa in the east-central part of the valley. The valley floor, most of which is between 2,175 and 2,400 ft in altitude, is about 300 mi^2 in area. The valley drains internally into China Lake, which is at the lowest altitude in the valley (about 2,150 ft). There is no perennial surface flow on the valley floor, and China Lake is dry except during rare extremely wet periods.

Indian Wells Valley has an arid climate; average rainfall on the valley floor is 4 to 6 in. Most of the precipitation occurs during October through March, with infrequent rainfall during summer and occasional snowfall during winter. Mean temperatures range from about 4°C in winter to about 28°C in summer, and average diurnal range is about 17°C throughout the year (Evert and Burgin, 1986, p. 11). The main growing season is April through October.

The population of Indian Wells Valley was about 25,000 in 1988. The principal communities, and their approximate populations in 1988, are: Ridgecrest, 22,000; Inyokern, 1,000; and China Lake, 500.

The source of water for public drinking-water supplies, for irrigation, and for military and industrial uses in Indian Wells Valley is ground water. Greatest pumpage is in the intermediate area between Ridgecrest and Inyokern (fig. 1). Many private wells serve as domestic drinking-water and irrigation supplies. In 1985, about 21,500 acre-ft of ground water was pumped from the valley (Berenbrock and Martin, 1991). The pumpage included about 7,000 acre-ft for municipal and domestic supplies; 6,900 acre-ft for irrigation (mostly in the northwestern part of the valley); 4,800 acre-ft for military use at China Lake Naval Air Weapons Station; and 2,800 acre-ft for industrial uses.

GEOHYDROLOGIC SETTING

Indian Wells Valley is a structural and topographic depression in the southwestern part of the Basin and Range Province. China Lake, which occupies part of Indian Wells Valley, is one member in a chain of remnant Pleistocene lakes in the rain shadow east of the Sierra Nevada in central California. Indian Wells Valley is located between the much smaller Rose and Salt Wells Valleys to the northwest and southeast, respectively. During the last pluvial episode, an ancestral lake covered Indian Wells Valley up to an altitude of 2,400 ft (St.-Amand, 1986, p. 12), permitting exchange of surface water and ground water between adjoining valleys. Today, Indian Wells Valley is virtually closed, and there is very little ground-water underflow to or from adjacent valleys (Warner, 1975, p. 11).

As the ancestral lake that occupied Indian Wells Valley receded, it left behind lacustrine deposits and several playas on the valley floor. The largest of the playas is China Lake in the east-central part of the valley. Several smaller playas occupy local topographic depressions in the valley (fig. 1). Mirror and Satellite Lakes to the south of China Lake are two playas of special significance to this study because ground water beneath them has a chemical composition that is different (conditions are less reducing) from that of water beneath China Lake, and because saline ground water in their lacustrine deposits is the source of degradation of ground-water quality in the Ridgecrest area.

For this study, the lithologic units mapped by Von Huene (1960), Zbur (1963), Moyle (1963), and Kunkel and Chase (1969) are grouped into "consolidated rocks" and "unconsolidated deposits" (fig. 2). The consolidated rocks include the basement complex, continental deposits, and volcanic rocks. The basement complex consists of pre-Tertiary igneous and metamorphic rocks that underlie the ground-water basin and crop out in the surrounding hills. The continental deposits of Tertiary age overlie the basement complex and are nearly impermeable except where fractured. The volcanic rocks include the Miocene Black Mountain Basalt near the El Paso

EXPLANATION

GEOLOGIC UNITS

UNCONSOLIDATED DEPOSITS

Quaternary–

▨ Sand dunes (Holocene)

□ Sand dunes and playa deposits (Holocene)

□ Playa deposits (Holocene)

□ Alluvium (Holocene and Pleistocene)

⁘ Lacustrine deposits (Pleistocene)

CONSOLIDATED ROCKS–

Quaternary and Tertiary

■ Volcanic rocks (Pleistocene and Miocene)

Tertiary–

■ Continental deposits (Pliocene and Miocene)

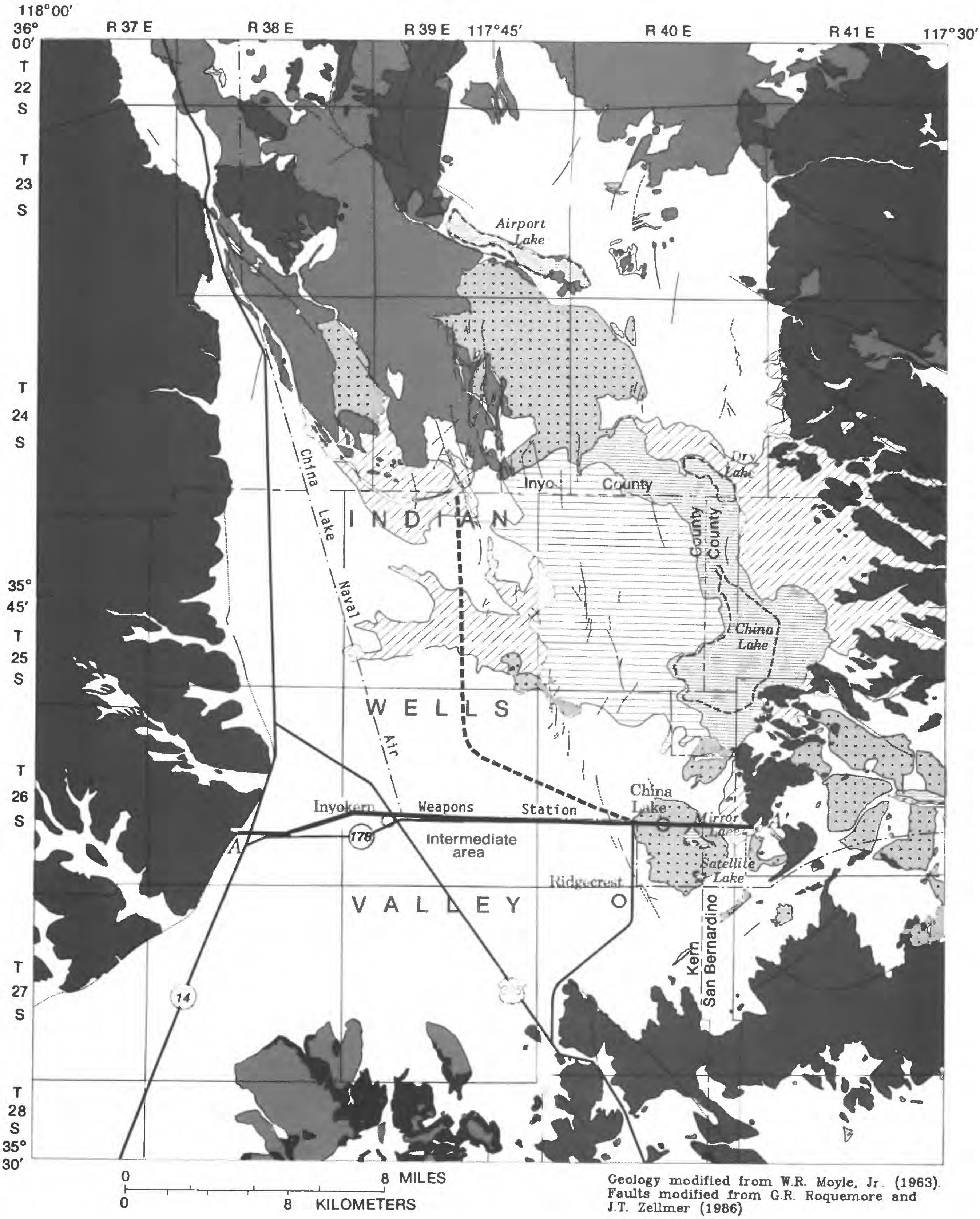
Pre-Tertiary–

■ Basement Complex

- - - FAULT–Dashed where approximately located

A——A′ LINE OF SECTION SHOWN IN FIGURES 16 AND 17

------ BOUNDARY OF SHALLOW AQUIFER (Kunkel and Chase, 1969)–Dashed where approximately located



Geology modified from W.R. Moyle, Jr. (1963). Faults modified from G.R. Roquemore and J.T. Zellmer (1986)

Figure 2. Generalized geology and location of geologic section in Indian Wells Valley. Modified from Berenbrock and Martin (1991).

Mountains and Pleistocene volcanic rocks near the Coso Mountains (Kunkel and Chase, 1969, p. 22; Duffield and Bacon, 1981; and Diggles and others, 1985, p. C6). The consolidated rocks form the lower boundary of the ground-water basin and much of the perimeter boundary.

The unconsolidated deposits include lacustrine, alluvium, playa, and sand-dune deposits. The lacustrine deposits of Pleistocene age and the playa deposits of Holocene age consist predominantly of silt and clay and are interbedded with and overlie the alluvium. The alluvium of Pleistocene and Holocene age includes the older and younger alluvium, alluvial fans, and stream-terrace deposits. The alluvium consists of unconsolidated gravel, sand, silt, and clay. The percentage of silt and clay in the alluvium generally increases toward the central and eastern parts of the valley. The sand-dune deposits, of Holocene age, consist of windblown sand.

The greatest thickness of unconsolidated deposits, about 2,000 ft, is in the west-central part of the valley (Zbur, 1963; Dutcher and Moyle, 1973, p. 9). The unconsolidated deposits vary greatly in lithology, both vertically and areally, especially in the central and eastern parts of the valley. The deposits in the southwestern part of the valley are more uniform, consisting primarily of fine to coarse sand and small amounts of silt. The absence of large amounts of silt and clay in the southwestern part of the valley indicate that the ancestral China Lake did not extend to this part of the valley.

On the basis of geophysical and geologic logs of selected wells, previous investigators have divided the unconsolidated deposits in the valley into two main aquifers: (1) The shallow aquifer (shallow water body of Kunkel and Chase, 1969) and (2) the deep aquifer (main water body of Kunkel and Chase, 1969).

The shallow aquifer includes lacustrine, alluvium, playa, and sand-dune deposits. Kunkel and Chase (1969) defined the shallow aquifer as extending from China Lake westward to the center of the valley and from the area south of Airport Lake southward to the community of China Lake. The fine-grained deposits of the shallow aquifer are as much as 300 ft thick (Berenbrock and Martin, 1991, fig. 4) where they overlie the deep aquifer in the eastern part of the valley. Ground water in the deep aquifer is confined or partly confined by the shallow deposits. The shallow aquifer generally does not yield water freely to wells although water from three shallow wells (well depth of about 55 ft) near the Ridgecrest Regional Wastewater Sewage Treatment evaporation ponds is pumped at a rate of about 60 gal/min (Don L. Greenfield, Supervisor, Ridgecrest Wastewater Treatment Facilities, oral commun., 1988). Water from the shallow aquifer is used only for fire protection and maintenance of a few buildings on China Lake Naval Air Weapons Station. Since 1987, the water table in the shallow aquifer has been artificially lowered by pumping around the Ridgecrest Regional Wastewater Treatment Facility ponds to prevent structural damage to nearby buildings resulting from high-water-table conditions.

The deep aquifer consists of alluvium and extends, in the eastern part of the valley, from the base of the shallow aquifer to consolidated rocks; in the western part of the valley, where the shallow aquifer is not present, the deep aquifer extends from land surface to the consolidated rocks. Kunkel and Chase (1969, p. 39) indicated that the total saturated thickness of the deep aquifer in the central part of the valley is at least 1,000 ft. The deep aquifer consists predominantly of fine to coarse sand and gravel of high permeability; in the central and eastern parts of the valley, however, it is interbedded with silt and clay layers of lacustrine deposits. The deep aquifer is the principal source of ground water in Indian Wells Valley, and it yields more than 1,000 gal/min to individual municipal and irrigation wells. Some wells in the intermediate area and at Inyokern yield more than 2,000 gal/min.

Before pumping began in the valley, ground water moved from areas of recharge along the margins of the valley toward China Lake through the deep aquifer and into the shallow aquifer (Warner, 1975, p. 12). Thus, leakage from the deep aquifer was virtually the only source of recharge to the shallow aquifer, and it was the only significant discharge from the deep aquifer. Discharge of ground water from the shallow aquifer was principally by evapotranspiration in and around China Lake.

Virtually all natural recharge to the ground-water system in Indian Wells Valley occurs as infiltration of surface runoff from the Sierra Nevada. Most of the runoff infiltrates near the mountain front where streams first cross the unconsolidated deposits of the valley, and the recharge is termed, in this report, "mountain-front recharge." Prior to pumping in the valley, mountain-front recharge to the ground-water system was approximately balanced by natural discharge, most of which occurred by evapotranspiration in and around China Lake. The evapotranspiration cannot be estimated precisely, but it probably averaged about 8,000 to 11,000 acre-ft/yr prior to 1920 (Berenbrock and Martin 1991, p. 14). Bloyd and Robson (1971, p. 12) estimated recharge and discharge under steady-state conditions to be 9,850 acre-ft/yr. About 6,300 acre-ft/yr of the recharge was attributed to infiltration from streams originating in the Sierra Nevada west of the valley, about 3,200 acre-ft/yr to infiltration from streams originating in the Coso and Argus Ranges north and northeast of the valley, and about 400 acre-ft/yr to infiltration from streams originating in the El Paso Mountains south of the valley. The estimated long-term average recharge from streamflow infiltration of about 10,000 acre-ft/yr was used in a ground-water flow model by Berenbrock and Martin (1991) for the period 1920-85; however, the results from their model simulations indicate that discharge by evapotranspiration had declined, as a result of pumping, to about 6,600 acre-ft/yr by 1985.

Since the 1920's, when pumpage became significant, ground-water movement in the deep aquifer has been altered by changes in water level. Water-level measurements made during spring 1988 (Supplemental Data A) indicate that pumping from the deep aquifer has caused a cone of depression in the intermediate and Ridgecrest areas (fig 3). Water level in the deep aquifer has declined as much as 100 ft in the center of the depression. Comparison of water levels for 1985 (Berenbrock and Martin, 1991, fig. 8) and 1988 (this report, fig. 3) indicates an additional decline of about 10 ft in the cone of depression.

Water-level measurements made during spring 1988 in wells tapping the shallow aquifer do not show any depressions in the water table that can be related to pumping (fig. 3). In fact, the water levels in spring 1988 were similar to those of 1920-21 (Berenbrock and Martin, 1991, fig. 7), except near the community of China Lake. Berenbrock and Martin (1991, p. 17) attributed the higher (1985) water levels near the community of China Lake to tree and shrubbery watering that has occurred since the 1950's; they estimated that about 90 acre-ft/yr of water is recharged to the shallow aquifer from tree and shrubbery watering.

Large hydraulic-head differences between wells in the shallow and deep aquifers indicate that ground water currently moves from the shallow aquifer into the deep aquifer. On the basis of model results, Berenbrock and Martin (1991, p. 43), estimated that 1,550 acre-ft flowed from the shallow aquifer into the deep aquifer during 1985; by contrast, prior to ground-water development in the valley, ground water flowed from the deep aquifer to the shallow aquifer.

METHODS OF INVESTIGATION

SAMPLING AND CHEMICAL ANALYSES

More than 80 wells were sampled and analyzed for major ions and selected minor constituents during 1987-88. Location of these wells is shown on plate 1. About two-thirds of the wells were sampled and analyzed by the USGS and one-third were sampled and analyzed by other agencies. The results of these analyses are given in Supplemental Data B. More than one-half of the sampled wells are (USGS oragency) observation wells. The ownership (or use) of the remaining wells, in descending order by number of wells, is: municipal, military, agricultural, domestic, industrial, and institutional (designated in Supplemental Data B). Some wells were sampled several times, but about 40 percent of the wells were sampled only once by the USGS during this study. Criteria for selection of sampling sites were based on areal distribution, availability of previous water-quality records, and accessibility to sites. An attempt was made to sample all wells that had been sampled during 1972 by Warner (1975); however, many of the wells could not be resampled in 1987-88 because they had either been destroyed or were no longer in service. Twenty of the wells sampled in 1972 (designated in Supplemental Data B) were resampled for this study.

Ground-water samples collected by the USGS were analyzed for inorganic constituents at the USGS National Water-Quality Laboratory in Arvada, Colorado, using standard methods described by Fishman and Friedman (1989). Many of these samples were tested for degree of saturation (saturation indices were calculated) with respect to gypsum, alkaline-earth carbonates, and fluorite using a computerized thermodynamic data base (Plummer and others, 1984). A few samples were analyzed for purgeable organic priority pollutants using methods described by Wershaw and others (1987). These results are described in a later section of this report. Samples from selected wells, springs, and surface-water sites obtained during 1985-89 were analyzed for stable hydrogen- and oxygen-isotope ratios. These results are given in tables 1 and 2.

MEASURES OF DISSOLVED-SOLIDS CONCENTRATION

Dissolved-solids concentration (DSC)[1] was calculated as the sum of soluble chemical constituents. For samples with incomplete chemical data (not all major constituents were analyzed), DSC can be estimated from the relation between DSC and specific conductance. The reliability of this estimate is best when the relative proportion of all major ions is about the same in all samples (Hem, 1985). Using data from this study, the slope of the regression between DSC, in milligrams per liter (mg/L), and specific conductance, in microsiemens per centimeter at 25°C (µS/cm), was found to be 0.75 for samples from the shallow aquifer and 0.67 for samples from the deep aquifer. This small difference reflects differences in relative proportions of major ions typically present in high-DSC and low-DSC water associated with the shallow aquifer and deep aquifer, respectively, and the

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[1] Commonly referred to as "total dissolved solids (TDS)."

non-ideal thermodynamic effects exerted by increasing ionic strength on electrochemical properties (decrease in activity coefficient and increase in ion-pair association with increase in concentration).

Three indicators of DSC are used interchangeably throughout this report although they are not precisely equal to one another: the sum of soluble chemical constituents computed directly from analyses of the individual constituents, residue on evaporation at 180°C, and specific conductance (as related to DSC) at 25°C. Hem (1985) discusses the various causes of small differences between computed sum of constituents and residue on evaporation (p. 156-157), and the relation between specific conductance and dissolved-solids concentration (p. 66-68). Historically, the USGS has used the sum of constituents to represent DSC of ground water in Indian Wells Valley; China Lake Naval Air Weapons Station and consulting agencies have used residue on evaporation (Koehler, 1971, p. 12; Warner, 1975, p. 20). To provide an equivalent measure of DSC in this report for ground-water samples collected by agencies other than the USGS and listed in Supplemental Data B, available data from these other agencies were used to compute sum-of-constituents values. Because silica usually was not analyzed by other agencies, the computed sum is slightly less than the total amount actually present. Both the sum-of-constituents and the residue-on-evaporation values for data from other agencies are given in Supplemental Data B.

HYDROGEN- AND OXYGEN-ISOTOPE ANALYSES

In addition to major-ion chemical constituents, the stable-isotope ratios of hydrogen (hydrogen-2, or deuterium, and hydrogen-1, or protium) and oxygen (oxygen-18 and oxygen-16) were determined for samples of ground water from Indian Wells Valley and surface water and spring water from the Coso and Argus Ranges and the Sierra Nevada (except oxygen isotopes were not determined in the Sierra Nevada samples). These isotopes commonly are used in hydrologic investigations to determine sources of recharge, evaporation, and mixing of ground water that has different isotopic signatures (see Fritz and Fontes, 1980, for applications).

For this study, isotope analyses were done on samples from 55 wells in Indian Wells Valley; 2 springs, 4 creeks, and 1 lake in the Sierra Nevada; and 8 springs in the Coso and Argus Ranges. Results are given in tables 1 and 2 using the conventional delta (δ) notation of Craig (1961b), in which δ is the deviation, in parts per thousand (permil), of the isotope ratio in the sample from the isotope ratio in Vienna Standard Mean Ocean Water (V-SMOW) normalized on scales such that $\delta^{18}O$ and δD values of Standard Light Antarctic Precipitation (SLAP) are -55.5 permil and -428 permil, respectively (Coplen, 1988). Hence, higher (less negative) values of $\delta^{18}O$ and δD represent enrichment in the heavier isotope of oxygen and hydrogen, respectively; and lower (more negative) δ values represent enrichment in the lighter isotope (depletion in the heavier isotope). The 2σ precision of oxygen- and hydrogen-isotope results is about 0.1 permil and 1.5 permil, respectively. Hydrogen-isotope-ratio analyses were done by the zinc hydrogen-gas generation technique (Kendall and Coplen, 1985). Oxygen-isotope-ratio analyses were done by a CO_2-equilibration technique (Epstein and Mayeda, 1953), which yields activities rather than concentrations. For samples with DSC less than about 100 g/L (grams per liter) and total alkaline-earth concentration less than about 10 g/L, such as those from Indian Wells Valley, concentrations and activities differ by less than the reported precision (Sofer and Gat, 1975).

Springs are more suitable indicators of the isotopic composition of recharge than are surface waters or precipitation because they represent shallow ground-water recharge after it is sufficiently below land surface to no longer be subjected to the effects of evaporation. Isotope ratios (table 2) for seven of the eight springs in the Coso and Argus Ranges plot (fig. 4) close to the global meteoric-water line of Craig (1961a). The lone exception is Coso Village Spring (δD= -85 permil and $\delta^{18}O$= -9.25 permil), which issues beneath a vernal pool and thus included standing water from the pool itself when sampled. The effect of evaporation from the pool is apparent in the higher (less negative) δ values; in addition, this sample (H in fig. 4) plots far to the right of (below) the global meteoric-water line. The seven other springs indicate a range in δD for recharge from about -90 to about -100 permil.

EXPLANATION

☐	UNCONSOLIDATED DEPOSITS
▨	CONSOLIDATED ROCKS
	WATER-LEVEL CONTOUR—Shows altitude, in feet above sea level. Contour interval 10 feet. Dashed where approximately located; queried where doubtful
2170 ———?—	Shallow aquifer
2180 ———?—	Deep aquifer
	DIRECTION OF GROUND-WATER MOVEMENT
→	Shallow aquifer
➞	Deep aquifer
	WELL AND NUMBER—See Supplemental Data A
14B1 ○	Shallow aquifer
● 1K1	Deep aquifer

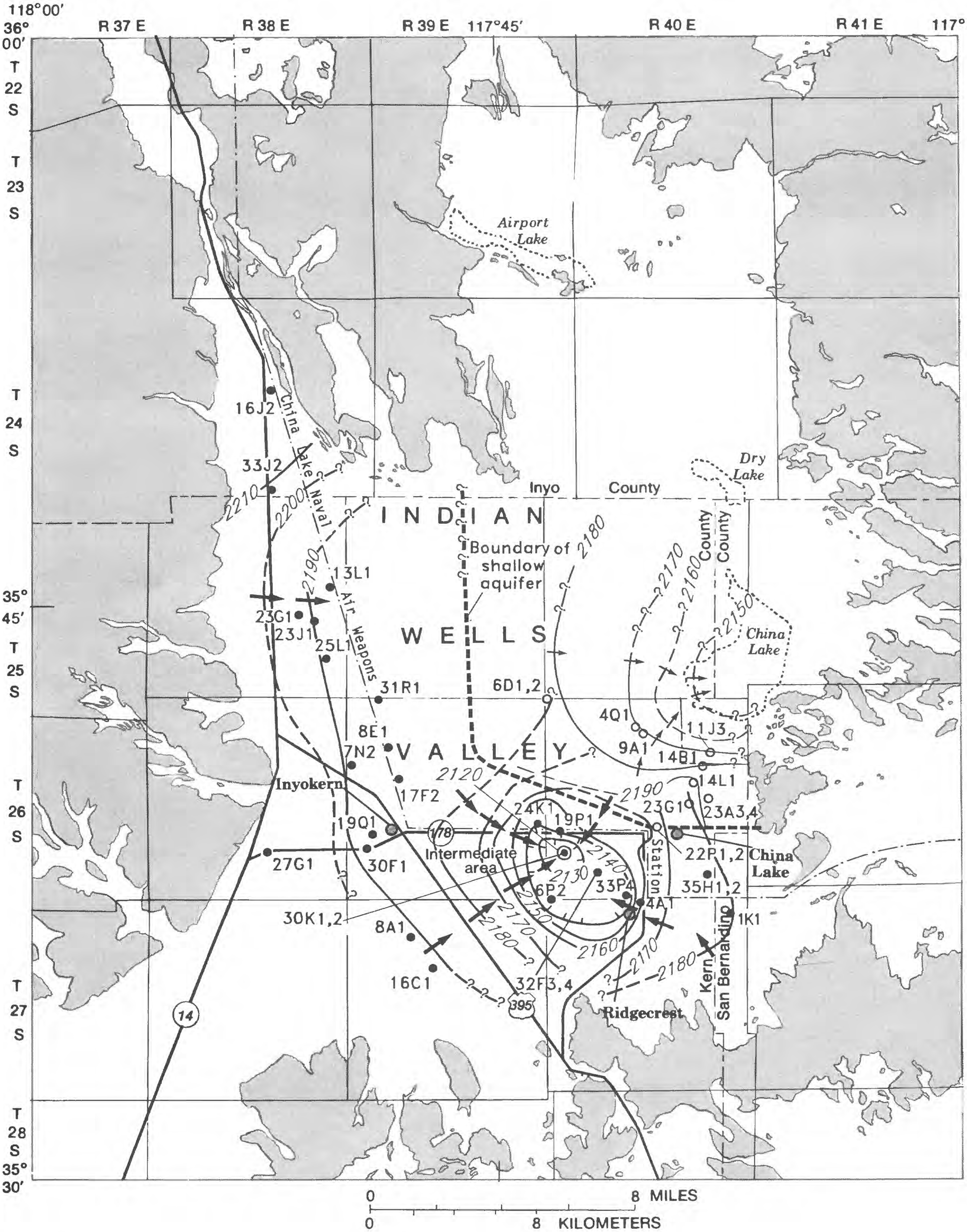


Figure 3. Water-level contours and direction of ground-water movement in the shallow and deep aquifers of Indian Wells Valley ground-water basin, spring 1988.

Table 1. Stable-isotope ratios for water samples from selected wells in Indian Wells Valley

[Altitude of land surface is given in feet above sea level. Depth of well and perforated interval are given in feet below land surface. µS/cm, microsiemens per centimeter at 25°C; permil, parts per thousand deviation from VSMOW (Vienna Standard Mean Ocean Water); --, no data available]

Well number	Altitude of land surface	Depth of well	Perforated interval	Sample date	Specific conductance (µS/cm)	Delta deuterium (permil)	Delta oxygen-18 (permil)	Monitored aquifer
				Northwestern Area				
24S/38E-16J2	2,585	611	251-611	01-22-86	2,010	-95.0	-11.60	Deep
24S/38E-33J2	2,480	375	240-375	01-23-86	--	-91.5	-12.00	Deep
24S/39E-33N1	2,354.5	161.4	--	01-10-86	1,620	-76.0	-10.40	Deep
25S/38E-11L1	2,445	400	--	07-23-87	540	-87.0	-11.65	Deep
25S/38E-23J1	2,376	630	240-630	04-16-86	660	-86.0	-11.80	Deep
				Western Area				
25S/38E-25J1	2,275	330	120-330	04-16-86	1,260	-90.5	-11.60	Deep
25S/38E-36A1	2,291	285	139-285	04-16-86	1,330	-94.0	-11.85	Deep
25S/38E-36B1	2,293	400	200-400	04-17-86	1,350	-91.5	-12.00	Deep
25S/39E-31D1	2,267	300	140-300	04-16-86	1,540	-90.0	-11.70	Deep
26S/38E-27G1	2,900.68	723	663-723	09-17-85	440	-93.0	-11.20	Deep
26S/38E-35B1	2,575	400	340-400	01-23-86	291	-104.0	-13.55	Deep
26S/39E-7N2	2,395.3	368	--	04-15-86	935	-94.0	-12.10	Deep
26S/39E-17F2	2,340	881	681-881	07-22-87	300	-102.0	-13.55	Deep
26S/39E-19P1	2,416	420.7	--	09-18-85	470	-94.5	-12.60	Deep
				Southwestern Area				
26S/37E-26L1	4,320	50	--	07-23-87	400	-90.5	-12.15	Deep
27S/38E-1G1	2,255	399	344-399	09-26-85	635	-84.0	-11.00	Deep
27S/39E-8M1	2,258	--	--	07-02-87	420	-91.5	-12.00	Deep
27S/39E-8M2	2,258	1,000	400-1,000	06-06-89	435	-92.5	-12.45	Deep
27S/39E-16C1	2,582.3	502	370-502	01-18-86	427	-93.0	-12.55	Deep
				China Lake Area				
25S/40E-20F1	2,179.5	182.6	--	01-09-86	675	-80.0	-10.50	Shallow
26S/40E-1A2	2,157.6	197.5	80-100; 110-130; 170-190	06-17-85	18,400	-97.5	-12.60	Shallow
26S/40E-4Q1	2,185	290	30-50; 70-90; 110-130	07-22-87	890	-94.5	-12.50	Shallow
26S/40E-6C1	2,195	620	500-600	07-24-87	90,000	-90.0	-10.60	Deep
26S/40E-6D1	2,216	320	276-300	07-24-87	21,300	-92.5	-12.05	Deep
26S/40E-11J3	2,174	7.6	--	06-11-85	15,000	-94.5	-11.55	Shallow
26S/40E-14B1	2,186.5	22	20-22	06-11-85	4,200	-80.0	-8.95	Shallow
26S/40E-14L1	2,201	57	55-57	06-11-85	2,600	-86.0	-10.55	Shallow
26S/40E-15N2	2,234.8	101	99-101	06-11-85	4,150	-101.0	-13.00	Shallow
26S/40E-17J1	2,262.35	97	95-97	06-10-85	410	-92.5	-12.00	Shallow
26S/40E-17R1	2,266.65	101	99-101	06-10-85	440	-91.0	-12.00	Shallow
26S/40E-21A1	2,250.58	104	102-104	06-10-85	760	-102.0	-12.65	Shallow
26S/40E-21E1	2,272.57	114	112-114	06-10-85	340	-92.5	-12.30	Shallow
26S/40E-22B1	2,232.45	63	61-63	06-12-85	6,500	-97.0	-12.55	Shallow
26S/40E-22H1	2,226.62	49	47-49	06-12-85	5,500	-91.0	-11.95	Shallow
26S/40E-22H2	2,227.03	77	75-77	06-12-85	12,500	-93.0	-12.30	Shallow
26S/40E-22H3	2,226.23	97	95-97	06-12-85	5,100	-97.0	-12.80	Shallow
26S/40E-22P2	2,262.8	75	73-75	06-10-85	1,800	-97.0	-12.50	Shallow
26S/40E-22P3	2,260	415	400-415	01-09-86	1,940	-102.0	-12.90	Deep

Table 1. Stable-isotope ratios for water samples from selected wells in Indian Wells Valley--*Continued*

Well number	Altitude of land surface	Depth of well	Perforated interval	Sample date	Specific conductance (µS/cm)	Delta deuterium (permil)	Delta oxygen-18 (permil)	Monitored aquifer
			China Lake Area--*Continued*					
26S/40E-22P4	2,260	215	200-215	01-09-86	2,210	-96.0	-12.40	Shallow
26S/40E-23D1	2,223	400	385-400	01-09-86	2,780	-101.0	-13.40	Deep
26S/40E-23D2	2,223	185	170-185	01-09-86	6,800	-101.0	-12.80	Shallow
26S/40E-23G1	2,215.04	57	55-57	06-11-85	9,750	-93.0	-12.05	Shallow
			Southeastern Area					
26S/40E-28J1	2,288.9	--	--	01-23-86	830	-96.0	-12.50	Deep
26S/40E-32F3	2,320	720	520-700	01-22-86	880	-105.0	-13.90	Deep
26S/40E-32K1	2,330	620	230-310; 340-380; 470-500; 520-600	01-22-86	675	-103.0	-13.75	Deep
26S/40E-36A1	2,247.2	260	80-90; 107-127; 187-195; 240-260	06-17-85	2,000	-94.5	-12.35	Deep
27S/40E-2J1	2,300	200	--	01-24-86	1,890	-99.0	-12.90	Deep
27S/40E-4B2	2,998	288	128-278	02-27-86	1,260	-97.0	-12.80	Deep
27S/40E-5D1	2,375	555	251-556	02-27-86	590	-104.0	-13.50	Deep
			Intermediate Area					
26S/39E-24M1	2,366.46	800	220-405; 450-620; 730-800	09-18-85	332	-93.0	--	Deep
26S/39E-25E1	2,345	387	179-260; 268-284; 291-384	02-27-86	405	-96.0	-12.75	Deep
26S/40E-30E2	2,345	378	205-378	02-27-86	383	-95.5	-12.70	Deep
26S/40E-30K1	2,340	800	250-800	01-22-86	364	-102.0	-8.20	Deep
26S/40E-30K2	2,340	802	220-470; 600-760	05-14-87	375	-99.5	-13.25	Deep
27S/40E-6D1	2,400	720	580-700	01-22-86	367	-108.0	-13.80	Deep

Data are insufficient from recharge areas in the eastern Sierra Nevada directly west of Indian Wells Valley to place as much reliability on isotopic composition as was done for the Coso and Argus Ranges; however, there is some evidence (see table 2) that the range is greater in the Sierra Nevada. For example, Sacatar Spring has a δD of -110 permil. (The high value of δD= -78 permil in Little Lake likely includes the effect of evaporation.) Despite the uncertainty in isotope ratios from the Sierra Nevada, it is recognized that colder and (or) wetter climates, reflecting either higher altitudes or an earlier pluvial period, produce lighter ratios; there is some evidence that this can explain, in part, some of the isotopic variations in ground water from Indian Wells Valley.

Given the range in isotopic composition of water in areas supplying recharge, it is not surprising that nearly all ground-water samples show δD values more negative than -90 permil and that most are more negative than -95 permil. A notable exception is shallow ground water in the discharge area of the lakebeds of present and ancestral China Lake where evapotranspiration results in δD values as high as -80 permil.

As used in subsequent sections of this report, the term "current recharge" refers to water that has been recharged between the present and some unspecified time in the recent past. It is intended to imply water of some effective average composition that reflects variations that occur over time scales ranging from annual to perhaps several hundred years (as distinct from climatic patterns that persist for thousands of years). The term "historical recharge" refers to older, perhaps late Pleistocene, ground water, whose presence in the basin is indicated by results of this study and a study by Smith and others (1992) in southeastern California.

Table 2. Stable-isotope ratios for samples from springs and surface-water sites in the Sierra Nevada and the Coso and Argus Ranges

[Designation of sites is shown on plate 1, except sites with italic letter designation are outside plate 1 area. Altitude of land surface is given in feet above sea level. µS/cm, microsiemens per centimeter at 25°C; permil, parts per thousand deviation from V-SMOW (Vienna Standard Mean Ocean Water); gal/min, gallon per minute; --, no data]

Designation	Spring or surface-water site	Altitude of land surface	Sample date	Specific conductance (µS/cm)	Delta deuterium (permil)	Delta oxygen-18 (permil)	Discharge rate (gal/min)
			Sierra Nevada[1]				
A	Sand Canyon Creek	3,120	01-24-86	--	-93	--	10
B	Little Lake	3,180	05-02-86	--	-78	--	--
C	Sand Canyon Spring	3,800	05-02-86	--	-92	--	.12
D	Sacatar Spring	3,840	05-02-86	--	-110	--	10-15
E	Chimney Creek	6,250	05-02-86	--	-105	--	5-10
F	Ninemile Canyon Creek	2,950	05-02-86	--	-96	--	5-10
G	Indian Wells Canyon Creek	3,815	05-02-86	--	-96	--	5
			Coso and Argus Ranges				
H	Coso Village Spring	5,820	07-13-89	615	-85.0	-9.25	1.4
I	Coles Spring	6,260	07-13-89	1,100	-96.0	-12.85	2.5
J	Crystal Spring	6,400	07-13-89	440	-102.0	-13.85	1.6
K	Tennessee Spring	6,200	06-07-89	410	-100.0	-13.60	3.5
L	New House Spring	6,280	06-07-89	360	-99.0	-13.50	2
M	Wildrose Spring	5,170	06-06-89	630	-93.0	-12.55	.12
N	Birchum Springs	5,680	07-13-89	450	-89.5	-12.25	.44
P	Mariposa Spring	6,320	07-13-89	1,180	-93.5	-13.10	2

[1] Isotope analyses done by G.I. Smith, U.S. Geological Survey, Menlo Park, California.

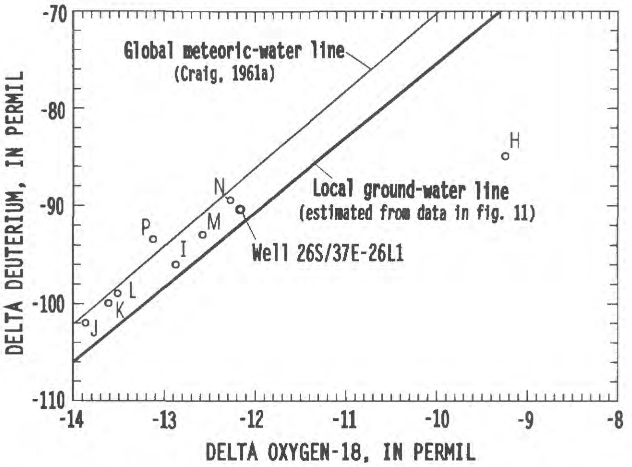


Figure 4. Relation between stable-isotope ratios of hydrogen and oxygen in samples from springs in the Coso and Argus Ranges. Data are given in table 2, and location of springs is shown on plate 1. (Well 26S/37E-26L1, in the Sierra Nevada in the western part of the study area, is included for comparison.)

GROUND-WATER FLOW DIRECTIONS AND TRAVELTIMES

A potential exists for saline ground water to migrate into the main water-producing zone (deep aquifer) in the intermediate and southeastern areas of Indian Wells Valley. The delineation of "areas" (pl. 1) in Indian Wells Valley is discussed in a later section, "Ground-Water Quality," which also contains an extensive discussion of the effect that pumping has on water quality by altering ground-water flow. Long-term pumping from the deep aquifer in the intermediate area and at Ridgecrest has caused a decline of hydraulic heads in these areas. Although saline water apparently has not migrated into the main water-producing zone on a regional scale, increased DSC in several wells near centers of heavy pumping has caused concern among water managers and planners. The problem posed by migrating saline water can be expected to intensify with Indian Wells Valley's rapid growth in population and accompanying demand for water.

The main processes affecting the fate and transport of solutes in ground water are: (1) advection by the ground-water flow system, (2) hydrodynamic dispersion (including diffusion and mechanical dispersion), and (3) geochemical reactions. Application of an advective-dispersive solute-transport simulation model was attempted, but the application was considered inappropriate because of (1) insufficient data on initial concentrations, on areas containing high-DSC water, and on the three-dimensional aspects of hydraulic properties; and (2) computational limitations (small cell sizes and hence large computation times required to avoid numerical dispersion and instabilities). Therefore, this report is limited to a simple analysis of three-dimensional advective transport of nonreactive constituents in the Indian Wells Valley ground-water system. This analysis does not consider chemical reactions, and assumes, as is generally valid for high permeability systems such as the deep aquifer in Indian Wells Valley, that the influence of advection will be much more important than that of dispersion. The results from this analysis represent average ground-water traveltimes and pathlines for nonreactive dissolved constituents. In addition to illustrating the effect that pumping has on ground-water flow and quality, results from this analysis also demonstrate the great antiquity of ground-water that is far removed from where it was recharged.

The computer model used in this study to simulate advective transport is MODPATH (Pollock, 1988, 1989). MODPATH employs a particle-tracking technique to compute pathlines and traveltimes based on the results of MODFLOW, the USGS modular three-dimensional finite-difference flow model (McDonald and Harbaugh, 1984). Results from the previous ground-water flow model of Indian Wells Valley developed by Berenbrock and Martin (1991) are used in this application of MODPATH. A complete description of MODPATH's theoretical development, solution techniques, and limitations is available in two reports by Pollock (1988, 1989).

To demonstrate the effect that pumping has on ground-water flow and traveltimes, particle tracking was simulated using both predevelopment steady-state (1920) as well as current (1988) ground-water conditions. The previous ground-water flow model developed by Berenbrock and Martin (1991) stopped simulations at the end of 1985. For this study, the distribution of ground-water pumpage, mountain-front recharge, and wastewater recharge was assumed to remain constant during 1985-88 because data to update these values (for 1986-88) were not available at the time. Subsequent simulations using the three-dimensional finite-difference flow model show this assumption to be reasonable: The model yields a decline in storage of 16,750 acre-ft during 1988 in comparison with 17,130 acre-ft during 1985 (Berenbrock and Martin, 1991, p. 39, table 5). The net result is that total storage depletion since 1920 increased from about 470,000 acre-ft in 1985 to about 520,000 acre-ft in 1988 (fig. 5). Accordingly, MODPATH simulations for 1985 and 1988 should yield similar results.

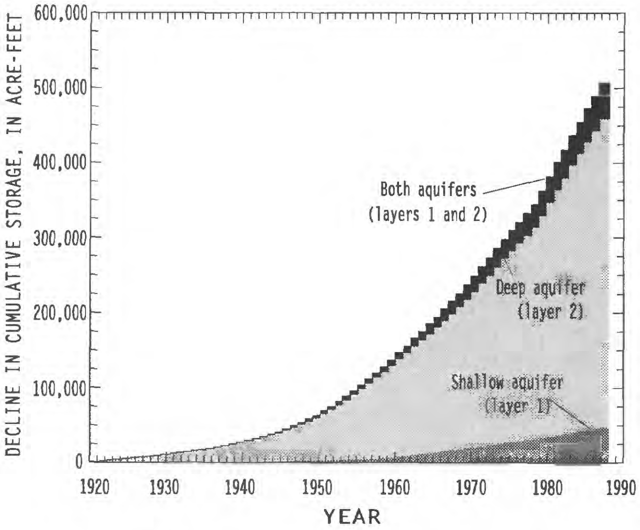


Figure 5. Simulated annual decline in ground-water storage in Indian Wells Valley, 1928-88, using the modular three-dimensional finite-difference flow model, MODFLOW.

Several particle-tracking simulations were completed to (1) assess the potential for saline ground water to migrate into the intermediate and southeastern areas of the valley from China, Mirror, and Satellite Lakes, and (2) determine the effect of ground-water pumping on flow direction and traveltimes from the Little Dixie Wash. Particles were tracked to discharge areas (represented by evapotranspiration or pumpage cells) to show general ground-water flow direction and distances traveled.

Pollock (1989, p. 19-21) lists several limitations of MODPATH. One important limitation is that MODPATH cannot simulate transient-flow systems. Therefore, the application herein considers the 1988 ground-water conditions in Indian Wells Valley, which are the result of transient flow, as steady-state conditions in order to use MODPATH. The error inherent in following this procedure cannot be quantified; however, the fact that calculated velocities differ little between the 1920 and 1988 simulations—except where pumpage causes a change in direction of ground-water flow, and consequently a change in the type of material through which the flow occurs—suggests that the error is small. This use of MODPATH assumes that hydraulic heads and the magnitude and direction of the flow velocity at any point in the model are constant throughout the simulation; in reality, however, hydraulic heads and flow velocities are known to be increasing or decreasing in response to ground-water pumping and (or) artificial recharge. Therefore, it must be emphasized that traveltimes calculated from these simulations should not be interpreted as quantitative predictions of the timing of possible degradation. The several limitations of MODPATH that are discussed by Pollock (1989) preclude such specific projections in Indian Wells Valley. These particle-tracking simulations provide only a general indication of relative rates and directions of solute movement.

Because estimates of porosity in the unconsolidated deposits are not available for Indian Wells Valley, porosities used in this model were based on values for various materials published by Bouwer (1978, p. 22) and Mercer and others (1982, p. 15). Porosities of 50 percent probably are representative of playa and lacustrine deposits and are assigned to cells in and around China, Airport, Mirror, and Satellite Lakes. Porosities of 30 percent probably are representative of the alluvium and are assigned to all other cells in the grid.

BETWEEN CHINA LAKE AND INTERMEDIATE AREA

Two simulations were completed to illustrate the change caused by pumping in ground-water movement between China Lake and the intermediate area. For the first simulation, particles were placed on the north face of eight model cells (in row 43) in layer 2 southwest of China Lake (figs. 6*A*,*B*) to demonstrate the effect that pumping in the intermediate and Ridgecrest areas has on ground-water flow and on traveltimes from the China Lake area. Particles were placed in cells that coincide with the southern-most extent of fine-grained deposits in the China Lake area (Berenbrock and Martin, 1991, fig. 3); these deposits probably contain saline water, as indicated in a later section of this report (see "China Lake Area").

Ground-water movement under 1920 (predevelopment) conditions, as shown in figure 6*A*, is to the north toward the area of evapotranspiration near China Lake. Particle positions are shown at their initial starting locations in the eight model cells (fig. 6*A*) and for each successive 1,000 years of simulation until all particles reach the evapotranspiration area (discharge point). Particles first reach the evapotranspiration area after 6,000 years, and all particles have reached the evapotranspiration area after 13,000 years. Dividing these traveltimes by horizontal distance yields a range for model-derived velocities (which represent both horizontal and vertical movement of ground water) of 0.7 to 2.5 ft/yr. Berenbrock and Martin (1991, Supplemental Data D1-D3, and oral commun., 1991) estimated that annual ground-water pumpage in the intermediate and Ridgecrest areas exceeded 5,100 acre-ft during 1963-88 and averaged about 6,500 acre-ft. In the simulation using 1988 (current) conditions, direction of the ground-water gradient has been reversed and flow is southward toward pumped wells in the intermediate and Ridgecrest areas (fig. 6*B*). The range of model-derived velocities for this simulation is 2.6 to 7.8 ft/yr, and particles from the former discharge area reach pumped wells in the intermediate and Ridgecrest areas in 1,000 to 5,000 years.

For the second simulation of ground-water movement between China Lake and the intermediate area, a generalized cross section *B-B′* along model column 26 was used to show the interaction of ground water from both model layers and the projected flow paths along the model cross section (fig. 7). In these simulations, one particle was placed in each of 16 cells (8 in layer 1 and 8 in layer 2) along the section. The particles were centered vertically on the north face of the cells. The simulation for 1920 conditions (fig. 7*A*) shows ground water moving northward and discharging at land surface. Simulated pathlines are much shallower for particles placed in rows closest to China Lake because they are nearer the area of discharge. In the simulation of 1988 conditions, not all the ground-water flow paths from layer 2 extend toward the discharge area to the north (fig. 7*B*). Extensive pumping since 1920, especially in the intermediate area, has caused a flow reversal in the main water-bearing zone (layer 2) (see fig. 3); hence, ground-water movement is to the south into model cells near the southeastern area (fig. 7*B*). The comparatively shorter traveltimes and higher ground-water velocities of particles entering cells closest to the southeastern area (rows 43 through 47) are caused by increased hydraulic gradients owing to pumping and by higher transmissivities toward the intermediate and southeastern areas (Berenbrock and Martin, 1991; p. 27-31). Particles located in layer 1 and near the southeastern area move downward toward layer 2; for particles in cells near the China Lake area (rows 41 through 38), flow direction is similar (northward) to that shown in the simulation of 1920 conditions.

Results of these particle-tracking simulations illustrate how pumping in the valley, especially in the intermediate area, has reversed the historical movement of ground water between China Lake and the intermediate area. They also show how a relatively nonreactive constituent—such as dissolved solids, whose concentration is high in ground water beneath China Lake—will travel very slowly to the intermediate area. Perhaps these slow traveltimes are the reason that DSC in wells in the intermediate area has not changed significantly during 1950-88. Note, however, that this analysis simulates only the regional movement of ground water. The estimated traveltimes do not represent the potential movement of solutes from local sources such as small lenses of lacustrine deposits that might be nearby.

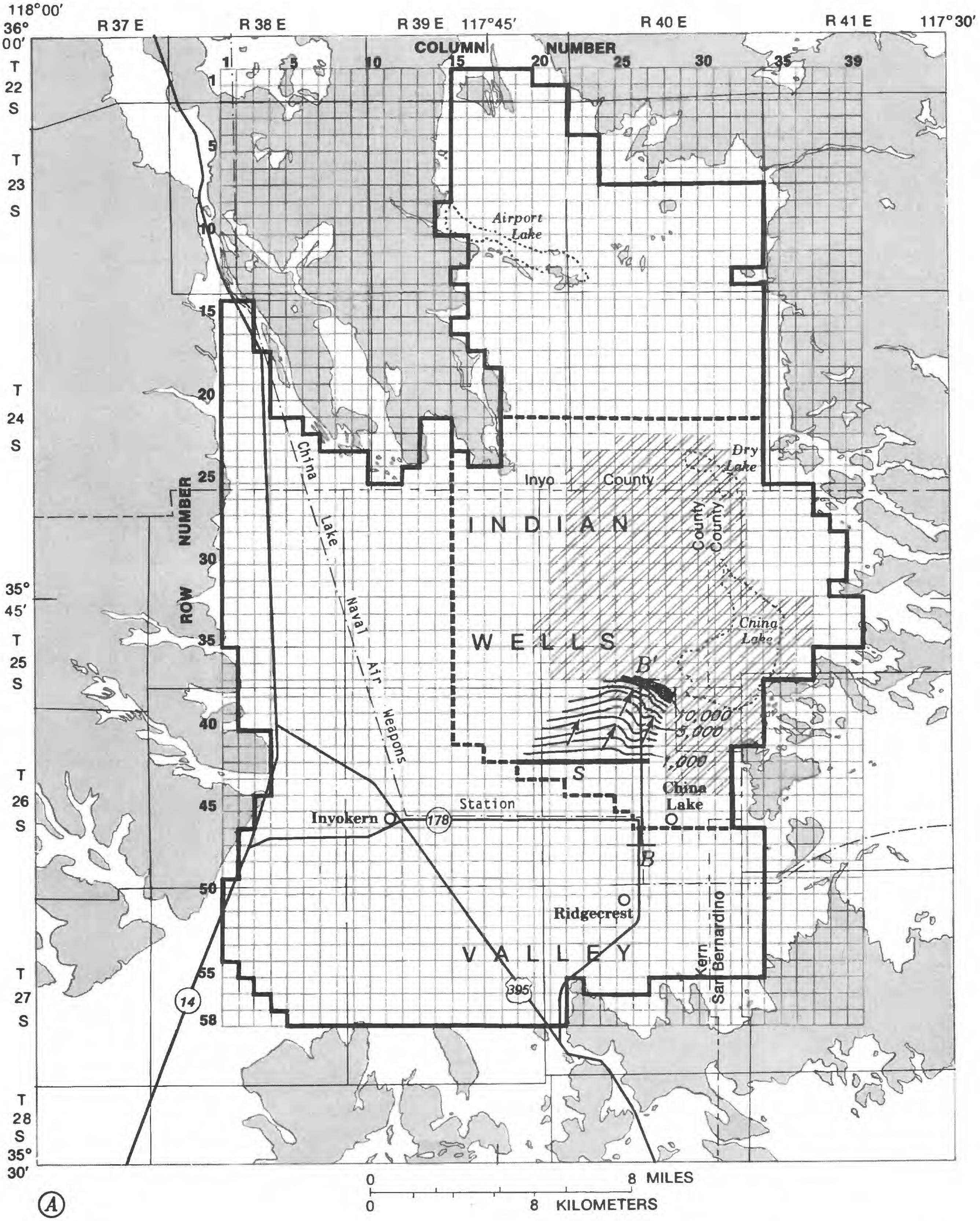


Figure 6. Model-simulated movement of ground water and maximum extent of travel of particles that initially were placed in the China Lake area. *A*, 1920 ground-water conditions. *B*, 1988 ground-water conditions.

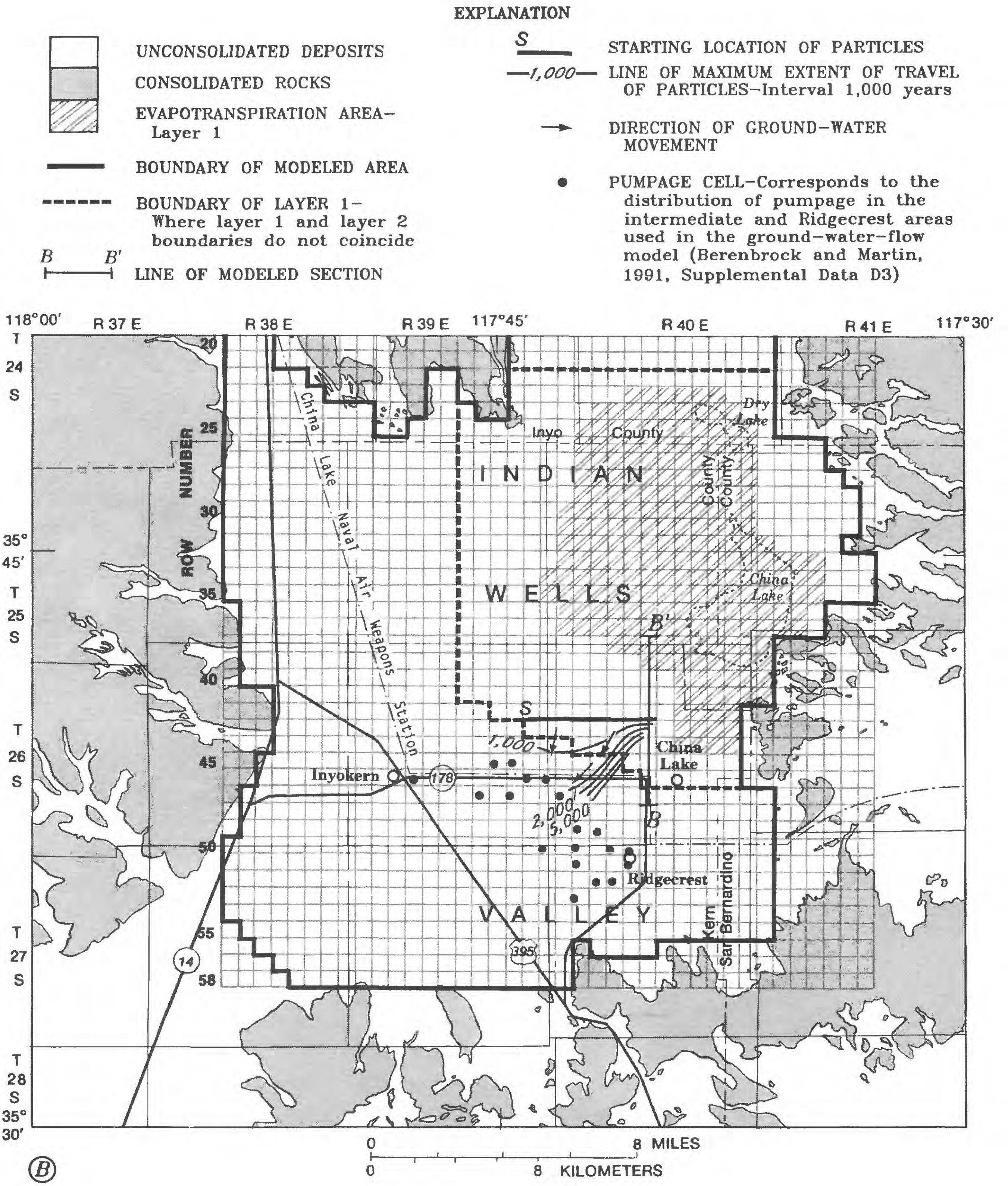


Figure 6. *Continued.*

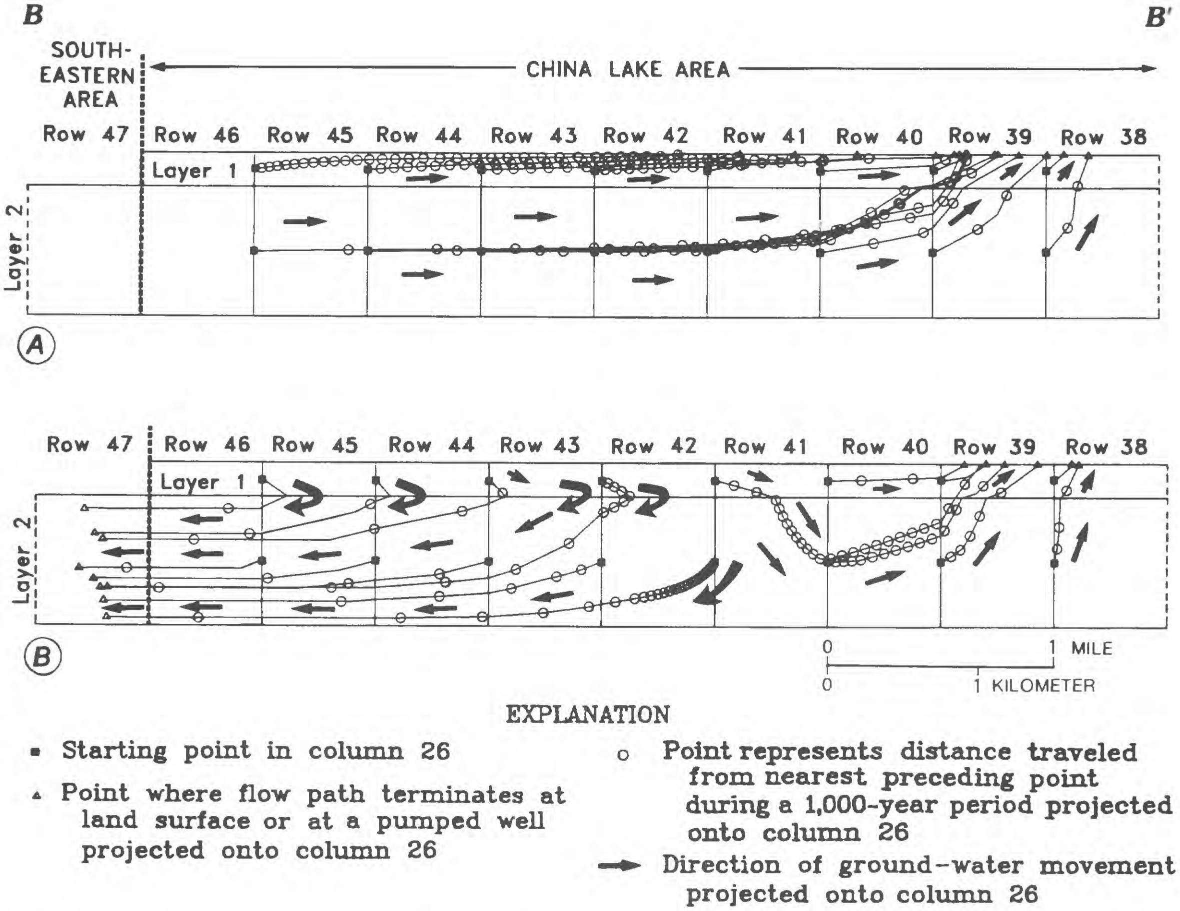


Figure 7. Model-simulated movement of ground water and the flow path of selected particles projected along a section matching model-column 26. ***A***, 1920 (steady-state) conditions. ***B***, 1988 ground-water conditions. Line of section is shown in figure 6.

MIRROR AND SATELLITE LAKES TO RIDGECREST

Model simulations also were used to demonstrate the effects that pumping at Ridgecrest has on ground-water flow and traveltimes from Mirror and Satellite Lakes. As discussed in a later section of this report (see "Southeastern Area" section), significant increases in DSC have occurred since 1972 in several wells at Ridgecrest. The increase is believed to be the result of induced migration of saline water from deposits beneath Mirror and (or) Satellite Lakes.

For the first simulation, particles were placed on the north face of cell (48,29) for the simulation using 1920 conditions and on the west face of three cells for the simulation using 1988 conditions. These model cells represent the northern and western edges of the deposits beneath Mirror and Satellite Lakes. The direction of particle movement for the simulation using 1920 conditions was similar to that of the China Lake area (see previous section) in that the particles move northward in both model layers (fig. 8*A*) until eventually discharged by evapotranspiration. The traveltime of particles to reach the evapotranspiration area ranged from 12,000 to more than 20,000 years. The low hydraulic conductivity and hydraulic gradient encountered northward in this part of the basin are the major factors causing low ground-water velocity (long traveltimes). For the simulation using 1988 conditions, particles move west toward Ridgecrest and the intermediate area (fig. 8*B*) where most pumping in the valley occurs. All particles reach the pumped wells by 4,000 years, and the range of model-derived velocities is 6.5 to 8.7 ft/yr.

In a second simulation, particle movement was tracked along model cross section *C-C'* between the Mirror and Satellite Lakes area and Ridgecrest (fig. 9) to estimate traveltime under 1988 conditions. For this simulation, particles were placed on the west face in only one cell (48,29) in layer 2. This cell, located at *C'*, represents the westernmost edge of deposits containing high-DSC ground water beneath Mirror and Satellite Lakes. As shown in figure 9, time required for particles to reach the large-capacity pumped wells 26S/40E-28J1 and 26S/40E-34N1 ranges from 1,300 to 1,400 years.

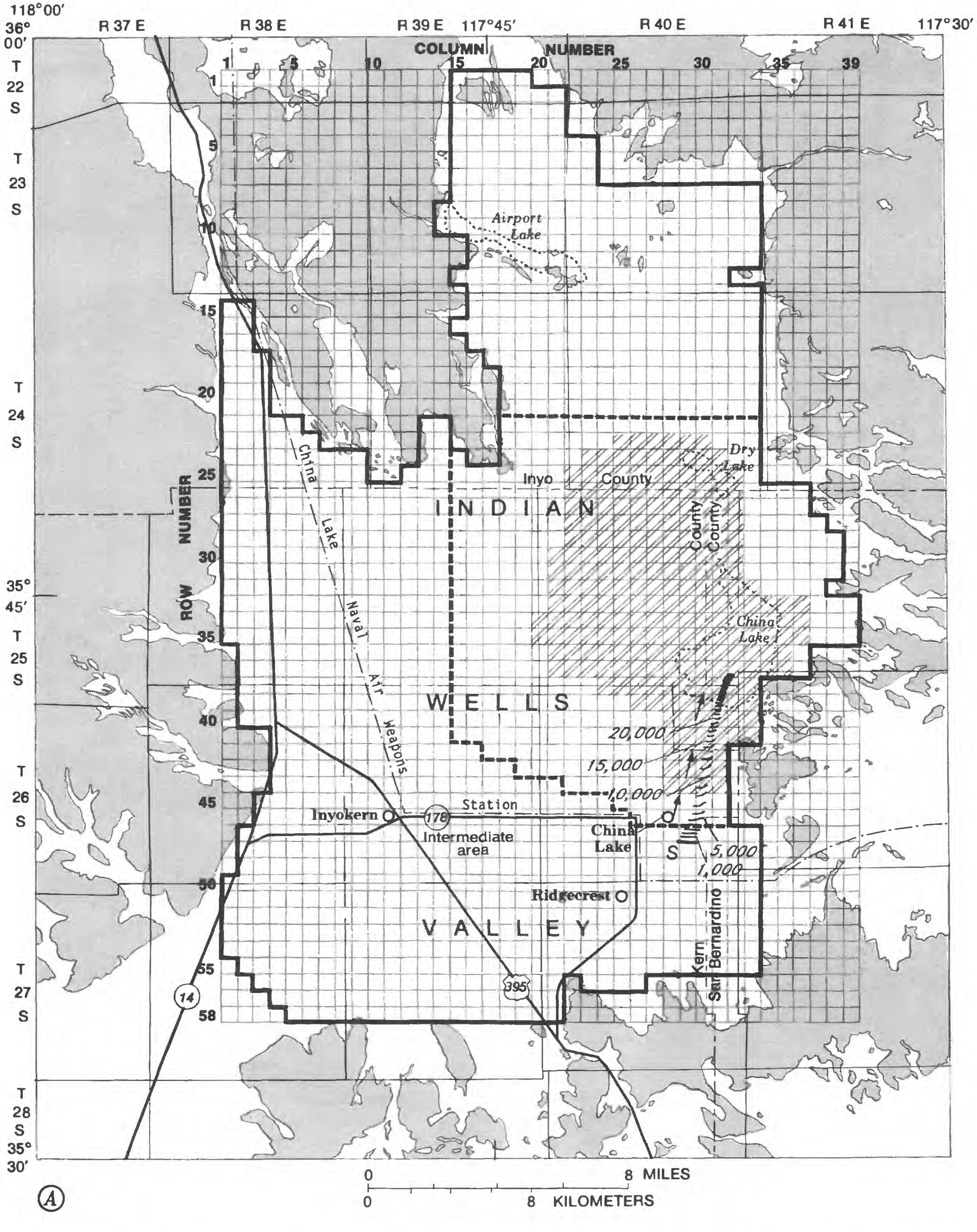


Figure 8. Model-simulated movement of ground water and maximum extent of travel of particles that initially were placed beneath Mirror and Satellite Lakes. ***A***, 1920 ground-water conditions. ***B***, 1988 ground-water conditions.

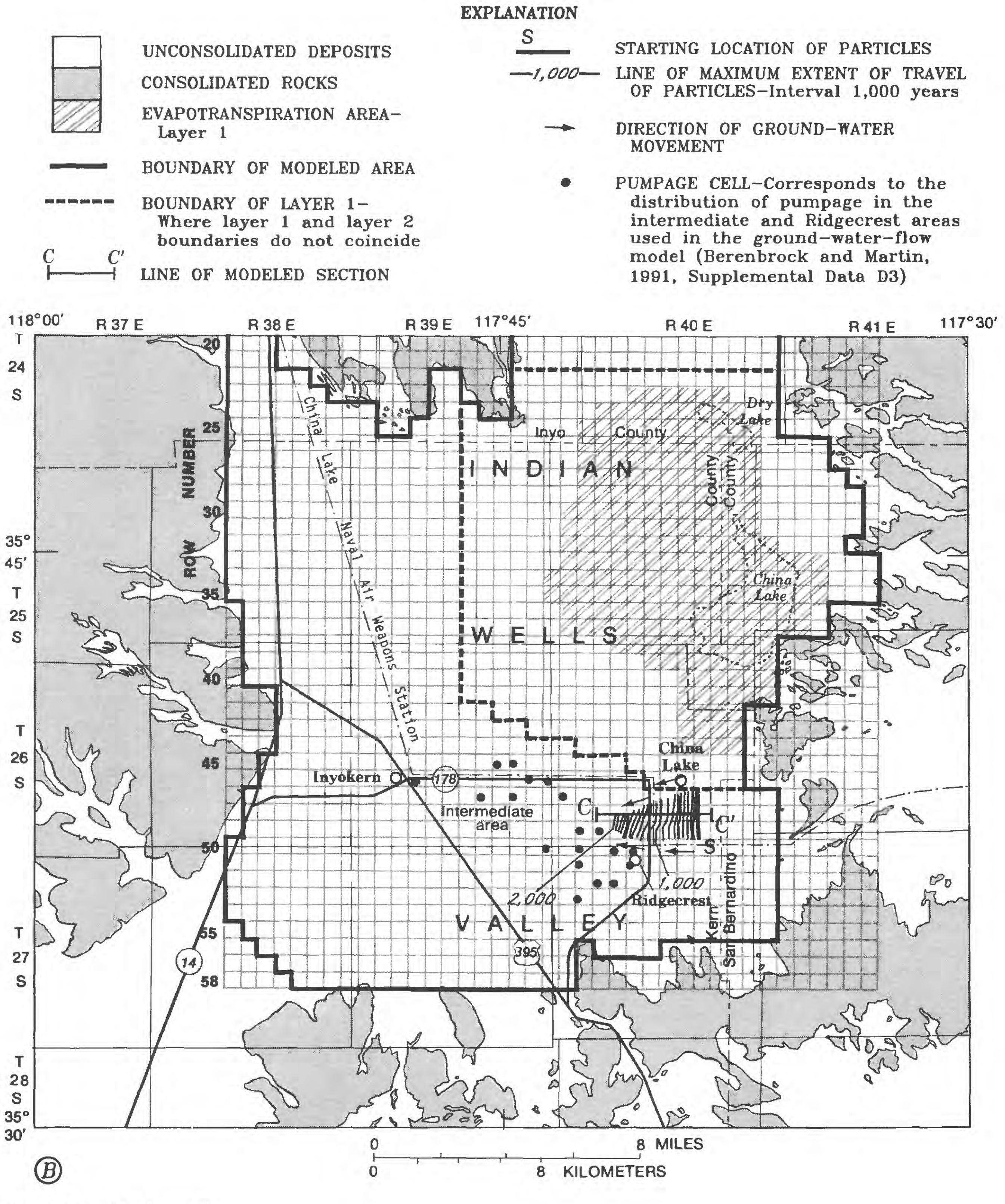
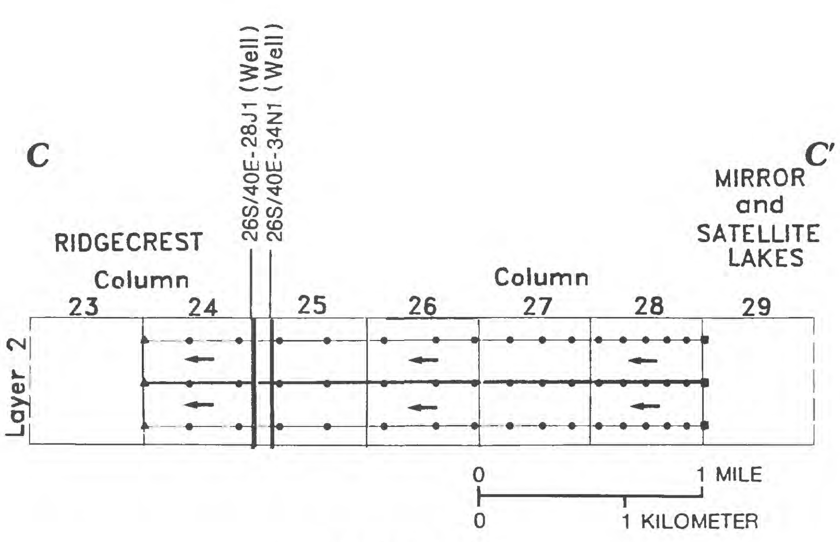


Figure 8. *Continued.*



EXPLANATION

■ Starting point in row 48

▲ Point where flow path terminates at a pumped well projected onto row 48

• Point represents distance traveled from nearest preceding point during a 1,000-year period projected onto row 48

← Direction of ground–water movement projected onto row 48

Figure 9. Model-simulated movement of ground water and the flow path of selected particles projected along a section matching model-row 48, for 1988 ground-water conditions. Line of section is shown in figure 8.

The simulated traveltime is much longer than that indicated by observed increases in DSC at the above wells since 1972, thus indicating that solute movement through the lower aquifer cannot be approximated with a single model layer. Interbedding of the more permeable alluvium with fine-grained lacustrine deposits in the central and eastern parts of the valley has produced considerable vertical variation in aquifer properties, ground-water velocities, and water quality. Because MODPATH and the modular (MODFLOW) flow model on which MODPATH is based vertically average their parameters throughout a cell's thickness, these models cannot simulate depth-related differences within a layer. Further complicating any meaningful comparison of times for model-calculated and recently observed (since 1972) increases in DSC in several wells at Ridgecrest is the possibility that observed increases may well be caused by local sources—such as movement of ground water from nearby layers of fine-grained deposits that contain high-DSC water—rather than by regional movement of ground water from Mirror and Satellite Lakes. Also, ground-water velocities through any coarse-grained strata that may be present could be much faster than the model indicates for regional flow. Additional drilling and sampling, especially in this part of the valley, would be needed to characterize the aquifers more precisely and to discretize the aquifer into a number of layers in a model.

FROM LITTLE DIXIE WASH

Model simulations were used to demonstrate the effect of ground-water pumping on flow direction and traveltime from the Little Dixie Wash, which contributes the largest quantity of low-DSC (less than 500 mg/L) ground water to the heavily pumped part of the intermediate area. For these simulations, 100 particles were placed throughout three model cells in Little Dixie Wash at the southwestern boundary of the model. In the simulation based on 1920 conditions (fig. 10*A*), ground water moves to the northeast toward the evapotranspiration area at China Lake. Lines of maximum extent of traveltimes of particles are shown at the initial starting locations and for each 1,000 years until the particles discharge. Some particles reach the evapotranspiration area in 3,000 years, and by 5,000 years most have reached the evapotranspiration area. The model-derived particle velocities ranged from 4 to 22 ft/yr.

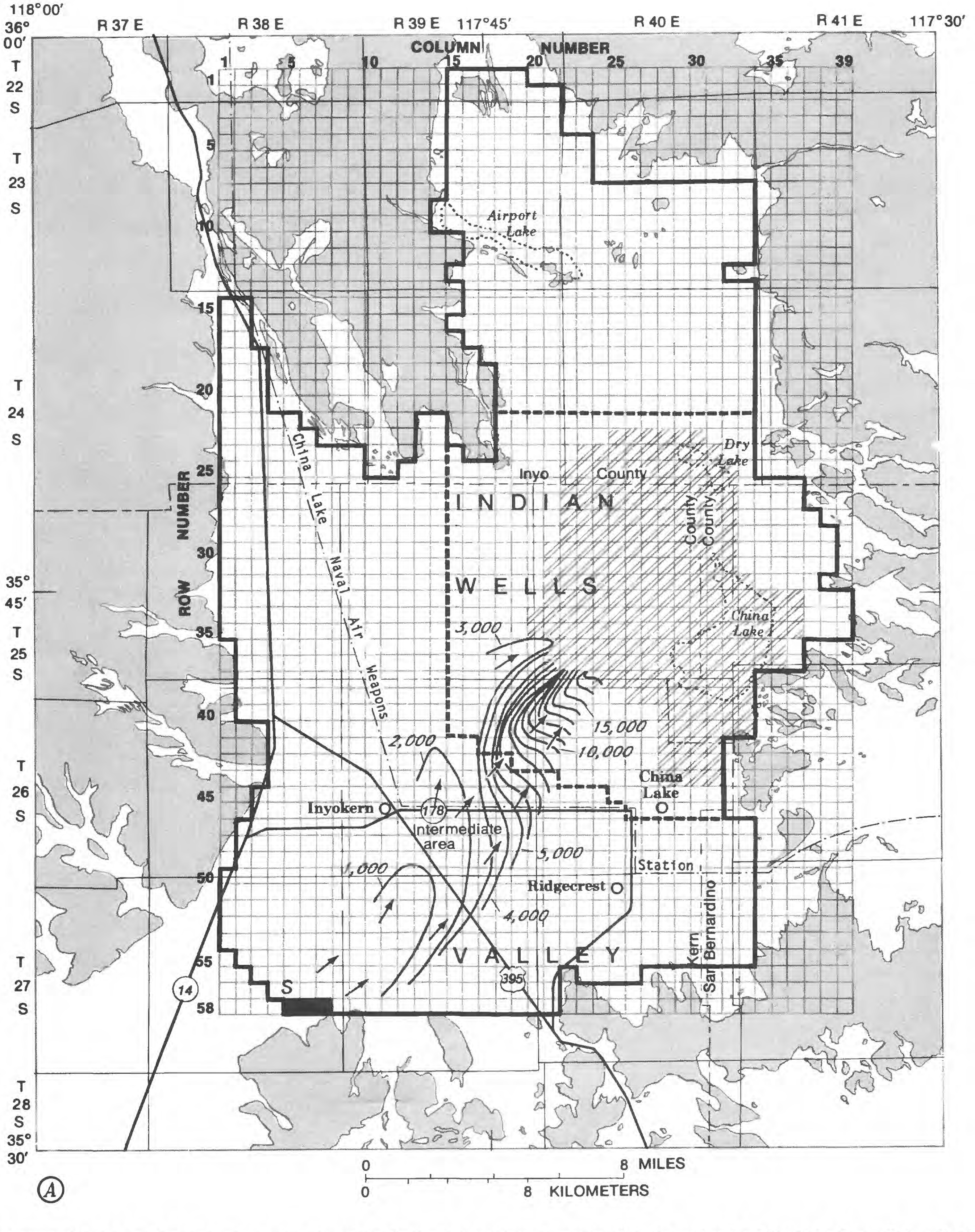


Figure 10. Model-simulated movement of ground water and maximum extent of travel of particles that initially were placed in the Little Dixie Wash area. *A*, 1920 ground-water conditions. *B*, 1988 ground-water conditions.

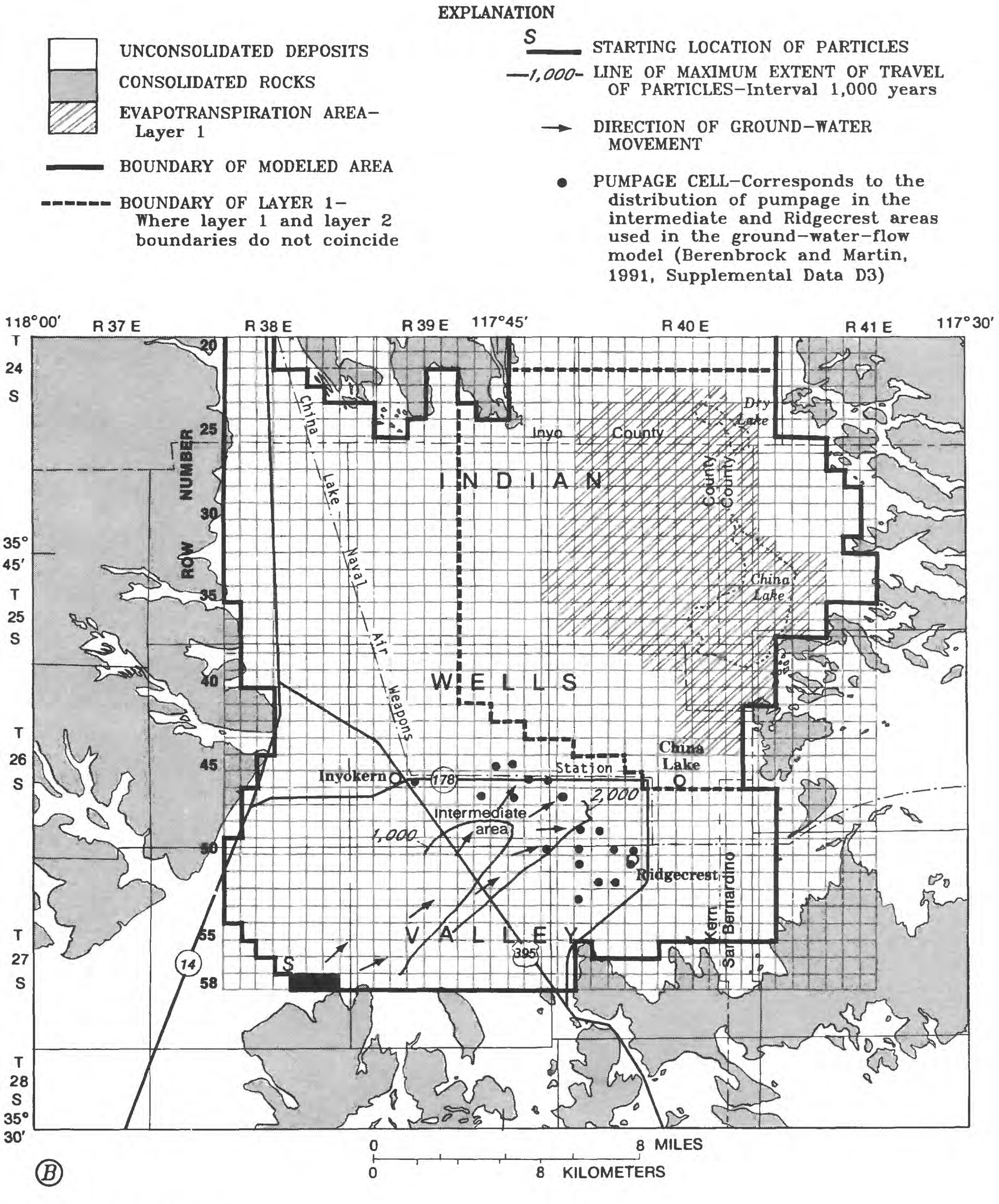


Figure 10. *Continued.*

In the simulation based on 1988 conditions, particles are deflected eastward, toward pumped wells in the intermediate area and at Ridgecrest (fig. 10*B*, fig. 3). Berenbrock and Martin (1991, p. 18) showed that ground-water development since the 1920's has caused this modification to the direction of ground-water movement in the deep aquifer. Model-derived velocities ranged from 15 to 39 ft/yr (approximately double those for 1920 conditions). All particles in this simulation reach the pumped wells within 3,000 years; most discharge within 1,000 to 2,000 years. This simulation illustrates how pumping has modified the direction of ground-water flow and traveltimes in this part of the basin, diverting low-DSC water from the west toward pumped wells at Ridgecrest and the intermediate area and away from the China Lake area.

For all simulations in Indian Wells Valley, the algorithm was operated to track particles forward along their pathlines. The algorithm could have been operated "in reverse" to track particles backward (Shafer, 1987; Pollock, 1989). Resulting ground-water traveltimes would be similar to those (of several thousand years between parts of the valley) obtained from forward simulations described herein, thus lending additional support to the isotopic evidence for the existence of ground water in the China Lake area that was recharged during an earlier pluvial period (see "Pleistocene Ground Water" section).

GROUND-WATER QUALITY

The quality of ground water in Indian Wells Valley varies both areally and vertically in the water-bearing deposits. Generally, sodium or calcium and bicarbonate are the predominant ions in most parts of the study area, especially where DSC is less than 500 mg/L, which is the recommended limit for drinking water (U.S. Environmental Protection Agency, 1979). In areas where higher DSC occurs, especially in the shallow aquifer, and where fine-grained lacustrine deposits are present in the deep aquifer, sodium and bicarbonate and chloride or sulfate commonly are predominant ions. As ground water moves from recharge areas along the mountain fronts through the alluvial deposits and eventually into the lacustrine deposits in the central and eastern parts of the valley, it acquires the chemical characteristics of high-DSC waters, through dissolution of evaporites in the ancestral China Lake deposits and (or) through evaporative concentration in discharge areas of the low-altitude lakebed. Similar processes probably have occurred on a smaller scale in the other (smaller) dry lakebeds of Indian Wells Valley such as those of Airport, Dry, Mirror, and Satellite Lakes.

In order to facilitate the discussion, in subsequent sections of this report, of water-quality characteristics in Indian Wells Valley, the valley has been divided into seven areas (see pl. 1): northeastern, northwestern, western, southwestern, China Lake, southeastern, and intermediate. This division into areas is somewhat artificial in that it is not based on differences in lithology or on any geohydrologic boundaries, but it does reflect common usage by local residents and by investigators in previous studies. For example, the geographic boundary of the China Lake area matches the boundary of layer 1 (the shallow aquifer) used in a model by Berenbrock and Martin (1991).

When viewed as a hydrologic unit, Indian Wells Valley consists of the mountain ranges and sloping coarse-grained alluvial deposits that grade into and interfinger with the fine-grained lacustrine deposits of the ancestral China Lake and the playa. It is the interfingering of these deposits, particularly at the margins of the ancestral China Lake, that makes interpretation of water-quality data from wells perforated over a wide depth interval especially difficult. Further compounding this difficulty is the fact that existing wells represent a wide range of physical conditions and uses, from well-maintained and continuously pumped supply wells to seldom-used observation wells. Finally, the absence of wells in appropriate areas, the combining of data collected at different times and by different agencies, and the paucity of long-term monitoring data from suitable wells impose additional problems. Accordingly, some generalizations are made in the subsequent interpretations that either are based on scant information or to which data from individual wells may not conform. Such generalizations are necessary to provide a conceptual framework that can explain variations in chemical concentrations and isotope ratios.

NORTHEASTERN AREA

No wells in the northeastern part of Indian Wells Valley were sampled during 1987-88 and few wells exist there (pl. 1). Berenbrock and Martin (1991, p. 55) stated that pumping probably has not affected ground-water conditions in this part of the valley; hence, the most recent data available from selected wells are used to assess water quality. High DSC may be attributable to fine-grained lacustrine (ancient lakebed) deposits in the southern part of this area. Although historical data indicate that DSC exceeds 1,000 mg/L and that sodium and chloride are the predominant ions in the southern part of this area, it is speculated that ground water upgradient and toward the north might have a much lower DSC and a corresponding shift toward calcium and bicarbonate predominance.

Some information on soluble salts and isotope ratios was obtained from eight springs in the mountains north of this area (table 2 and pl. 1). Three of the springs are 15 to 20 mi north of Airport Lake in a geothermal area (outside pl. 1 area). The

Coso Village Spring's isotopic datum is believed to be anomalous because it plots far beneath the global meteoric-water line (refer to discussion in previous section). Despite evidence of extensive evaporation from the pool fed by this spring, specific conductance in the pool (615 μS/cm) is not especially high. However, isotopic data from the two other springs, Coles Spring and Mariposa Spring, plot near the global meteoric-water line, even though these springs are 2 to 3 times higher in DSC (on the basis of specific-conductance measurements) than are other springs in the Coso and Argus Ranges (see table 2). The high DSC in Coles and Mariposa Springs might result from increased dissolution of minerals at higher temperatures while the water itself retains the isotopic signature of recharge (δD about -95 permil). There is a need for wells, particularly near and north of the dry lakebed of Airport Lake, to answer questions on whether (and if so, where) high-DSC ground water originates in this part of the valley.

NORTHWESTERN AREA

Four wells in the northwestern area of Indian Wells Valley were sampled during 1987-88 (pl. 1). Because there is little pumping in this part of the valley, ground-water conditions are unlikely to have changed. Therefore, to increase the number of wells on which the areal assessment of water quality is based, the most recent historical data from selected wells are used, along with the 1987-88 data. The DSC ranges from about 300 mg/L in the western and southern part of this area to almost 1,000 mg/L in the eastern part and to more than 1,000 mg/L in the far north (see pl. 1). Low DSC is attributed to proximity to mountain-front recharge from streams emerging from the Sierra Nevada (Sand, Boulder, County Line, and Noname Canyons). Higher DSC to the east is attributed to the fine-grained deposits at the maximum extension of the ancient China Lake shoreline. These areal differences are especially evident for chloride concentration, which is several times higher in wells 25S/39E-4R1 and 25S/39E-9J1 to the east than in wells 25S/38E-11L1 and 25S/38E-13J1 to the west (Supplemental Data B and pl. 1).

Highest DSC occurs in the extreme northern part of this area, where quality is influenced by underflow of isotopically light (δD less [more negative] than -100 permil) saline water from Rose Valley to the north (Gleason and others, 1992). For example, well 24S/38E-16J2 is perforated from 251 to 611 ft, and the water has a DSC of 1,320 mg/L. Despite high DSC, its isotopic composition is much lighter (δD= -95 permil) than that of water from well 24S/39E-33N1 in the eastern part of this area (pl. 1), in which DSC is only a little less at 1,100 mg/L but δD is -76 permil. (High δD of -78 permil in table 2 for Little Lake is caused by surface evaporation and is not representative of underflow from Rose Valley.) These isotopic and DSC patterns indicate that underflow from Rose Valley is not a significant contributor to the slightly saline ground water in the southern and eastern parts of the area near the perimeter of the lakebed of ancestral China Lake.

The DSC at wells 25S/39E-4R1 and 25S/39E-9J1 shows fluctuations as great as 20 percent but no discernible trend since the early 1950's (see pl. 1). These sporadic fluctuations probably can be attributed to varying proportions of ground water of differing quality entering the well at the time of sampling. Well 25S/39E-4R1 is perforated from 100 to 200 ft below land surface (Supplemental Data B) and, although no data are available for the other well, it is likely to be perforated over a similarly long interval, thus increasing the chance that perforations intercept multiple zones of differing texture and water quality. Both wells are pumped sporadically for military uses; pumping, or lack thereof, prior to sample collection could affect the proportion of water entering the well from different zones.

WESTERN AREA

Recharge of ground water in the western area of Indian Wells Valley is from stream losses along the Sierra Nevada mountain front and by ground-water flow from the southwestern area (discussed in the next section of this report). The spatial pattern of DSC in water from wells in the western area (pl. 1) is related to the presence of lacustrine or alluvial deposits, the proportion of which varies both areally and vertically. Wells in the northern part of the western area, where land-surface altitude is mostly below 2,300 ft, have DSC values near or exceeding 1,000 mg/L and δD values of approximately -90 permil. In the southern part, where altitudes are mostly above 2,400 ft, DSC generally is less than 300 mg/L, and δD is more negative than -100 permil in some wells (26S/38E-35B1 and 26S/39E-17F2). These highly negative isotope ratios are associated with great depth (well 26S/39E-17F2 is perforated at 681-881 ft below land surface) or proximity to the mountains (altitude at well 26S/38E-35B1 is 2,575 ft) and are indicative of isotopically light recharge originating either from a higher altitude or during a colder and wetter climatic period.

The chemical data from well 25S/38E-36A1 differ markedly from those of nearby wells 25S/38E-36B1, 25S/38E-25J1, and 25S/39E-31D1 and are believed to be unrepresentative of natural ground-water chemical composition in the deposits at this well. Its DSC is four times greater than that in the other three wells; and, although chloride concentration usually equals or exceeds sulfate concentration in the slightly saline ground water from lacustrine deposits near the edge of the ancestral lake, sulfate is four times more abundant (on a weight basis) than is chloride in well

25S/38E-36A1. The alkaline-earth (calcium and magnesium) concentrations also are anomalously high. This well is connected to fertilizing equipment, and fertilizer likely is the source of the very high nitrate concentration (315 mg/L as nitrate, which equals 71 mg/L as nitrogen) (Supplemental Data B).

Warner (1975, p. 33, fig. 10) concluded that ground-water DSC increases with increasing depth in Indian Wells Valley. He depicts a shallow zone with DSC less than 300 mg/L, a middle zone with DSC from 300 to 500 mg/L, and a deeper zone with DSC ranging from 500 to greater than 1,000 mg/L. He based this conclusion on the observation that many deep wells in Indian Wells Valley penetrate a section of blue clay, presumably ancient lake deposits, at greater depths; and on results of chemical analyses of water reportedly from each of the three perforated intervals in well 26S/39E-19K1. However, this generalization based on the multiple-depth, inflatable-packer test done on the well in January 1965 may be questionable. The test was done to explain the cause of an increase in DSC from 632 mg/L in 1960 to 1,250 mg/L in 1963, although it is noted here that DSC again was down to 707 mg/L in 1964 (J.A. Westphal, USGS, written commun., 1966). Results of the 1965 test, along with depth of perforated intervals open to the well during the test, are summarized in table 3. Furthermore, Warner's generalized lithologic section (Warner, 1975, fig. 2) shows alternating zones of fine-grained and coarse-grained deposits. Thus, the differences in water quality (rather than being simply a direct relation with depth) might be due primarily to differing lithologies. A more detailed discussion of this follows.

Lack of internal consistency among the three indicators of DSC is shown by data in table 3. Specifically, sum of constituents is much too low on the basis of values of residue on evaporation and specific conductance. Further inconsistencies among the data are indicated by lower values of residue on evaporation and specific conductance when the well was open to either the upper (270 to 540 ft) or the middle (590 to 625 ft) perforated intervals than when it was open to both perforations simultaneously (table 3). These inconsistencies—as well as the rather wide fluctuation in DSC between 1960 and 1964, and the dramatic drop (compare data from upper interval only) between January and June 1965, after the well was permanently sealed above the middle perforation—raise considerable uncertainty about whether a simple relation exists between DSC and depth. It seems more likely that the changes in DSC of ground water are complex and local, and that they might be related to the presence or absence of interfingering fine-grained lacustrine or coarse-grained alluvial deposits associated with repeated advance and retreat of the ancestral China Lake shore. In support of this hypothesis, the absence of any fine-grained lacustrine deposits at depth in nearby Little Dixie Wash, only a short distance east of well 26S/39E-19K1, results in low DSC (199 mg/L) at well 26S/39E-20R2, even though it is perforated from 600 to 900 ft below land surface.

Table 3. Three indicators of dissolved-solids concentration in water from historical multiple-depth inflatable-packer tests in well 26S/39E-19K1

[Data are from unpublished written communications describing the results of tests done by U.S. Navy personnel in 1965 and referred to by Warner (1975, p. 33 and fig. 10). Perforated intervals, in feet below land surface: upper, 270-540; middle, 590-625; lower, 700-790]

Month/ year	Perforated interval	Sum of constituents (mg/L)	Residue on evaporation (mg/L)	Specific conductance (µS/cm)
1/65	Middle + lower	1,150	1,660	2,370
1/65	Middle	956	1,170	1,670
1/65	Upper + middle	813	1,400	1,870
1/65	Upper	709	1,090	1,560
6/65	Upper	454	798	1,000
4/87	Upper	393	393	780

Casual inspection of water-quality and perforated-depth data in Supplemental Data B seems to imply higher DSC at shallower depths and lower DSC at greater depths. However, this association might be fortuitous because most high-DSC wells in the western area are perforated at shallower depths (less than 400 ft below land surface) and are located in fine-grained lacustrine deposits at the north end of this area; by contrast, the low-DSC wells generally are perforated at greater depths (more than 400 ft below land surface) and are located at higher altitudes in the coarse-grained alluvial deposits at the south end of this area. Deeper wells in the northern part and shallower wells in the southern part of the western area would be needed to establish whether a simple depth-dependent water-quality pattern exists.

There are some large well-to-well differences in major- and minor-ion concentrations that might imply the existence of local variations in evaporite composition within the western area. (A more complete discussion of the causes of variations in major-ion composition is given in a subsequent section on the China Lake area.) Evaporite minerals—including sodium chloride, sodium carbonate, and some borates—accumulated in the playa and lacustrine beds as China Lake receded (St.-Amand, 1986, p. 15). Reference has been made earlier in this report to the association between high ground-water chloride concentrations and lacustrine deposits. An extreme example is well 25S/39E-31R1 (Supplemental Data B), in which chloride constitutes 30 percent of dissolved solids by weight. Similarly, boron concentrations in ground water always are low (less than 1,000 µg/L) where lacustrine deposits are absent, but increase markedly within the area of lacustrine deposits. For example, boron concentration is 5,400 µg/L in well 26S/39E-8E1 (DSC=724 mg/L) and 18,000 µg/L in well 25S/39E-31R1 (DSC=1,890 mg/L). Finally, the presence of sodium carbonate is suggested at wells 25S/39E-31R1, 26S/39E-8E1, and 26S/39E-17F2, which have pH values of 9.4, 9.0, and 8.8, respectively. Such high pH values can occur only if there is insufficient calcium (calcite) to buffer the water. Coincidentally, such low calcium concentrations permit fluoride (where present) concentrations to rise dramatically (although even the high-fluoride ground water remains undersaturated with respect to fluorite in all but a very few wells in Indian Wells Valley). Calcium is virtually absent (1.5 mg/L) in well 26S/39E-8E1, resulting in a fluoride concentration of 9.5 mg/L, which far exceeds the drinking-water standard (California Department of Health, 1977) of 1.4 mg/L.

A comparison of chemical composition between well 26S/39E-8E1 and well 26S/39E-17F2, about 1 mi apart, might be instructive to any generalizations on the relation between ground-water quality and depth, particularly near the outer margins of the ancient lake. Both wells are perforated at great depth—570 to 880 ft for well 26S/39E-8E1, where land-surface altitude is 2,318 ft, and 681 to 881 ft for well 26S/39E-17F2, where land-surface altitude is slightly higher at 2,340 ft. High boron concentration (5,400 µg/L) and pH (9.0), and moderate chloride concentration (58 mg/L) and DSC (724 mg/L) indicate that well 26S/39E-8E1 penetrates the lacustrine deposits. Although no isotope data are available from well 26S/39E-8E1, comparison with data from other nearby wells (for example, δD in well 26S/39E-7N2 is -94 permil) indicate that its δD probably is no more negative than -95 permil. Compare these data with low boron (600 µg/L), chloride (7.2 mg/L), and DSC (173 mg/L) but high pH (8.8) in well 26S/39E-17F2, for which δD is highly negative at -102 permil. The comparison suggests that lower-DSC, sodium carbonate- and fluoride-rich ground water, recharged at higher altitudes or during an earlier pluvial period, might be present beneath higher-DSC ground water at the margins of the ancestral lake—a conclusion opposite to that presented by Warner (1975).

Historical data since the early 1950's from wells 26S/39E-5F1 and 26S/39E-19P1 were evaluated and used to discern trends for DSC in the western area. As in the northwestern area, DSC values fluctuate but no trend is evident (see pl. 1).

SOUTHWESTERN AREA

The scant data that were collected from the southwestern area during 1987-88 (Supplemental Data B and pl. 1) have been supplemented for this assessment with additional historical data. Water-level contours indicate that the direction of ground-water movement in this area is from southwest to northeast (fig. 3), and isotopic evidence suggests that recharge originated as recent precipitation in the Sierra Nevada. Because the altitude of the Sierra Nevada generally decreases toward the south, isotopic ratios in recharge would be expected to be higher (less negative) in the southwestern area than in the western area. Meridional trends in the Sierra Nevada also could be associated with different storm tracks, with a greater proportion of precipitation in the southern part of the range from the subtropical ocean as opposed to the Gulf of Alaska (Benson and Klieforth, 1989; Friedman and others, 1992; and Ingraham and Taylor, 1991). Delta-D contours mapped by Buchanan (1989, fig. 3*A*, p. 121), on the basis of data from nonthermal springs in the Great Basin, indicate that gradients are as high as 0.3 permil/km in southern Nevada. A gradient of that magnitude would be sufficient to account for the north-south isotope trend exhibited by ground water on the west side of Indian Wells Valley. In fact, except for two wells (27S/38E-31D1 and 28S/37E-13F1) in the extreme southwest corner (pl. 1), the lightest (most negative) δD recorded by this study from wells in alluvial deposits of the southwestern area is only -93 permil (table 1), and δD=-90.5 permil for a shallow domestic well (well 26S/37E-26L1) at an altitude of 4,320 ft in Walker Canyon in the Sierra Nevada (pl. 1). Delta D is at least 5 permil higher (less negative) in alluvial deposits of the southwestern area than in alluvial deposits near the mountains in the western area.

The southwestern area is on an elevated spur of the valley between the Sierra Nevada and the El Paso Mountains. The entire area lies above the maximum altitude of the ancestral China Lake and, hence, is free of any fine-grained lacustrine deposits and the associated high-DSC ground water. As a result, DSC is low and chloride, at about 20 to 30 mg/L, composes only about 10 percent (on a weight basis) of the dissolved solids. The lone exception is well 28S/37E-13F1 in the far southwest corner of this area, for which the chloride concentration at a depth of 400 ft (perforated interval not known) was 138 mg/L and DSC was 419 mg/L in 1953. Sampling in late 1990 confirmed the historical data. The well may penetrate the relatively shallow continental deposits in this part of the valley and therefore would not be representative of water quality in the overlying alluvium. The very different chemical characteristics then would imply a water of distinct ion ratios (or water type) in the continental deposits that contributes little mixing to ground water in the overlying alluvium. Isotope ratios from ground water in the continental deposits would establish whether a distinct difference exists between ratios in these less permeable deposits and in the overlying alluvium.

A highly negative δD value of -108 permil in a 1982 sample from well 27S/38E-31D1, a 375-foot-deep well 3.5 mi northeast of well 28S/37E-13F1 (Gleason and others, 1992; G.I. Smith, USGS, written commun., 1990), was considered so anomalous (at 15 permil more negative than any hydrogen-isotope ratios measured in the present study from other wells in the southwestern area) that it was resampled in late 1990 for hydrogen and oxygen isotopes. In the samples collected in 1990 from this and another well nearby, δD=-105.5 permil and $\delta^{18}O$= -13.50 permil for well 27S/38E-31D1 and δD=-101.5 permil and $\delta^{18}O$= -13.40 permil (near the local ground-water line) for well 28S/37E-13F1 (Tyler Coplen, USGS, written commun., 1990), confirming the 1982 analyses. The highly negative δD values indicate that, in most of the southwestern area, little of the ground water comes from recharge in the far southwestern corner, where the isotopically light composition matches more closely that of ground water in basins immediately south of Indian Wells Valley (Gleason and others, 1992).

Pumpage in the southwestern area has been negligible, and it is unlikely to have affected ground-water conditions. Although well 27S/39E-7R1 has been sampled only four times since 1946, DSC at this well has shown very little historical variation (pl. 1).

CHINA LAKE AREA

The chemical and isotopic characteristics of ground water in the China Lake area show an extremely broad range. Several factors—that also operate, although to a lesser extent, in other areas of Indian Wells Valley—are responsible for this range in composition, including:

1. Proportion of alluvial or lacustrine (and playa) deposits tapped by wells.
2. Presence of historical (that is, older—perhaps late Pleistocene) recharge.
3. Evaporative concentration and (or) dissolution of evaporite minerals.
4. Microbial sulfate reduction in the aquifer.
5. Human activities—including pumping, surface disposal of wastewater, and discharge of contaminants.

The first four factors are discussed in greater detail in the section, "Geochemical Processes and Historical Recharge." Human activities exert a more localized influence on ground-water quality, and examples of such activities are noted in this section.

Fresh ground water exists in the southwest part of the China Lake area, where the presence of alluvium and flow of low-DSC ground water from the southwest (and intermediate) areas predominates. The result is DSC less than 300 mg/L at wells 26S/40E-17J1 and 26S/40E-17R1 in the extreme southwest part of the China Lake area, increasing to about 500 mg/L at wells 26S/40E-4Q1 and 26S/40E-9A1 about 2 mi northeast, with further increases in the direction of decreasing land altitude toward the China Lake playa (pl. 1).

Nearly all the China Lake area is underlain by a large thickness of lacustrine deposits from the ancestral lake. Ground water pumped from wells tapping these deposits in the deep aquifer generally is slightly saline (DSC= 1,200-1,500 mg/L). Some deep wells in and near the China Lake playa produce highly saline water; for example, DSC=60,700 mg/L in well 26S/40E-6C1.

Low DSC at well 25S/40E-20F1 is particularly puzzling given its location so far from the source of fresh ground water. Its DSC of about 500 mg/L (calculated from specific conductance in table 1) suggests that water in this well could not have flowed through lacustrine deposits for any significant distance. And the isotopic data (δD= -80 permil, $\delta^{18}O$=-10.5 permil) indicate (on the basis of deviation from the meteoric-water line) that the water has undergone little or no evaporation. This δD value is at least 10 permil more positive (less negative) than that of any other sample from the China Lake area that also shows little or no evaporation (fig. 11). The result is so unusual that the possibility must be considered that water in the well was diluted by surface runoff which collects in the immediate vicinity of the well after rains and which might flow down alongside the well casing. Present-day precipitation has an average annual δD of -81 permil at Inyokern and -85 permil at Walker Pass in the Sierra Nevada (Friedman and others, 1992, and G.I. Smith, USGS, written commun., 1990).

Moderately saline ground water with DSC as high as several thousand milligrams per liter can be produced from wells at shallow depths. The high DSC is due to evaporative concentration. An example is well 26S/40E-14B1 (perforated from 20 to 22 ft), in which DSC is about 3,000 mg/L and δD is -80 permil. Evaporation at some stage in the history of this water is indicated in figure 11 by the large deviation of its δD-$\delta^{18}O$ datum from the meteoric-water line.

In fact, ground water at well 26S/40E-14B1 may have been even more saline and isotopically enriched under natural conditions. This well is near the northern edge of an area where the Ridgecrest Wastewater Treatment Facility maintains evaporation

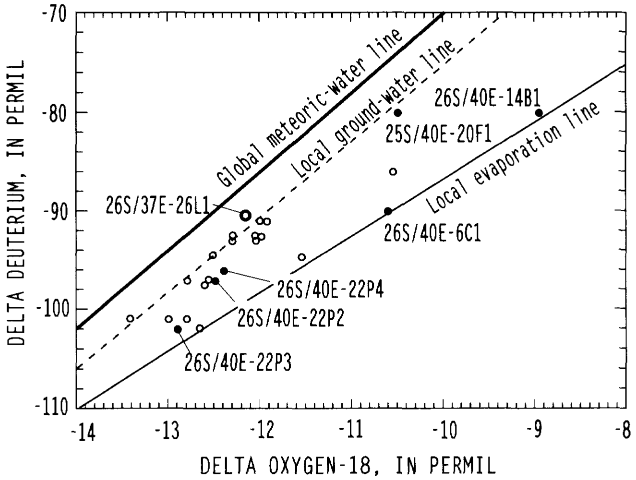


Figure 11. Relation between stable-isotope ratios of hydrogen and oxygen in samples from selected wells in the China Lake area. Data are given in table 1, and location of wells is shown on plate 1. (Wells that are referred to in the text are numbered on the graph; well 26S/37E-26L1, in the Sierra Nevada in the western part of the study area, is included for comparison.

ponds containing low-DSC treated wastewater (500 mg/L, according to Warner, 1975, p. 24-25, and Don L. Greenfield, Supervisor, Ridgecrest Wastewater Treatment Facilities, oral commun., 1989). Because water is supplied from deep wells, the treated wastewater is presumed to be isotopically light in comparison with shallow ground water. Infiltration from the ponds has created a local ground-water mound leading to possible dilution in shallow wells by isotopically lighter, less saline water. Ground water at well 26S/40E-14L1 (perforated from 55 to 57 ft), a short distance south of well 26S/40E-14B1, is clearly affected by wastewater. Water from this well is only about half as saline as water from well 26S/40E-14B1 and it has a δD of -86 permil. A mixing model used to match the chemical and isotopic composition at well 26S/40E-14L1 from the composition of wastewater and ground water at well 26S/40E-14B1 requires that the infiltrating wastewater that reaches well 26S/40E-14L1 have a δD of -95 permil.

Disposal of synthetic organic contaminants is another example of how human activities affect ground-water quality in the China Lake area. Between 1984 and 1988, the USGS analyzed for purgeable organic priority pollutants (U.S. Environmental Protection Agency, 1985) 26 samples from 15 wells near the China Lake Naval Air Weapons Station's industrial-waste ponds near the community of China Lake. Some early results were presented by Berenbrock (1987, p. 55). The purpose in summarizing results in table 4 is not to provide a comprehensive assessment of contamination but rather to illustrate another way in which human activities affect the ground-water quality. Only those wells in which concentrations exceeded the detection limit, which was 3 μg/L in most samples, are listed in table 4, along with proposed or recommended maximum contaminant levels in drinking water.

Unequivocal evidence of contamination generally requires detection in more than one sample and over a period of time. Nevertheless, the occurrence of vinyl chloride in wells 26S/40E-15N2 and 26S/40E-22H3 is mentioned because this compound can be produced by bacterial dechlorination of trichloroethylene under anoxic conditions (Barrio-Lage and others, 1986). Trichloroethylene apparently has been used in the area—as indicated by trichloroethylene concentrations that ranged from 94 to 270 μg/L in four samples (collected from well 26S/40E-22B1) over a period of 4 years (table 4).

In addition to the wide range in ground-water DSC, very large differences in relative concentrations of major ions exist in ground water from the China Lake area. For low-DSC ground water in the alluvium, concentrations of the major cations (sodium and the alkaline earths) and the major anions (chloride, sulfate, and bicarbonate) are somewhat comparable, with no overwhelming abundance of any particular ion. However, the more saline ground water in the China Lake lacustrine (and playa) deposits typically has very high sodium accompanied by nearly negligible alkaline-earth concentrations, and high chloride and (or) bicarbonate accompanied by low sulfate concentrations. The waters are poorly buffered and typically have high fluoride concentrations and pH values close to or exceeding 9 (see Supplemental Data B).

As discussed in more detail in a later section of this report, sulfate reduction is the geochemical process most likely responsible for the large shift in relative concentrations of major ions and of some minor constituents. The process extends throughout all but the very shallow lacustrine deposits of China Lake. However, the lacustrine deposits within 200 to 400 ft of land surface in and near Mirror Lake, on the southern boundary of the China Lake area, appear to be much less reducing. Accordingly, wells that tap these deposits produce water with DSC a little higher than 1,000 mg/L but with substantial alkaline-earth and sulfate concentrations; however, the water still is undersaturated (saturation indices range from slightly less than 1 to 10^{-2}) with respect to gypsum. (A similar geochemical pattern exists in Satellite Lake, which is immediately south of Mirror Lake and in the southeastern area discussed in the next section.) Water-quality characteristics (including depletion in sulfate) indicate that the deposits more than 200 to 400 ft below land surface are more reducing (wells perforated between 200 and 400 ft to delineate more

Table 4. Concentrations of purgeable organic priority pollutants in water samples from selected wells in Indian Wells Valley

[Compounds designated as priority pollutants by U.S. Environmental Protection Agency (1986, p. 738-740). MCL, maximum contaminant level; μg/L, micrograms per liter; <, less than value indicated]

Well	Date	Vinyl chloride (μg/L)	Chloroform (μg/L)	1,2-trans-dichloroethene (μg/L)	Trichloroethylene (μg/L)	Ethyl benzene (μg/L)	1,1,1-Trichloroethane (μg/L)	Toluene (μg/L)	Methylene chloride (μg/L)
26S/40E-15N2	06-11-85	6.2	<3	<3	<3	<3	<3	<3	<3
26S/40E-22A1	05-31-87	<3	<3	<3	5.5	<3	<3	<3	<3
	07-06-88	<3	<3	<3	3.2	<3	<3	<3	<3
26S/40E-22B1	08-09-84	<3	12	6	94	<3	<3	<3	<3
	04-17-86	<3	7.9	5.3	170	<3	<3	<3	<3
	06-02-87	<3	17	10	270	<3	<3	<3	<3
	07-07-88	<3	9.7	3.1	170	<3	<3	<3	<3
26S/40E-22H3	08-09-84	7.9	<3	<3	<3	<3	<3	<3	<3
26S/40E-22P3	01-09-86	<3	<3	<3	<3	4	23	<3	<3
26S/40E-22P4	08-09-84	<3	<3	<3	<3	<3	<3	<3	4
26S/40E-23D1	01-09-96	<3	<3	<3	<3	<3	8	3.1	<3
Proposed or recommended MCL (μg/L)		[1]2	None	None	[1]5	None	[1]200	[2]2,000	None

[1] U.S. Environmental Protection Agency, 1987.

[2] U.S. Environmental Protection Agency, 1985.

precisely the depth to this geochemical transition are not available), and thus chemical composition in ground water from the few deep wells in Mirror (and Satellite) Lake is similar to chemical composition from wells in China Lake.

Historical data from wells 26S/39E-12R2, 25S/40E-8A1, 26S/40E-5P1, and 26S/40E-22P1, and less extensive historical data from well 26S/40E-22B1, show no appreciable trend for ground-water DSC in the China Lake area since the early 1950's. Historical data are shown on plate 1. (A 1988 analysis from nearby well 26S/40E-4Q1 was substituted for well 26S/40E-5P1 because the latter well could not be sampled during this study.) Apparently, pumping in the intermediate area and at Ridgecrest has not caused a degradation of ground-water quality in this part of the valley, in contrast to observed DSC increases in the southeastern and intermediate areas that are discussed in the next two sections of this report.

SOUTHEASTERN AREA

The DSC of ground water in the southeastern area ranges from about 300 to 400 mg/L at the western end to greater than 1,000 mg/L near Satellite Lake (pl. 1). The only deep well located on Satellite Lake from which chemical data were obtained in the 1987-88 sampling is 26S/40E-27E3, which is perforated from 400 to 460 ft below land surface (Supplemental Data B). Its low DSC of 479 mg/L (obtained from a sample collected February 26, 1989) indicated that perhaps high-DSC ground waters and their associated lacustrine deposits do not extend to great depth beneath Satellite Lake. The low DSC of 285 mg/L obtained from a 1988 sampling of well 26S/40E-35H2, which is perforated from 360 to 500 ft, seemed to support this conclusion; however, because the well was believed to have been inadequately developed, it was resampled in 1990. (Data not given in Supplemental Data B.) The second sample had a DSC of 1,630 mg/L and the same high-chloride, low-sulfate concentrations found in deep deposits beneath Mirror Lake. Deeper wells are needed at both Mirror and Satellite Lakes to determine the depth to which lacustrine deposits extend beneath the lakes and to define precisely the depth at which the transition to highly reduced low-sulfate ground water occurs.

In contrast to the general absence of water-quality trends observed in other, less populated parts of Indian Wells Valley, several wells in the southeastern area show clear evidence of increasing DSC. Examples shown on plate 1 include the production wells 26S/40E-34N1, 26S/40E-28J1, and 27S/40E-4C2 (which was abandoned and sealed in the early 1980's because of high DSC) in eastern Ridgecrest and the observation well 26S/40E-36A1. Extensive pumping near Ridgecrest and in the intermediate area to the west (discussed in the next section) has created a cone of depression (see fig. 3) inducing flow of high-DSC ground water from beneath Satellite and Mirror Lakes toward the production wells to the west. As shown by the graphs in figure 12, increasing DSC is accompanied by a shift toward increasing relative concentration of alkaline-earth cations and sulfate at the expense of sodium and (carbonate) alkalinity. Such compositional changes would be expected (on the basis of the aforementioned chemical characteristics of shallow ground water beneath Satellite and Mirror Lakes) and are inconsistent with flow from the chloride-rich, sulfate-poor ground water beneath the China Lake playa. Kunkel and Chase (1969, p. 53) also concluded that this water was not the source of high DSC in the eastern and southern parts of Ridgecrest.

On the basis of small deviation of its isotopic values from the local ground-water line developed using data from the China Lake area (discussed in "Pleistocene Ground Water" section), ground water from the southeastern area shows little or no evidence of evaporation. The most negative isotope ratios measured in ground water from alluvial and lacustrine deposits of Indian Wells Valley (table 1) were found in deep wells of the southeastern (δD=-105 permil in well 26S/40E-32F3) and intermediate areas (δD=-108 permil in well 27S/40E-6D1). Both wells are perforated from more than 500 to 700 ft below land surface. Such δD values are 10 to 15 permil more negative than those of ground water from the southwestern area, and this difference indicates that the source of most of the deep ground water could not be from flow of recently recharged ground water from the southwestern area. The most likely source of recharge to the southeastern (and intermediate) areas is from precipitation in the higher altitudes of the Sierra Nevada toward the west and north and (or) from recharge during pluvial periods, as was indicated for the western and China Lake areas.

An alternative possibility is that recharge was derived from the El Paso Mountains to the south during an earlier pluvial period. (Present-day climate is too arid for significant recharge.) This alternative seems unlikely because ground-water traveltimes from there probably would be too short. Altitudes lower than those of the Sierra Nevada also would seem to favor less negative isotopic ratios than are found in the deep wells. No data are available from wells in or near the El Paso Mountains that could provide isotopic information.

Another alternative is that the highly negative δD values reflect input of ground water from deeper continental deposits. However, chemical evidence (isotopic data from continental deposits are not available) in the southwestern area indicates that there is little or no contribution of ground water from continental deposits to the alluvium. Two wells at altitudes near or above the ancestral lake shoreline and for which historical chemical data exist are 27S/40E-10R1 and 27S/40E-15L1 (pl. 1). Well 27S/40E-10R1 is a 262-foot-deep well at an altitude of 2,380 ft and 27S/40E-15L1 is a 277.5-foot-deep well at an altitude of 2,470 ft. Perforated intervals are not known for either well. Chemical concentrations (from historical data in WATSTORE, a USGS computerized data base) in water from well 27S/40E-10R1 are so unusual that they are believed to represent anomalous local conditions (extensive mineral dissolution and redox reactions). Between 1974 and 1980, iron concentration was about 100 mg/L, pH was acidic (4.7 to 6.5), and chloride exceeded sulfate by a factor of 100. Well 27S/40E-15L1, for which chemical data exist for 1961-74, had a DSC of 1,000-1,500 mg/L, and chloride constituted 50 percent of the dissolved solids. This ground water is much too saline to contribute significantly to the less saline and isotopically light ground water that occupies the deep aquifer farther north of the El Paso Mountains. It is not known for certain whether well 27S/40E-15L1 is perforated entirely within alluvium, rather than partially in the deeper continental deposits. Chloride constitutes 50 percent of the dissolved solids, and similarly high chloride proportions were noted in the one well (28S/37E-13F1) thought to penetrate continental deposits in the southwestern area.

INTERMEDIATE AREA

The DSC in ground water over nearly all the intermediate area is between 200 and 300 mg/L, and chloride constitutes about 10 percent of the dissolved solids. These chemical characteristics are similar to those of the southwestern area, although ground water in the intermediate area is isotopically lighter (δD is 10 to 15 permil more negative). Only in the extreme northeastern part of the intermediate area is DSC relatively high (see pl. 1). The higher DSC is associated with proximity to the ancestral China Lake deposits. On the basis of historical data in WATSTORE, it can be calculated that the increase in DSC is accompanied by an increase to 20 percent in the proportion of chloride to dissolved solids at wells 26S/39E-13R4, 26S/39E-2N1, and 26S/39E-11E1. Depth-related chemical data seem to indicate that higher DSC is correlated with increasing depth in the aquifer. The DSC at well 26S/39E-13R3, which is perforated between 150 and 260 ft, is only 223 mg/L; by contrast, DSC at well 26S/39E-13R4, perforated from 640 to 800 ft, is 826 mg/L.

The intermediate area contains several large-capacity production wells, and the possibility exists that pumping from these wells could induce flow of high-DSC ground water from the China Lake area. At present, however, there is little indication of DSC change in the area. Wells 26S/39E-24P1, 26S/39E-23J1, and 26S/39E-24M1 show no trend (pl. 1). The

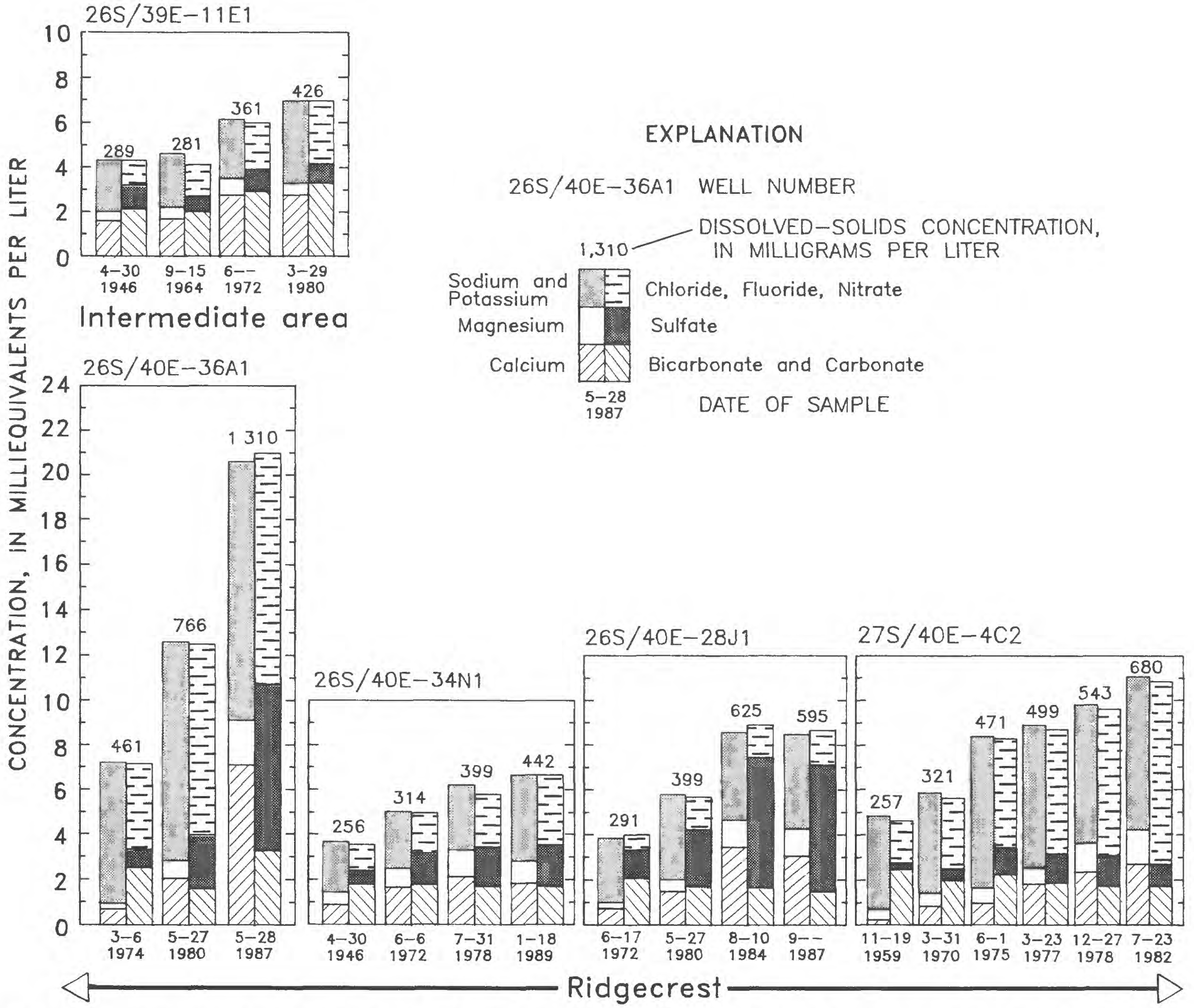


Figure 12. Historical trends in dissolved-solids concentration and major-ion proportions in samples from selected wells at Ridgecrest and in the intermediate area.

only well that has a discernable trend is 26S/39E-11E1, in which the increase is from about 250 mg/L in the mid-1960's to about 450 mg/L in 1980 (pl. 1). The graph in figure 12 indicates that the increase in DSC is accompanied by a small increase in the proportion of chloride, as would be expected if pumping was inducing southward flow of ground water from lacustrine deposits of China Lake.

The isotopic composition at well 27S/40E-6D1, a deep production well perforated from 580 to 700 ft, is mentioned here because its δD of -108 permil is the lightest ratio measured for this study. Curiously, the position of its δD-δ^{18}O datum point is closer to the evaporation line than it is to the local ground-water line (although the two lines are separated by only 4 permil on the δD axis at this point), indicating that this ground water has been subjected to some evaporation during its history. The isotopically light character of deep ground water in the intermediate area (table 1) is consistent with recharge during a pluvial period and (or) recharge from precipitation in the higher altitudes of the Sierra Nevada toward the north, as was noted for the western, China Lake, and southeastern areas.

The isotopic data (δD=-102.0 permil and δ^{18}O= -8.20 permil) from well 26S/40E-30K1, perforated from 250 to 800 ft, are so anomalous that they are believed to be erroneous. The datum (not shown) lies so far to the right of even the evaporation line in figure 11 that it could only be produced by extensive dissolution of ^{18}O-bearing minerals or by extreme evaporation of the sample during storage at the laboratory. Unfortunately, no sample was available for reanalysis by the laboratory. We note, however, that values of δD=-99.5 permil and $\delta^{18}O$= -13.25 permil (table 1) in the adjacent well 26S/40E-30K2, which is perforated from 220-470 and from 600-760 ft, are similar to other isotopic data from deep deposits in Indian Wells Valley.

GEOCHEMICAL PROCESSES AND HISTORICAL RECHARGE

Several geochemical reactions, evaporation, and location and timing of recharge cause a wide range in salinity, relative concentration of major ions (also minor constituents), and isotope ratios in ground water from Indian Wells Valley. Ground-water data from the China Lake area were chosen to illustrate the significance of these processes because of the prevalence of lacustrine deposits, a shallow water table from which evapotranspiration can occur, and the long time required for ground water to move from the areas of mountain-front recharge. The elucidation of these processes provides a framework for]understanding the wide range in ground-water quality, determining what kind of (and from where) additional data are needed, estimating isotopic characteristics of historical (late Pleistocene) recharge, and anticipating possible degradation of the ground-water resource.

DISSOLUTION OF EVAPORITES

Within the topographically lowest parts of the valley and near the China Lake playa, ground-water DSC can reach extremely high concentrations, even in wells that are several hundred feet deep. One example is well 26S/40E-6C1, perforated from 500 to 600 ft, in which DSC exceeded 60,000 mg/L and numerous salt crystals (not identified) coated the sampling pump when it was withdrawn. Smith and Pratt (1957) reported crystals of gaylussite, a rather soluble sodium-calcium carbonate pentahydrate, from 100 to 200 ft below land surface in a core from China Lake. These observations, along with very high chemical concentrations in well 26S/40E-6C1 (and a few other wells), elicit the possibility that re-solution of evaporite minerals is a source of increasing DSC in ground water from the lacustrine and playa deposits in this part of Indian Wells Valley.

The relation between specific conductance and δD can be used to ascertain whether re-solution of soluble minerals or concentration by partial evaporation (from the water table) is the cause of increased salinity in ground water. If evaporation is the only process, δD should increase with increasing DSC. If re-solution of evaporites is the dominant process, δD and DSC would not be correlated. (Refer to Fritz and Fontes, 1980, p. 121-125; and Schroeder and others, 1991, for examples of the foregoing application.) To some extent, the reasoning is compromised in Indian Wells Valley because the isotopic signature of precipitation is variable in the surrounding mountains and likely has changed through time. Nevertheless, large scatter in the plot between specific conductance and δD shown in figure 13, particularly in the high conductance range above about 2,000 µS/cm, does suggest that re-solution of evaporites is an important process.

BACTERIAL SULFATE REDUCTION

Because chloride salts are highly soluble, it might be anticipated that chloride should become an increasing proportion of DSC as DSC increases. However, as shown in figure 14, there are numerous wells in the China Lake area in which DSC exceeds 1,000 mg/L but in which chloride constitutes less than 10 percent of the dissolved solids. As shown in the figure, several samples of this type come from the Mirror Lake lacustrine deposits that crop out at the far southern end of the China Lake area. (Similar patterns of concentration exist in the Satellite Lake deposits.)

The plot of chloride against sulfate concentration in figure 15 provides an indication of the very large variation in major-ion chemical composition of ground water. The figure illustrates that wells in the Mirror Lake area (except for two deep wells, 26S/40E-22P1 and 26S/40E-22P3) contain sulfate-rich ground water. A few shallow wells elsewhere in the China Lake area also contain sulfate-rich ground water. Shown in this figure, also, are several wells at which the odor of hydrogen sulfide was prominent during sampling. Note that chloride concentration exceeds sulfate concentration in those same wells. Bacterial sulfate reduction in the fine-grained lacustrine (and the playa) deposits of Indian Wells Valley very likely is the main cause of wide variations in major-ion chemical composition.

Sulfate reduction generates alkalinity according to the idealized stoichiometric reaction:

$$SO_4^{2-} + 2CH_2O \rightarrow H_2S + 2HCO_3^- .$$

In principle, the progress of this reaction could be followed along a ground-water flow path, but the acquisition of data appropriate to this task was not specifically sought for this study. However, the relative importance of sulfate reduction—along with precipitation of carbonates, dissolution of halite, and ion exchange on clays—can be illustrated by comparison of depth-related data at a single location. Wells 26S/40E-22P2 and 26S/40E-22P4 (perforated from 73 to 75 and from 200 to 215 ft, respectively) and 26S/40E-22P1 and 26S/40E-22P3 (perforated

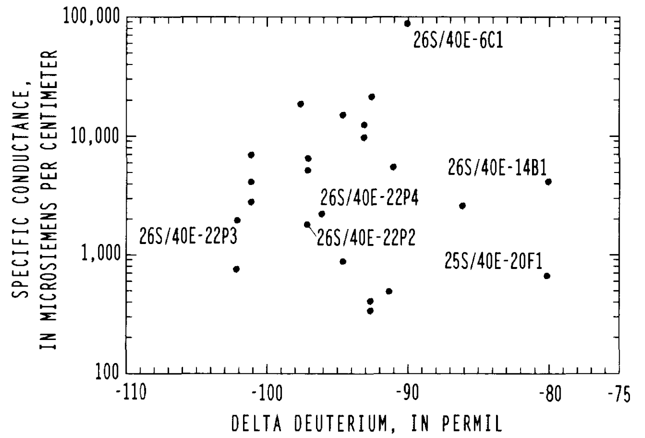


Figure 13. Relation between specific conductance and delta deuterium in samples from wells in the China Lake area. (Wells that are referred to in the text are numbered on the graph; location of wells is shown on plate 1.)

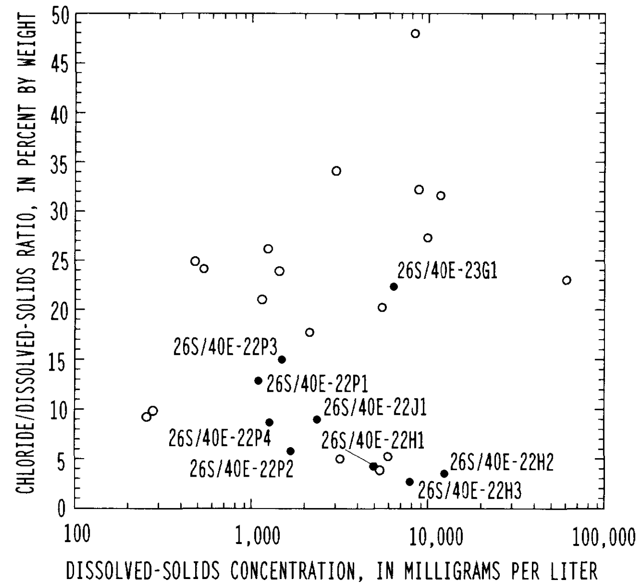


Figure 14. Relation between percent chloride and dissolved-solids concentration in samples from wells in the China Lake area. Solid circles designate wells in lacustrine deposits of Mirror Lake. (Wells that are referred to in the text are numbered on the graph; location of wells is shown on plate 1.)

from 530 to 830 and 400 to 415 ft, respectively) are chosen for this example. Data from the two shallow wells and from the two deep wells were averaged for the calculation.

Validity of this comparison, based on mass- and electrical-balance calculations, rests on the assumption that the shallow and deep ground water were the same, or at least very similar, before reactions altered their composition. The assumption is reasonably good for dissolved solids as DSC is only slightly higher in the shallow wells, and isotopic data from shallow and deep wells show about equal deviation from the meteoric-water line, thus demonstrating equivalent evaporation histories. However, δD is about 5 permil more negative at -102 permil in well 26S/40E-22P3 than in the two shallow wells, suggesting that the deep ground water there originated during a colder and (or) wetter period or at higher altitudes than did the shallow ground water.

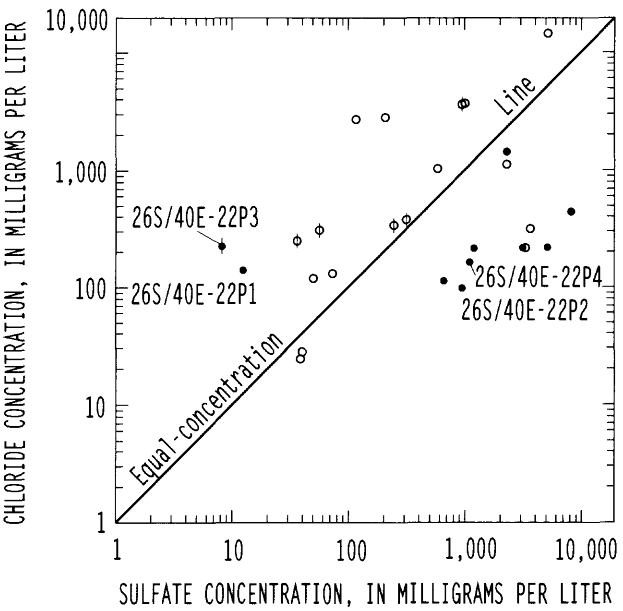


Figure 15. Relation between chloride and sulfate concentrations in samples from wells in the China Lake area. Solid circles designate wells in lacustrine deposits of Mirror Lake; circles with a vertical line through them designate wells in which a strong odor of hydrogen sulfide was noted during sampling. (Wells that are referred to in the text are numbered on the graph; location of wells is shown on plate 1.)

Comparison of average major-ion concentrations in the two shallow wells (Na^++K^+=9 meq/L, Mg^{++}+Ca^{++}=16 meq/L, Cl^-=3 meq/L, SO_4^{2-}=21 meq/L, and HCO_3^-=2 meq/L) with concentrations in the two deep wells (Na^++K^+=21 meq/L, Mg^{++}+Ca^{++}=0.2 meq/L, Cl^-=6 meq/L, SO_4^{2-}=0.2 meq/L, HCO_3^-=16 meq/L) shows the expected increase in (bicarbonate) alkalinity associated with sulfate reduction in the deep deposits. Subtracting concentrations (and reporting gains and losses per liter of water) shows that microbial sulfate reduction has removed 21 meq SO_4^{2-} in the deep zone, which according to the reaction should yield the same equivalent of bicarbonate (plus carbonate if pH is sufficiently high). Increase in dissolved HCO_3^- (alkalinity) is only 14 meq; therefore, the remaining 7 meq is presumed to precipitate alkaline-earth carbonates. However, the decrease of 16 meq in dissolved $Mg^{++}+Ca^{++}$ exceeds by 9 meq the amount removed by precipitation of carbonates. The remainder is removed by ion exchange with Na^+ (and very minor amounts of K^+), which increases by 12 meq/L in the deep ground water. And the extra 3 meq Na^+ comes from dissolution of that much additional halite in the deep zone in comparison with the shallow zone. In summary, calculations based on mass balance and electroneutrality illustrate the following difference between the deep (greater than 400 ft) and shallow (less than 200 ft) zones at one site (in Mirror Lake):

Reaction/process	Net difference, in milliequivalents
Sulfate reduction and alkalinity generation	21
Carbonate precipitation	7
Ion exchange	9
Halite dissolution	3

It is apparent from this calculation that bacterial sulfate reduction can have a profound effect on the chemical composition of ground water in the China Lake area. In fact, the process occurs throughout the valley's lacustrine deposits, probably even to some extent at rather shallow depths in the example cited (the 1954 field log for well 26S/40E-22P1 indicated initial appearance of gray clay as shallow as 30 ft). This process also explains why nitrate (whose reduction precedes that of sulfate) is detected in few deep wells but averages 2 mg/L (as nitrogen) in shallow wells from the China Lake area, and why manganese and iron (whose solubilities are greater in reduced oxidation states) concentrations generally are much higher in deep than in shallow wells (see Supplemental Data B). For some reason (possible explanations include cropping out of the lacustrine deposits, coarser texture, higher altitude, or less organic matter), the near-surface environment seems less reducing in the vicinity of Mirror and Satellite Lakes than in the much larger China Lake, and this difference has allowed the retention of sulfate-rich ground water to a greater depth (somewhere between 200 and 400 ft below land surface for the example cited) in the Mirror and Satellite Lake deposits. As a result, these shallower ground waters beneath Mirror and Satellite Lakes are only slightly undersaturated (saturation indexes greater than 10^{-2}) with respect to gypsum, in contrast to the high-chloride, low-sulfate deeper ground waters, which have saturation indexes of 10^{-4} to 10^{-5}.

PLEISTOCENE GROUND WATER

Two pieces of evidence suggest that some ground water in the China Lake area might be as old as late Pleistocene. As already noted, ground water is isotopically lighter in deep wells than in shallow wells (excluding those containing water that has undergone substantial evaporation), indicating that the deep ground water could have been recharged during a colder/wetter time. Merlivat and Jouzel (1979) concluded that hydrogen-isotope depletions exceeding 5 permil relative to present-day meteoric water reflect paleorecharge in a colder and more humid climate. Also, as discussed in a preceding section of this report, an advective ground-water model inferred traveltimes of at least several thousand years between parts of the valley.

A more detailed interpretation of hydrogen- and oxygen-isotope data can be used to estimate the isotopic composition of recharge in the past. Because this estimate falls substantially below (more negative than) isotope ratios found in springs or in precipitation, it lends additional credence to the argument for "old" ground water in Indian Wells Valley. Ground waters that most readily suggest the existence of great age, on the basis of isotopic values alone, are those that are isotopically light (δD more negative than -100 permil) despite showing evidence of evaporation (deviation from the meteoric-water line). As shown in figure 11, one sample from the China Lake area that has both characteristics is from well 26S/40E-22P3, which coincidentally is a deep well that was used in the foregoing discussion on sulfate reduction. Samples of water from other wells in the China Lake area that have both characteristics also are evident in the figure.

To extrapolate back in time to the average isotopic composition of recharge requires determination of:

1. The local meteoric-water line, and
2. A local evaporation line.

Isotopic data from only the China Lake area were used to determine these lines because this part of the valley is most likely to contain the oldest ground water. Although data from a few adjacent wells in other parts of Indian Wells Valley also could have been used, those data would have minimal effect on the calculation.

The local meteoric-water line is assumed to be best approximated by a local ground-water line (fig. 11) that is drawn through the locus of isotopic-datum values that lie closest to the global meteoric-water line. These values presumably represent meteoric waters that have been subjected to the least evaporation. The local line deduced in this way is almost parallel to and about 4 permil (on the δD axis) below the global meteoric-water line, coincidentally closely matching the isotopic relation between δD and $\delta^{18}O$ in precipitation averaged for eight North American stations (Fritz and Fontes, 1980, p. 32).

Most isotopic data from springs in the Coso (additional isotopic data in Fournier and Thompson, 1980) and Argus Ranges and from a shallow well (26S/37E-26L1) in the Sierra Nevada fall between the global meteoric-water line and a local ground-water line, as can be seen in figure 4.

A local evaporation line (fig. 11) is estimated by drawing a line through the locus of isotopic-datum points that lie farthest from the meteoric-water (also ground-water) line. Slope of the evaporation line is about 5.8, which compares favorably with a present-day slope of 5.4 in the Imperial Valley of southern California (Michel and Schroeder, 1989; Schroeder and others, 1991) although it is emphasized that the processes of evaporation in the two arid areas may be dissimilar. Also implicit in this interpretation is the assumption that evaporative processes with a much lower slope have not occurred locally or in the past. The assumption appears to be supported by isotopic data from the central Imperial Valley and might, therefore, be valid in other small closed or semiclosed basins in arid areas of the Western United States. The aforementioned slope of 5.4 in the Imperial Valley was based on evaporation of irrigation drainwater which penetrates only several feet below land surface and which tritium analyses indicate has been recharged during the past several years. This slope is virtually identical to that of 5.5 estimated here for much older ground water from as much as 1,000 ft below land surface in the central Imperial Valley (Coplen and others, 1975, fig. 3, p. 18).

The evaporation line and local ground-water line (when extended) intersect at about δD= -125 permil, much more negative than any a ratio in springs, creeks, and precipitation today. Even with the uncertainty presented by possible historical shifts in local meteoric-water and evaporation lines, such isotopically light recharge requires a sustained period of colder and (or) wetter climate than that of today. Smith and others (1979) and Friedman and Smith (1972) suggested that δD might have been 50 to 85 permil more negative during Pleistocene pluvial periods on the basis of comparisons with values for Sierra Nevada snows during the excessively wet winter of 1968-69. A Pleistocene/Holocene difference of about 20 to 35 permil was found in plant cellulose from fossil packrat middens in east-central Nevada (Siegel, 1983) and the four corners area of the southwestern United States (Long and others, 1990, fig. 17.5, p. 391). A similar shift was measured in radiocarbon-dated ground water from the San Juan Basin of New Mexico (Phillips and others, 1986, figs. 2 and 3, p. 183). A somewhat smaller difference was found between Pleistocene and late Holocene ground water from the Amargosa Desert in west-central Nevada; however, the youngest water present had a radiocarbon age of 9,000 years (Claassen, 1985). The difference estimated from this study is 40 to 45 permil when compared with present precipitation, and 15 to 35 permil when compared with current recharge. Differences based on comparisons with recharge water rather than precipitation itself would tend to be less because recharge and subsequent transport "average" the water's characteristics over a period of many years, and because recharge tends to be greater from higher altitudes and during periods of greater precipitation, which favor depletion in the heavier isotope. Using the formulation of Yurtsever (1975) that relates oxygen-isotope ratios in precipitation and surface mean temperatures at continental locations, and ignoring possible past changes in altitude at which recharge occurred, one obtains an average difference of 2 to 3 degrees Celsius, between "recent" and "old" recharge to Indian Wells Valley.

WATER-QUALITY CHANGES ALONG A GEOLOGIC SECTION

Chemical and isotopic data along a west-east section (fig. 16 and table 5) can be used to summarize the changes in, and processes that affect, ground-water quality in Indian Wells Valley. Lithology and water-table altitude (Kunkel and Chase, 1969; Berenbrock and Martin, 1991) along this same section are shown in figure 17. As shown in the figures, land surface declines steeply eastward from near the Sierra Nevada on the west and then gently toward its lowest point in the playa on the east side of the valley. Delta deuterium decreases (becomes more negative) easterly from about -95 permil near the Sierra Nevada to about -105 permil for deep ground water in the central and eastern parts of the valley. Shallow ground water from the playa is significantly enriched in deuterium by evaporation.

Lateral variations in water quality are minor in the alluvium between the Sierra Nevada and Ridgecrest (DSC generally is between 200 and 300 mg/L), although changes with depth (associated with changes in lithology) are observed in samples from well 26S/40E-32F3 in this section (see fig. 16). Sulfate concentrations (on a weight basis) are slightly greater than chloride concentrations in the alluvium, and pH is about 8. The expected change in chemical quality associated with the transition from alluvial to lacustrine deposits begins to manifest itself between wells 26S/40E-30E2 and 26S/40E-30K1 (and 26S/40E-30K2), where (from west to east) there is an increase in sodium content from about 40 percent to about 80 percent of total cation concentration, and the chloride/sulfate ratio changes from slightly less than 1 to slightly greater than 1. These trends intensify farther to the east: In samples from wells beneath the playa near the community of China Lake (see fig. 16), DSC exceeds 1,000 mg/L and chloride concentration greatly exceeds sulfate concentration in deep zones subjected to microbial sulfate reduction (refer to data from wells 26S/40E-22P1 and 26S/40E-22P3 in table 5). There is some evidence that DSC decreases at the

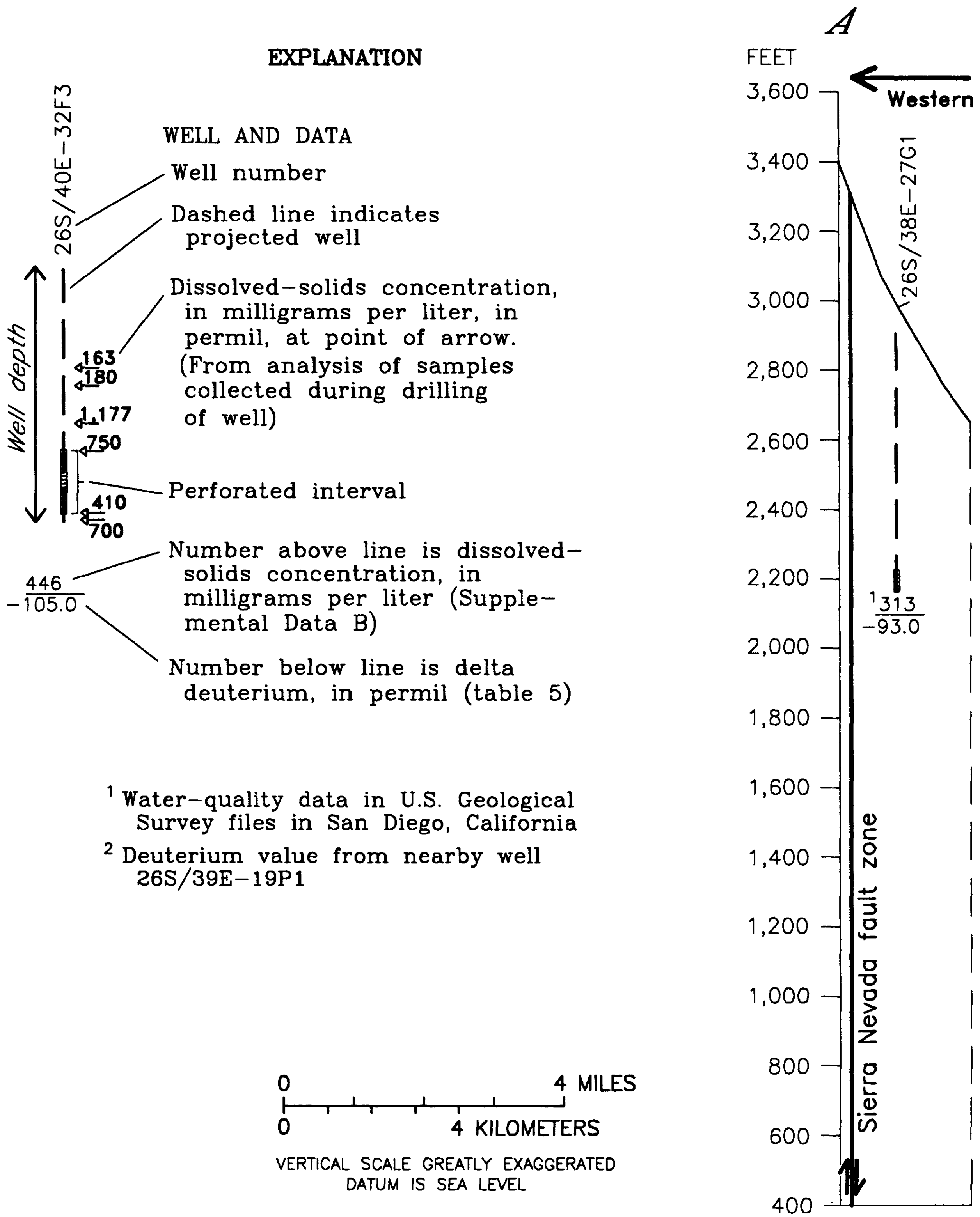


Figure 16. Generalized section of Indian Wells Valley ground-water basin showing vertical and lateral distribution of dissolved-solids concentration and delta deuterium. Line of section is shown in figure 2 and on plate 1; geology for this section is shown in figure 17.

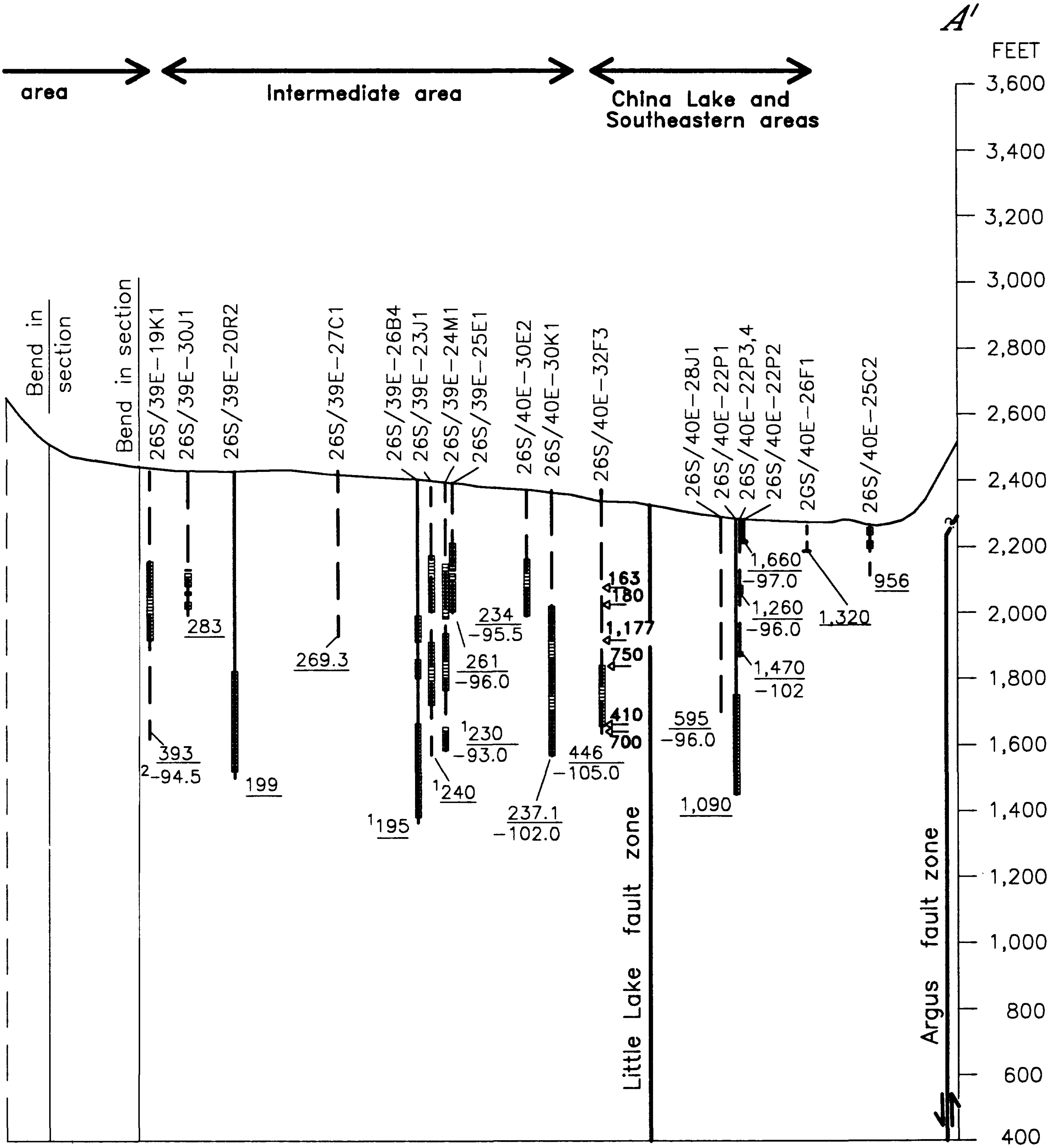


Figure 16. *Continued.*

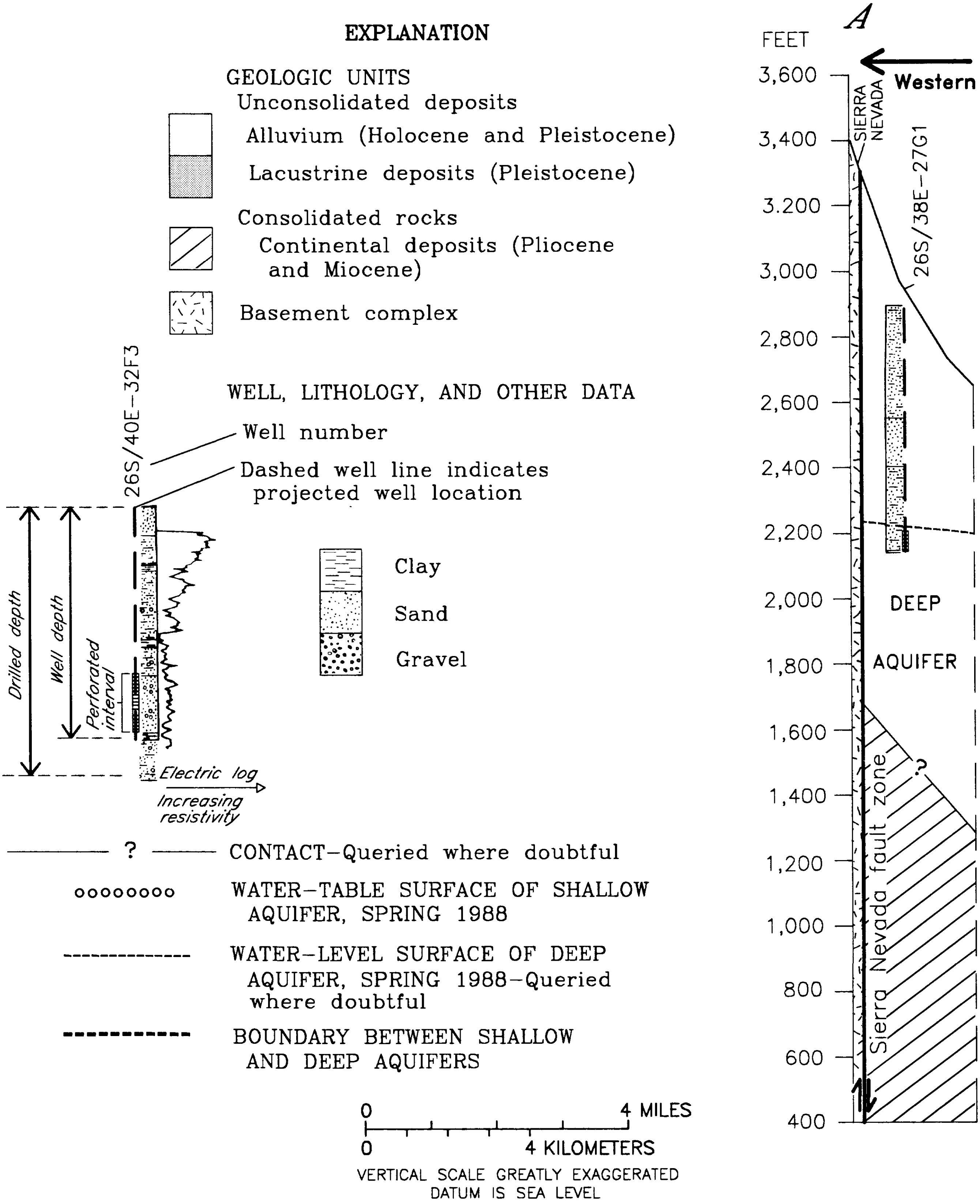
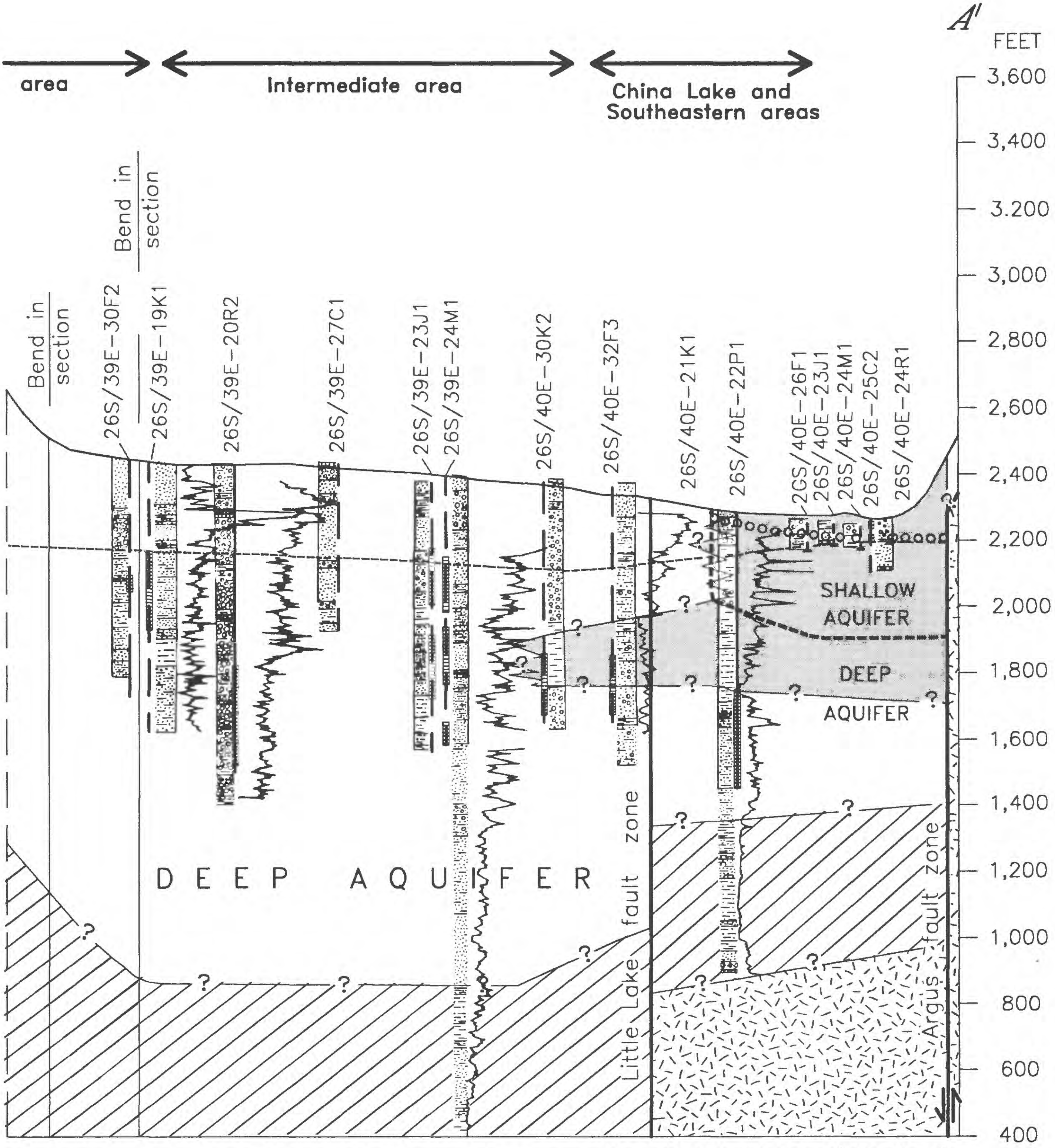


Figure 17. Geologic section of Indian Wells Valley ground-water basin. From Berenbrock and Martin, 1991, figure 4. (Line of section shown in figure 2 and on plate 1. Location of wells is shown on plate 1.)



From Kunkel and Chase (1969, fig. 3); modified by Berenbrock and Martin (1991)

Figure 17. *Continued.*

Table 5. Selected chemical and isotopic data for water samples from wells along section *A-A′* in Indian Wells Valley

[Reproduced from data in Supplemental Data B and from files in the San Diego Office of U.S. Geological Survey. Wells are listed in west-to-east order along section *A-A′* (fig. 16). Altitude of land surface is given in feet above sea level; depth of well and perforated interval are given in feet below land surface. µS/cm; microsiemens per centimeter at 25 °C; mg/L, milligrams per liter; --, no data available]

Well number	Altitude of land surface	Depth of well	Perforated interval	Specific conductance (µS/cm)	pH (standard units)	Percent sodium	Sulfate (mg/L)	Chloride (mg/L)	Delta deuterium (permil)
26S/38E-27G1	2,900.68	723	663–723	440	6.3	63	130	16	-93.0
26S/39E-19K1	2,406.2	803	270–540	780	8.1	42	80	136	[1]-94.5
26S/39E-30J1	2,441.1	430	294–298; 306–325; 330–343; 360–370; 393–413	430	8.3	56.3	58	24	--
26S/39E-20R2	2,421	920	600–900	365	8.2	59	40	27	--
26S/39E-27C1	2,425	500	--	400	8.02	43.0	50.9	30.2	--
26S/39E-26B4	2,384	1,030	410–490; 540–560; 735–1,015	300	9.2	80	25	16	--
26S/39E-23J1	2,361.2	800	210–380; 470–660	445	8.0	46	49	29	--
26S/39E-24M1	2,366.46	800	220–405; 450–620; 730–800	332	8.1	44	42	25	-93.0
26S/39E-25E1	2,345	387	179–260; 268–284; 391–384	450	7.9	40	47	32	-96.0
26S/40E-30E2	2,345	378	205–378	410	8.1	38	39	27	-95.5
26S/40E-30K1	2,340	800	250–800	380	8.28	80.2	22.0	28.7	-102.0
26S/40E-30K2	2,340	802	220–470; 600–760	375	8.9	73	27	36	-99.5
26S/40E-32F3	2,320	720	520–700	775	9.1	97	28	110	-105.0
26S/40E-28J1	2,288.9	--	--	870	8.4	43	270	55.6	-96.0
26S/40E-22P1	2,258.7	830	530–830	1,830	8.6	99	12	150	--
26S/40E-22P2	2,262.8	75	73–75	2,180	7.2	39	960	96	-97.0
26S/40E-22P3	2,260	415	400–415	2,230	8.8	99	11	230	-102.0
26S/40E-22P4	2,260	215	200–215	2,250	7.3	28	1,100	140	-96.0
26S/40E-25C2	2,255	160	20–40 60-80	1,710	8.0	69	240	280	--
26S/40E-26F1	2,225	77	75–77	2,150	7.7	56	490	270	--

[1] Isotopic data from nearby well 26S/39E-19P1.

far east end of the section (fig. 16) as land-surface altitude begins to rise (for example, in well 26S/40E-25C2).

Water-quality variations along this section provide an indication, along with the previous discussions, of the kind, and location, of additional data that would be useful in better defining the cause of chemical and isotopic variations and estimating isotopic differences between the late Pleistocene and the present.

1. Chemical-quality data from very deep wells (just above the continental deposits) in the playas, especially beneath Mirror and Satellite Lakes, would aid in determining the depth to which fine-grained deposits and high-DSC ground waters extend.
2. Additional wells at shallow and moderate depths in and west (downgradient) of Mirror and Satellite Lakes would define more precisely the zones in which extensive sulfate reduction has or has not occurred.
3. Depth-related chemical and isotopic data, in addition to geochemical processes described in

this report, could be used to model solute transport and chemical changes if suitable wells were available along flow paths upgradient of wells that have shown recent increases in DSC.

4. More precisely delineated depth-related water-quality data and accompanying lithologic data (including solid-phase chemical, isotopic, and mineralogic analyses) are needed from central parts of the valley— especially where advance and retreat of the ancestral China Lake has resulted in interfingering of coarse-grained alluvial and fine-grained lacustrine deposits.
5. Data are needed from places where few or no wells currently exist. One example is the northeastern area, where data that would characterize recharge in and north of Airport Lake are unavailable. Another is in and near the El Paso Mountains, where recharge may have occurred during wetter and colder climates.
6. Chemical and isotopic data from the continental deposits could more fully characterize differences between ground water in these deposits and ground water in the alluvial and lacustrine deposits.
7. Isotopic data from the base of the alluvium close to the Sierra Nevada could place a minimum constraint on δD values in recently recharged water and perhaps more definitively distinguish current from historical (Pleistocene) recharge.
8. Geochemical and isotopic data similar to those collected for this study could be obtained from other small closed or semiclosed Pleistocene basins in the Western United States to determine the extent to which similar processes of evaporative concentration, evaporite dissolution, ion exchange, and microbial sulfate reduction occur; and to establish whether reliable isotopic maps of late Pleistocene recharge can be constructed using ground-water data. Buchanan (1989) has suggested similar use of isotopic (and chemical) data to distinguish "contemporary high-altitude" from "paleo-fluid" recharge in geothermal systems within the Great Basin. The procurement of detailed isotopic data sets from all small basins in the Western Unites States would enhance interpretations made on a large regional scale such as those of Ingraham and Taylor (1991) and Smith and others (1992).

SUMMARY AND CONCLUSIONS

Ground water is virtually the sole source of water for residents of Indian Wells Valley. This report presents the results of a study that updates an appraisal that was done by the U.S. Geological Survey in 1972. In the 1972 study, it was found that pumping had not yet degraded water quality, but more recent data indicate that dissolved-solids concentration (DSC) has increased since 1972 in several wells. These increases were caused by pumpage-induced flow from nearby lacustrine (and playa) deposits that contain saline ground water.

For this study, major ions and selected minor constituents were analyzed in samples from more than 80 wells, and stable hydrogen and oxygen isotopes were analyzed in samples from 55 wells. Variability in location and timing of recharge, evapotranspiration on the valley floor, and several geochemical processes in the aquifer produce a very wide range in chemical and isotopic composition of the ground water. The ranges observed were about 200 to 60,000 mg/L for DSC, nearly zero to almost half (of DSC) in the proportion of some individual ions, and -108 to -76 permil for delta deuterium (δD).

Natural recharge to the ground-water system occurs as surface runoff from the surrounding mountains that infiltrates alluvial deposits on the margins of the valley. Natural discharge is by evapotranspiration on the valley floor in and around China Lake. Recharge originates predominantly in the Sierra Nevada on the west side of the valley; smaller amounts originate in the Coso and Argus Ranges on the north and east sides, and less than 5 percent originates in the El Paso Mountains on the south side. As ground water moves from alluvium into the lacustrine deposits of the ancestral China Lake, its DSC increases from a few hundred to more than 1,000 mg/L. Much larger increases in DSC occur in deposits near the China Lake playa (especially in the shallow deposits).

In addition to a general trend of increasing DSC as ground water moves through the valley, sulfate (likely microbial) reduction, carbonate precipitation, ion exchange, and evaporite dissolution in lacustrine deposits exert a large influence on the concentration and proportion of major ions. Comparison of depth-related concentrations at one location beneath Mirror Lake shows the relative importance of these processes. Stoichiometric calculations indicate that ground water from depths greater than 400 ft below land surface has the following incremental difference from ground water at depths of less than 200 ft: 21 meq/L sulfate reduction accompanied by generation of carbonate alkalinity, 7 meq/L precipitation of alkaline-earth carbonates, 9 meq/L ion exchange of aqueous alkaline-earth ions for sodium ions on clays, and 3 meq/L halite dissolution. The net result of these processes is that the predominant ions in deep ground water at this location are sodium (greater than 99 percent of total cations), bicarbonate, and chloride; and the deep ground water is poorly buffered (pH is about 9).

Deposits in which sulfate reduction is intensive seem to extend upward to shallower depths in China Lake than in the much smaller Mirror and Satellite Lakes, where less reducing conditions persist to greater depth. Accordingly, at those wells that exhibit a temporal trend of increasing DSC, induced flow from China Lake might be expected to cause an increase in the chloride-to-sulfate ratio, and induced flow from Mirror and Satellite Lakes might be expected to cause an increase in the sulfate-to-chloride ratio. There is some evidence for the former change in the intermediate area and more evidence for the latter change at Ridgecrest, but DSC has increased at only a few wells and the change in relative concentrations still is small in those wells.

Delta deuterium in deep ground water from alluvial deposits on the west side of the valley ranges from about -90 permil at the south end to about -105 permil at the north end. Enrichment in deuterium toward the south probably results from decreasing altitude of the Sierra Nevada in this direction. The range in isotope ratios is not as large in current recharge from the Coso and Argus Ranges, where δD ranges from about -90 to about -100 permil, as it is in current recharge from the Sierra Nevada.

Isotope ratios generally are more negative in ground water from greater depth, especially in areas of the valley farthest from the sources of recharge. This pattern may result partly because this ground water is recharged at higher altitudes and is less affected by evaporation, but it also seems to indicate recharge during an earlier (presumably late Pleistocene) pluvial period. A local ground-water line, estimated from the relation between hydrogen- and oxygen-isotope ratios in ground-water samples minimally affected (that is, having isotopic composition closest to the global meteoric-water line) by evaporation was found to be nearly parallel, and about 4 permil more negative on the δD axis, to the global meteoric-water line defined by $\delta D = 8\delta^{18}O + 10$. An evaporation line with a slope of about 5.8 was estimated from the isotope ratios in samples of ground water that had been most affected by evaporation at some time in its history as indicated by maximum deviation from the meteoric-water line. These two lines intersect at δD about -125 permil, which is 15 to 35 permil more negative than current recharge.

An advective transport model, MODPATH, was used to simulate the effects of pumping on ground-water flow paths and traveltimes in Indian Wells Valley. Historical water-level data and the MODPATH simulations indicate that pumping in the intermediate area has reversed the hydraulic gradient between China Lake and the intermediate area, and that pumping at Ridgecrest has induced flow from Mirror and Satellite Lakes. Simulation of advective transport of ground water through alluvium from Little Dixie Wash, a few miles west of pumped wells in the intermediate area, suggests that pumping has significantly increased the velocity of ground-water movement but that traveltimes from Little Dixie Wash to the pumped wells still may be 1,000 years or more.

Traveltimes as great as several thousand years were obtained between other parts of the valley; where flow paths pass through lacustrine deposits, traveltimes increase markedly. Such long traveltimes give additional credibility to the isotopic evidence for existence of late Pleistocene ground water in the eastern and central parts of Indian Wells Valley. The traveltimes also demonstrate that the recently observed increase in DSC at several wells between the area of heavy pumping and the playas originates either from increased flow through permeable strata not accounted for in the simplified model or from local sources (lacustrine deposits) near the wells.

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SUPPLEMENTAL DATA

Supplemental Data A. Altitude of water level for selected wells in Indian Wells Valley, spring 1988

[Altitudes are given in feet above sea level; depth to water in feet below land surface. Winter or summer data are given for those wells for which spring data are not available]

Well number	Altitude of land surface	Date	Depth to water	Altitude of water level	Monitored aquifer
24S/38E-16J2	2,585	03/29/88	318.80	2,266.20	Deep
24S/38E-33J2	2,480	03/29/88	274.98	2,205.02	Deep
25S/38E-13L1	2,320	03/29/88	135.43	2,184.57	Deep
25S/38E-23G1	2,412	03/19/88	233.72	2,188.28	Deep
25S/38E-25L1	2,329.2	03/19/88	142.93	2,186.27	Deep
25S/39E-31R1	2,260	01/12/88	73.22	2,186.78	Deep
26S/38E-27G1	2,900.68	01/12/88	659.50	2,241.18	Deep
26S/38E-7N2	2,395.3	03/19/88	210.43	2,184.87	Deep
26S/39E-8E1	2,318	03/08/88	134.50	2,183.50	Deep
26S/39E-17F2	2,340	01/12/88	171.98	2,168.02	Deep
26S/39E-19Q1	2,418.3	03/19/88	232.65	2,185.65	Deep
26S/39E-24K1	2,347.4	03/19/88	202.59	2,144.81	Deep
26S/39E-30F1	2,433.5	03/19/88	248.04	2,185.46	Deep
26S/40E-4Q1	2,185	03/02/88	12.67	2,172.33	Shallow
26S/40E-6D1	2,216	03/02/88	33.67	2,182.33	Shallow
26S/40E-6D2	2,216	03/02/88	34.42	2,181.58	Shallow
26S/40E-9A1	2,187	03/02/88	14.72	2,172.28	Shallow
26S/40E-11J3	2,174	03/18/88	3.37	2,170.63	Shallow
26S/40E-14B1	2,186.5	03/18/88	1.00	2,185.5	Shallow
26S/40E-14L1	2,201	03/18/88	5.18	2,195.82	Shallow
26S/40E-19P1	2,336	03/29/88	192.98	2,143.02	Shallow
26S/40E-22P1	2,258.7	01/11/88	103.92	2,154.78	Shallow
26S/40E-22P3	2,260	01/11/88	96.34	2,163.66	Shallow
26S/40E-23A3	2,210	01/12/88	24.59	2,185.41	Shallow
26S/40E-23A4	2,210	01/12/88	20.08	2,189.92	Shallow
26S/40E-23G1	2,215.04	07/07/88	23.78	2,191.26	Shallow
26S/40E-32F3	2,320	01/11/88	184.18	2,135.82	Shallow
26S/40E-32F4	2,320	01/11/88	185.84	2,135.16	Shallow
26S/40E-34H1	2,243	01/12/88	67.92	2,175.08	Shallow
26S/40E-35H2	2,243	01/12/88	72.14	2,170.86	Shallow
27S/39E-8A1	2,535	01/14/88	342.46	2,192.54	Shallow
27S/39E-16C1	2,582.3	06/09/88	389.98	2,192.32	Shallow
27S/40E-4A1	2,305	01/11/88	146.67	2,158.33	Shallow
27S/40E-6D2	2,400	01/11/88	271.68	2,128.32	Shallow

Supplemental Data B. Chemical analyses of ground water in Indian Wells Valley, 1987-88

[Altitude of land surface is given in feet above sea level. Depth of well and perforated interval are given in feet below chloride, nitrate, sulfate, and silica (Fishman and Friedman, 1989, p. 459-460). μS/cm, microsiemens per centimeter at --, no data available.]

Well number	Date	Time	Altitude of land surface	Depth of well, total	Perforated interval	Temperature (°C)	Specific conductance (μS/cm)	pH (standard units)
			Northwestern area					
25S/38E-11L1	09-20-87	1300	2,445	400	--	20.5	500	8.0
	07-09-88	1325				26.5	500	8.0
25S/38E-13J1	07-06-88	1430	2,295	154.5	--	27.0	570	8.2
25S/39E-4R1[1]	[2]06----87	--	2,252.6	200	100-200	--	1,300	7.6
	07-08-88	1105				26.0	1,220	7.6
25S/39E-9J1[1]	01-19-89	1445	2,248.1	200	--	24.0	930	7.7
			Western area					
25S/38E-25J1	[3]08-25-88	--	2,275	330	120-330	--	1,400	7.7
25S/38E-36A1	[3]08-25-88	--	2,291	285	139-285	--	4,400	7.5
25S/38E-36B1	[3]07-29-88	--	2,293	400	200-400	--	1,160	7.5
25S/39E-31D1	[3]07-29-88	--	2,267	300	140-300	--	1,280	7.5
25S/39E-31R1	05-31-87	1125	2,260	480	120-180	22.0	3,820	9.4
26S/39E-5F1[1]	07-08-88	1205	2,276.7	200	100-200	24.0	850	7.4
26S/39E-8E1	05-31-87	1540	2,318	880	570-880	26.0	1,350	9.0
26S/39E-17F2	05-31-87	1500	2,340	881	681-881	22.0	360	8.8
26S/39E-18K2	07-01-88	0850	2,388	310	290-310	28.5	530	7.3
26S/39E-19K1	[2]04-01-87	0900	2,406.2	803	270-540	--	780	8.1
26S/39E-30J1	[4]05-14-87	1255	2,441.1	430	294-298; 306-325; 330-343; 360-370; 393-413	--	430	8.30
	07-07-88	1115				30.5	445	8.1
			Southwestern area					
26S/37E-26L1	07-23-87	1500	4,320	50	--	27.5	400	7.2
27S/39E-7R1	[3]08-31-88	1200	2,563.3	515	434-514	--	440	7.75
27S/39E-8M1	07-02-87	1200	2,558	--	(5)	32.0	420	8.1
	[3]07-02-87	1600		--	(6)	29.0	430	7.96
	[3]07-06-87	1055		--	(7)	27.4	400	7.98
	[3]07-06-87	1315		--	(8)	28.5	430	7.77
	[3]07-06-87	1635		--	(9)	30.1	430	8.02
	[3]07-06-87	1755		--	(10)	29.7	420	7.7
			China Lake area					
25S/39E-12R2[1]	[2]04-28-87	0900	2,196.1	150	65-140	--	1,240	7.8
	07-08-88	1030				24.5	1,150	7.7
26S/40E-1A2	06-01-87	1920	2,157.6	197.5	80-100; 110-130; 170-190	24.0	18,500	--
26S/40E-4Q1	05-30-87	0930	2,185	290	30-50; 70-90; 110-130	21.0	875	9.0
26S/40E-6C1	05-30-87	1845	2,195	620	500-600	20.5	90,000	10.1
26S/40E-6D1	05-30-87	1800	2,216	320	276-300	23.5	16,000	10.1
26S/40E-6D2	05-30-87	1820	2,216	260	120-200	24.0	14,500	9.7
26S/40E-9A1	05-29-87	1810	2,187	100	20-40; 60-80	23.0	950	8.0
26S/40E-11J3	07-01-88	1530	2,174	7.6	--	27.5	11,600	7.5
26S/40E-14B1[1]	05-27-87	1250	2,186.5	22	20-22	23.5	6,400	7.9
	07-06-88	1800				22.0	4,770	7.5

Footnotes at end of table.

land surface. Calculated by adding the following constituents: calcium, magnesium, sodium, potassium, alkalinity, 25°C; °C, degrees Celsius; mg/L, milligrams per liter; µg/L, micrograms per liter; <, actual value less than value shown;

Hardness (mg/L, as $CaCO_3$)	Hardness, noncarbonate (mg/L, as $CaCO_3$)	Calcium, dissolved (mg/L, as Ca)	Magnesium, dissolved (mg/L, as Mg)	Sodium, dissolved (mg/L, as Na)	Percent sodium	Sodium adsorption ratio	Potassium, dissolved (mg/L, as K)	Alkalinity, field (mg/L, as $CaCO_3$)	Alkalinity, lab (mg/L, as $CaCO_3$)
				Northwestern area--*Continued*					
94	0	24	8.3	61	57	3	4.8	--	193
100	0	26	8.4	74	60	3	4.9	184	194
140	0	33	13	70	52	3	5.8	179	176
275	--	50	37	166	55	4	16	--	370
290	0	52	39	170	55	4	11	345	356
230	0	46	29	110	49	3	9.6	321	336
				Western area--*Continued*					
162	--	40	15	253	77	9	5.4	--	365
1,271	--	309	121	486	45	6	7.7	--	372
214	--	61	15	183	64	5	4.9	--	325
258	--	72	19	185	60	5	4.4	--	315
150	0	41	11	710	95	25	6.1	--	684
210	37	59	14	120	55	4	3.7	168	164
5	0	1.5	.28	280	99	55	2.8	544	557
15	0	4.9	.64	52	85	6	3.7	106	118
120	32	38	6.0	51	48	2	2.3	88	91
185	--	60	9	64	42	2	4	--	60
92	--	29.4	3.4	54.3	56.3	2.5	3.4	103	--
98	0	31	4.9	55	54	3	2.3	108	110
				Southwestern area--*Continued*					
--	--	--	--	--	--	--	--	123	123
108.4	--	32.2	6.8	50.7	49.7	2.2	2.7	--	110.9
97	5	30	5.3	49	51	2	3.4	92	101
91.2	--	21.6	3.0	52.6	62.2	2.8	2.7	--	110.5
99.6	--	32.4	4.6	48.7	50.8	2.1	2.3	--	107.1
103.2	--	34.9	3.9	45.8	48.5	2.0	2.2	--	109.2
103.6	--	34.9	4.0	47.4	49.2	2.0	2.4	--	109.7
117.3	--	40.7	3.4	41.3	43.1	1.7	2.6	--	106.6
				China Lake area--*Continued*					
205	--	40	25	177	63	5	20	--	305
220	0	43	27	180	63	5	11	298	303
5	0	1.7	.3	4,300	99	830	42	--	4,180
68	0	20	4.3	110	76	6	4.7	196	110
11	0	3.1	.75	24,000	100	3,300	74	--	28,200
12	0	2.9	1.2	4,000	100	510	10	--	4,810
12	0	3.1	1.0	3,500	100	460	17	3,360	4,100
130	0	41	6.7	110	64	4	4.6	182	186
400	76	87	44	2,800	93	63	32	323	343
860	550	250	57	890	69	14	19	307	320
660	370	190	45	830	73	15	15	290	309

Supplemental Data B. Chemical analyses of ground water in Indian Wells Valley, 1987-88--*Continued*

Well number	Sulfate, dissolved (mg/L, as SO_4)	Chloride, dissolved (mg/L, as Cl)	Fluoride, dissolved (mg/L, as F)	Silica, dissolved (mg/L, as SiO_2)	Solids, residue at 180 °C, dissolved (mg/L)	Solids, sum of constituents, dissolved (mg/L)	Nitrogen, NO_2+NO_3 dissolved (mg/L, as N)	Nitrogen ammonia, dissolved (mg/L, as N)	Nitrogen ammonia + organic, dissolved (mg/L, as N)
Northwestern area--*Continued*									
25S/38E-11L1	41	12	1.4	24	--	296	0.83	--	--
	40	11	1.3	26	--	306	.89	0.010	<0.20
25S/38E-13J1	62	24	.40	22	--	343	.68	1.90	1.7
25S/39E-4R1[1]	105	118	.6	--	764	715	--	--	--
	150	110	.70	48	--	792	.37	.030	.30
25S/39E-9J1[1]	88	59	.70	47	--	584	.16	.020	.20
Western area--*Continued*									
25S/38E-25J1	114	163	1.4	--	875	812	--	--	--
25S/38E-36A1	1,220	310	.63	--	3,210	2,992	--	--	--
25S/38E-36B1	141	95.9	.73	--	755	702	--	--	--
25S/39E-31D1	185	106	.97	--	845	767	--	--	--
25S/39E-31R1	110	550	2.1	35	--	1,890	.14	--	--
26S/39E-5F1[1]	180	75	.60	40	--	594	<.10	.020	.20
26S/39E-8E1	11	58	9.5	29	--	724	<.10	--	--
26S/39E-17F2	6.7	7.2	3.2	30	--	173	<.10	--	--
26S/39E-18K2	63	55	.6	35	--	312	1.50	.020	.70
26S/39E-19K1	80	136	.7	--	393	393	--	--	--
26S/39E-30J1	58	24	.9	--	252.8	243	--	--	--
	54	27	.6	36	--	283	1.70	.030	.20
Southwestern area--*Continued*									
26S/37E-26L1	--	--	--	--	--	--	--	--	--
27S/39E-7R1	57.5	25.4	.7	--	258.8	225	--	--	--
27S/39E-8M1	56	32	1.2	10	274	263	1.80	.90	.60
	45.6	25.5	1.0	--	246.4	225	--	--	--
	47.3	26.0	.9	--	253.8	234	--	--	--
	46.5	25.7	.9	--	262.3	231	--	--	--
	47.2	25.9	.9	--	252.0	234	--	--	--
	50.1	28.3	.8	--	260.8	235	--	--	--
China Lake area--*Continued*									
25S/39E-12R2[1]	129	117	.8	--	692	692	--	--	--
	140	120	.80	52	--	760	.47	<.010	.50
26S/40E-1A2	950	3,700	3.7	69	--	11,700	<1.0	--	--
26S/40E-4Q1	52	120	.7	49	--	481	.23	--	--
26S/40E-6C1	5,200	14,000	150	48	--	60,700	<.10	--	--
26S/40E-6D1	120	2,700	11	34	--	9,850	<.10	--	--
26S/40E-6D2	210	2,800	10	22	--	8,710	<.10	--	--
26S/40E-9A1	76	130	.7	54	--	535	.20	--	--
26S/40E-11J3	1,000	3,800	2.7	110	--	8,090	.12	.030	.70
26S/40E-14B1[1]	710	1,300	1.2	69	--	3,500	3.10	--	--
	580	1,000	1.1	70	--	2,930	3.10	.060	.60

Footnotes at end of table.

Nitrogen, organic, dissolved (mg/L, as N)	Nitrate, total (mg/L as NO_3)	Phosphorus, dissolved (mg/L, as P)	Boron, dissolved (µg/L, as B)	Iron, dissolved (µg/L, as Fe)	Manganese, dissolved (µg/L, as Mn)	Predominant ions	Monitored aquifer	Well use
				Northwestern area--*Continued*				
--	--	--	380	150	2	Na HCO_3	Deep	Domestic
--	--	0.050	510	<3	1	Na HCO_3	Deep	
0	--	.010	230	3	110	Na HCO_3	Deep	Observation
--	1	--	--	--	--	Na HCO_3	Deep	Military
.27	--	.020	2,800	<3	50	Na HCO_3		
.18	--	.020	1,400	150	75	Na HCO_3	Deep	Military
				Western area--*Continued*				
--	2.2	--	--	<50	<10	Na HCO_3	Deep	Agricultural
--	315	--	--	<50	<10	Na Ca SO_4	Deep	Agricultural
--	6.2	--	--	80	50	Na HCO_3	Deep	Agricultural
--	6.2	--	--	70	10	Na HCO_3 SO_4 Cl	Deep	Agricultural
--	--	--	18,000	40	30	Na Cl HCO_3	Deep	Observation
.18	--	.010	510	260	65	Na SO_4 HCO_3 Cl	Deep	Military
--	--	--	5,400	40	5	Na HCO_3	Deep	Observation
--	--	--	600	23	7	Na HCO_3	Deep	Observation
.68	--	.020	160	1,400	18	Na Ca HCO3 Cl SO_4	Deep	Domestic
--	4	--	--	--	--	Ca Na Cl	Deep	Military
--	8.9	--	--	20	<20	Na HCO_3	Deep	Municipal
.17	--	.010	200	<3	<1	Na HCO_3		
				Southwestern area--*Continued*				
--	--	--	--	--	--	--	Deep	Domestic
--	7.6	--	--	<100	40	Na Ca HCO_3	Deep	Observation
.51	--	.040	170	150	180	Na HCO_3 SO_4	Deep	Observation
--	7.9	--	--	410	80	Na HCO_3		
--	8.5	--	--	400	30	Na HCO_3		
--	6.3	--	--	430	60	Na Ca HCO_3		
--	6.6	--	--	300	30	Na Ca HCO_3		
--	5.6	--	--	40	110	Ca Na HCO_3		
				China Lake area--*Continued*				
--	1	--	--	600	<30	Na HCO_3	Shallow	Military
--	--	.020	2,800	<3	7	Na HCO_3 SO_4 Cl		
--	--	--	170,000	740	20	Na Cl	Shallow	Observation
--	--	--	1,700	12	6	Na Cl	Shallow	Observation
--	--	--	270,000	620	18	Na HCO_3	Deep	Observation
--	--	--	84,000	80	20	Na HCO_3	Deep	Observation
--	--	--	130,000	140	20	Na Cl HCO_3	Shallow	Observation
--	--	--	2,100	16	57	Na Cl HCO_3	Shallow	Observation
.67	--	.060	21,000	30	10	Na Cl	Shallow	Observation
--	--	--	7,900	60	100	Na Cl	Shallow	Observation
.54	--	.060	6,300	10	50	Na Cl	Shallow	Observation

Supplemental Data B. Chemical analyses of ground water in Indian Wells Valley, 1987-88--*Continued*

Well number	Date	Time	Altitude of land surface	Depth of well, total	Perforated interval	Temperature (°C)	Specific conductance (μS/cm)	pH (standard units)
			China Lake area--*Continued*					
26S/40E-14L1[1]	06-30-88	1815	2,201	57	55-57	24.0	2,200	7.3
26S/40E-15N2	07-09-88	0945	2,234.83	101	99-101	26.0	3,950	8.6
26S/40E-17J1[1]	07-02-88	1050	2,262.35	97	95-97	25.0	415	7.5
26S/40E-17R1[1]	07-02-88	0930	2,266.65	101	99-101	24.5	450	7.8
26S/40E-22A1	05-31-87	0900	2,227	153	35-75	22.0	9,400	7.5
	07-06-88	1650				24.5	6,100	7.5
26S/40E-22B1[1]	06-02-87	1140	2,232.45	63	61-63	25.0	7,200	7.7
	07-07-88	0930				26.5	5,900	7.9
26S/40E-22H1[1]	06-29-88	1745	2,226.62	49	47-49	27.0	4,800	7.2
26S/40E-22H2[1]	06-29-88	1850	2,227.03	77	75-77	26.0	11,000	7.2
26S/40E-22H3[1]	06-02-87	1100	2,226.23	97	95-97	26.0	9,200	7.4
	07-07-88	0900				27.0	6,000	7.5
26S/40E-22J1[1]	07-01-88	1430	2,236.88	71	69-71	28.0	3,100	7.1
26S/40E-22P1[1]	05-27-87	1830	2,258.7	830	530-830	31.0	1,830	8.6
	08-09-88	1730				32.0	1,880	8.6
26S/40E-22P2[1]	07-01-88	1350	2,262.8	75	73-75	28.5	2,180	7.2
26S/40E-22P3	05-26-87	1740	2,260	415	400-415	27.0	2,230	8.8
	08-08-88	1710				27.5	2,220	8.8
26S/40E-22P4	05-26-87	1635	2,260	215	200-215	25.0	2,250	7.3
	08-09-88	1040				25.5	1,670	7.4
26S/40E-23B2	05-27-87	1035	2,210	360	300-340	24.5	2,180	8.7
	01-18-89	1630				23.5	2,000	8.8
26S/40E-23B3	05-27-87	1250	2,210	240	180-220	24.0	2,150	8.8
	07-07-88	1340				25.0	2,150	8.8
26S/40E-23D1	05-26-87	2005	2,223	400	385-400	24.5	2,900	9.0
26S/40E-23D2	05-26-87	2010	2,223	185	170-185	23.0	8,500	8.3
26S/40E-23G1[1]	07-07-88	1700	2,213.04	57	55-57	24.0	8,550	7.9
			Southeastern area					
26S/40E-25C2	05-28-87	1835	2,255	160	20-40; 60-80	24.0	1,710	8.0
26S/40E-26F1[1]	07-08-88	1710	2,225	77	75-77	25.5	2,150	7.7
26S/40E-27D1	07-08-88	1420	2,266.71	160	100-140	26.5	2,350	7.9
26S/40E-27D2	07-08-88	1400	2,264	--	--	25.5	2,150	7.9
26S/40E-27E3	07-08-88	1345	2,269.06	480	400-460	26.5	640	9.1
26S/40E-28J1	[13]09----87	--	2,288.9	--	--	--	870	8.4
26S/40E-32F3	05-28-87	1500	2,320	720	520-700	33.5	775	9.1
	[3]01-07-88	1345				--	880	8.27
26S/40E-32K1	[3]01-07-88	1330	2,330	620	230-310; 340-380; 470-500; 520-600	--	420	8.27
26S/40E-33P4	05-28-87	1610	2,300	304	169-182; 198-216; 233-252; 256-272; 278-290	27.0	660	8.7
	01-17-89	1900				26.0	580	8.7
26S/40E-34N1	01-18-89	1030	2,290.4	232	135-142; 146-155; 176-181	24.0	730	7.9
26S/40E-35Q2[1]	01-19-89	1030	2,251.47	127	125-127	21.5	1,500	7.9

Footnotes at end of table.

Hardness (mg/L, as $CaCO_3$)	Hardness, noncarbonate (mg/L, as $CaCO_3$)	Calcium, dissolved (mg/L, as Ca)	Magnesium, dissolved (mg/L, as Mg)	Sodium, dissolved (mg/L, as Na)	Percent sodium	Sodium adsorption ratio	Potassium, dissolved (mg/L, as K)	Alkalinity, field (mg/L, as $CaCO_3$)	Alkalinity, lab (mg/L, as $CaCO_3$)
China Lake area--*Continued*									
280	0	73	23	390	75	11	10	404	420
32	0	5.3	4.6	1,200	98	94	15	925	939
100	0	30	6.0	43	47	2	4.2	121	122
110	0	32	6.3	42	45	2	4.8	125	132
2,400	2,200	420	320	1,500	57	14	59	162	170
2,100	1,900	310	320	960	49	9	45	164	171
920	640	120	150	1,300	74	19	41	282	287
920	630	88	170	1,300	75	19	36	290	300
2,400	2,300	500	290	540	32	5	53	138	146
3,900	3,600	440	670	2,400	57	17	49	235	247
3,000	2,700	450	450	1,400	50	11	45	272	276
2,700	2,400	190	530	1,500	55	13	40	270	266
1,000	770	170	140	370	44	5	29	230	237
6	0	1.7	.53	410	99	73	4.4	741	771
6	0	1.7	.52	440	99	78	4.1	738	764
690	570	150	76	210	39	4	20	119	115
8	0	2.1	.70	520	99	82	7.1	880	880
5	0	1.2	.60	650	99	120	7.1	874	898
940	850	270	65	170	28	3	20	93	97
590	490	170	41	130	32	2	11	101	111
12	0	2.7	1.2	440	97	58	13	487	477
15	0	3.4	1.7	470	97	53	12	482	471
12	0	2.7	1.2	430	97	56	12	760	542
16	0	3.9	1.5	470	97	53	10	558	570
15	0	3.2	1.8	720	98	82	11	1,030	724
490	200	56	86	1,600	87	32	38	295	231
570	140	87	86	2,000	88	37	31	434	442
Southeastern area--*Continued*									
220	120	60	16	230	69	7	9.0	98	120
440	270	110	40	270	56	6	13	173	175
1,400	1,300	140	70	220	45	4	32	109	91
560	430	150	46	230	46	4	22	133	127
--	--	--	--	--	--	--	--	195	--
212	--	60	15	83	43	3	26	--	73
9	0	3.5	.06	150	97	23	.8	165	171
28.8	--	10.7	.5	183.0	92.8	14.8	2.0	--	71.3
27.6	--	7.3	2.2	82.7	86.1	6.9	1.4	--	110.9
25	0	7.7	1.4	110	89	10	2.7	86	105
31	0	9.6	1.8	110	87	9	3.1	87	101
150	57	39	12	82	53	3	6.9	90	92
220	120	62	16	210	66	6	15	100	108

Supplemental Data B. Chemical analyses of ground water in Indian Wells Valley, 1987-88--*Continued*

Well number	Sulfate, dissolved (mg/L, as SO_4)	Chloride, dissolved (mg/L, as Cl)	Fluoride, dissolved (mg/L, as F)	Silica, dissolved (mg/L, as SiO_2)	Solids, residue at 180 °C, dissolved (mg/L)	Solids, sum of constituents, dissolved (mg/L)	Nitrogen, NO_2+NO_3 dissolved (mg/L, as N)	Nitrogen ammonia, dissolved (mg/L, as N)	Nitrogen ammonia + organic, dissolved (mg/L, as N)
China Lake area--*Continued*									
26S/40E-14L1[1]	250	340	1.2	71	--	1,420	3.00	0.020	1.3
26S/40E-15N2	1,100	160	1.5	56	--	3,170	<.10	.010	3.3
26S/40E-17J1[1]	40	24	.8	45	--	266	<.10	<.010	.60
26S/40E-17R1[1]	41	28	.7	52	--	282	<.10	.030	.70
26S/40E-22A1	5,200	400	1.1	68	--	8,080	1.20	--	--
	3,700	310	1.0	70	--	5,830	2.10	.120	.70
26S/40E-22B1[1]	3,200	210	1.2	68	--	5,290	3.70	--	--
	3,200	210	1.1	72	--	5,280	3.20	.100	.60
26S/40E-22H1[1]	3,100	210	2.0	83	--	4,870	1.20	.100	1.0
26S/40E-22H2[1]	8,100	430	2.7	46	--	12,300	1.40	.150	1.0
26S/40E-22H3[1]	5,500	260	2.1	69	--	8,350	<.10	--	--
	5,100	210	1.9	73	--	7,820	.10	1.90	1.8
26S/40E-22J1[1]	1,200	210	2.3	69	--	2,330	1.00	.080	.80
26S/40E-22P1[1]	12	150	3.5	41	--	1,070	<.10	--	--
	13	140	2.9	39	--	1,090	<.10	.350	1.0
26S/40E-22P2[1]	960	96	1.8	77	--	1,660	<.10	.850	1.3
26S/40E-22P3	11	230	5.1	26	--	1,340	<.10	--	--
	8.6	220	4.6	39	--	1,470	<.10	1.10	1.8
26S/40E-22P4	1,100	140	.5	63	--	1,890	1.10	--	--
	670	110	.5	62	--	1,260	.90	.030	.40
26S/40E-23B2	42	340	2.6	54	--	1,190	<.10	--	--
	58	320	2.3	52	--	1,220	<.10	1.60	1.8
26S/40E-23B3	31	250	2.5	48	--	1,240	<.10	--	--
	38	240	2.2	22	--	1,140	<.10	9.90	2.2
26S/40E-23D1	330	370	6.0	14	--	2,090	<.10	--	--
26S/40E-23D2	2,300	1,100	1.7	46	--	5,420	<.10	--	--
26S/40E-23G1[1]	2,300	1,400	2.3	67	--	6,260	<.10	.910	1.2
Southeastern area--*Continued*									
26S/40E-25C2	240	280	1.2	56	--	956	.85	--	--
26S/40E-26F1[1]	490	270	.5	18	--	1,320	<.10	.290	.50
26S/40E-27D1	1,200	35	.8	90	--	1,940	<.10	1.20	1.2
26S/40E-27D2	810	120	.6	22	--	1,490	1.70	1.40	1.8
26S/40E-27E3	--	--	--	--	--	--	--	--	--
26S/40E-28J1	270	55.6	.38	--	595	554	--	--	--
26S/40E-32F3	28	110	.8	53	--	446	<.10	--	--
	48.9	221.8	.7	--	560.5	511	--	--	--
26S/40E-32K1	24.3	37.3	1.3	--	267.5	227	--	--	--
26S/40E-33P4	27	96	1.2	37	--	343	1.70	--	--
	29	100	1.0	37	--	353	1.90	<.010	.30
26S/40E-34N1	86	110	.6	46	--	442	1.10	.020	.60
26S/40E-35Q2[1]	43	420	.7	65	--	894	<.10	.240	.50

Footnotes at end of table.

Nitrogen, organic, dissolved (mg/L, as N)	Nitrate, total (mg/L as NO_3)	Phosphorus, dissolved (mg/L, as P)	Boron, dissolved (µg/L, as B)	Iron, dissolved (µg/L, as Fe)	Manganese, dissolved (µg/L, as Mn)	Predominant ions	Monitored aquifer	Well use
				China Lake area--*Continued*				
1.3	--	0.060	3,700	1,300	40	Na Cl HCO_3 SO_4	Shallow	Observation
3.3	--	.710	69,000	110	40	Na SO_4 HCO_3	Shallow	Observation
--	--	.010	200	220	31	Na Cl HCO_3	Shallow	Observation
.67	--	.020	200	21	29	Na Cl HCO_3	Shallow	Observation
--	--	--	12,000	50	160	Na SO_4	Shallow	Observation
.58	--	.020	5,100	30	40	Na SO_4		
--	--	--	10,000	200	170	Na SO_4	Shallow	Observation
.50	--	.060	10,000	220	160	Na SO_4		
.90	--	.050	2,600	120	260	Ca Mg Na SO_4	Shallow	Observation
.85	--	.050	11,000	520	290	Na SO_4	Shallow	Observation
--	--	--	11,000	990	460	Na SO_4	Shallow	Observation
0	--	.050	9,300	160	400	Na SO_4		
.72	--	.010	1,900	30	170	Na Mg Ca SO_4	Shallow	Observation
--	--	--	4,900	60	5	Na HCO_3	Deep	Observation
.65	--	.170	4,700	35	3	$NaHCO_3$		
.45	--	.010	510	180	81	Na Ca Mg SO_4	Shallow	Observation
--	--	--	15,000	110	20	Na HCO_3	Deep	Observation
.70	--	.310	14,000	100	<10	Na HCO_3		
--	--	--	710	90	130	Na Mg SO_4	Shallow	Observation
.37	--	.110	520	40	15	Ca Na Mg SO_4		
--	--	--	5,800	40	20	Na HCO_3 Cl	Deep	Observation
.20	--	.140	5,200	50	10	Na HCO_3 Cl		
--	--	.170	5,700	60	10	Na HCO_3	Shallow	Observation
0	--	.180	5,900	20	<10	Na HCO_3		
--	--	--	19,000	180	20	Na HCO_3 Cl SO_4	Shallow	Observation
--	--	.050	19,000	40	100	Na SO_4	Shallow	Observation
.29	--	.120	25,000	180	100	Na SO_4 Cl	Shallow	Observation
				Southeastern area--*Continued*				
--	--	--	1,300	<3	35	Na Cl	Deep	Observation
.21	--	.030	1,800	210	10	Na SO_4 Cl	Deep	Observation
0	--	.010	500	<10	130	Ca SO_4	Deep	Observation
.40	--	.020	660	20	100	Na Cl Mg SO_4	Deep	Observation
--	--	--	--	--	--	--	Deep	Observation
--	0.4	--	--	230	20	Na Ca SO_4	Deep	Institutional
--	--	--	880	<3	3	Na HCO_3 Cl	Deep	Municipal
--	<1.0	--	--	<100	<30	Na Cl		
--	4.7	--	--	<100	<30	Na HCO_3	Deep	Municipal
--	--	--	660	<3	5	Na Cl	Deep	Municipal
--	--	.010	620	13	3	Na Cl		
.58	--	.020	560	48	6	Na Cl HCO_3 SO_4	Deep	Municipal
.26	--	.010	1,900	92	36	Na Cl	Deep	Observation
--	--	--	2,900	40	50	Na Cl SO_4	Deep	Observation

Supplemental Data B. Chemical analyses of ground water in Indian Wells Valley, 1987-88--*Continued*

Well number	Date	Time	Altitude of land surface	Depth of well, total	Perforated interval	Temper-ature (°C)	Specific conduct-ance (μS/cm)	pH (standard units)
			Southeastern area--*Continued*					
26S/40E-36A1	05-28-87	1135	2,247.2	270.2	80-90, 107-127; 187-195; 240-260	21.5	2,320	7.5
27S/40E-1M2	07-08-88	0925	2,290	210	150-210	25.0	990	8.1
27S/40E-2G3	07-08-88	2020	2,258	500	--	25.0	2,600	8.2
27S/40E-2H1	06-30-88	1610	2,275	200	119-197	29.5	1,200	7.9
27S/40E-2J1[1]	06-01-87	1330	2,300	220	--	25.0	2,200	7.7
	06-30-88	1410				26.0	2,000	7.7
27S/40E-4B2	05-29-87	0825	2,298	288	128-278	26.0	1,400	8.0
	[11]06-11-87	0836				--	1,200	8.01
27S/40E-5D1	05-29-87	0845	2,375	555	251-555	25.0	620	8.8
27S/40E-6H1	06-02-87	1500	2,395	402	220-400	29.5	560	8.9
	[3]01-14-88	1050				--	500	8.21
27S/40E-6R2	[3]02-11-88	--	2,400	--	--	--	1,200	8.02
27S/40E-8A1	[3]01-14-88	1015	2,345	--	--	--	600	8.26
27S/40E-8B1	06-02-87	1445	2,380	400	200-400	29.5	720	8.7
	[3]01-14-88	0921				--	630	8.28
27S/40E-8B2	[3]01-14-88	0931	2,365	400	200-400	--	670	8.17
27S/40E-8C2	[3]01-14-88	0900	2,388	400	220-400	--	670	7.81
27S/40E-8F1	[3]01-14-88	0844	2,397	400	--	--	790	8.26
27S/40E-8Q2	06-02-87	1415	2,430	367	--	30.0	1,240	8.8
	[3]01-14-88	0830				--	790	8.19
			Intermediate area					
26S/39E-13R3	05-30-87	1645	2,318	300	150-170; 200-220; 240-260	26.0	365	8.0
26S/39E-13R4	01-18-89	1500	2,318	800	640-800	33.0	1,300	9.0
26S/39E-20R2	05-29-87	1330	2,421	920	600-900	28.5	365	8.2
26S/39E-24P1[1]	05-29-87	1205	2,355	800	250-350; 490-580; 640-780	30.5	360	9.0
26S/39E-25E1	05-29-87	0910	2,345	387	179-260; 268-284; 291-384	26.0	450	7.9
	[11]06-12-87	--			--	400	8.10	
26S/39E-26B3	[3]01-07-88	1415	2,384	--	--	--	400	8.11
26S/39E-27C1	[3]01-07-88	1430	2,415	500	--	--	440	8.02
26S/40E-30E2	05-29-87	0930	2,345	378	205-378	27.0	410	8.1
	[3]06-11-87	--			--	390	8.16	
26S/40E-30K1	[3]01-07-88	--	2,340	800	250-800	--	380	8.28
26S/40E-30K2	05-14-87	1600	2,340	802	220-470; 600-760	32.0	375	8.9
	[3]01-07-88	--				--	370	8.2
27S/39E-12M1	[12]06----87	--	2,530	--	--	--	550	8.0
	[12]09----87	--				--	550	7.8
27S/40E-6D1	05-28-87	1440	2,400	720	580-700	33.0	380	9.0
	[3]01-07-88	1400				--	440	8.2

Footnotes at end of table.

Hardness (mg/L, as $CaCO_3$)	Hardness, noncarbonate (mg/L, as $CaCO_3$)	Calcium, dissolved (mg/L, as Ca)	Magnesium, dissolved (mg/L, as Mg)	Sodium, dissolved (mg/L, as Na)	Percent sodium	Sodium adsorption ratio	Potassium, dissolved (mg/L, as K)	Alkalinity, field (mg/L, as $CaCO_3$)	Alkalinity, lab (mg/L, as $CaCO_3$)
				Southeastern area--*Continued*					
450	290	140	24	260	55	6	8.7	162	173
91	0	0	3.8	180	80	9	4.1	155	157
250	140	69	20	410	77	12	11	114	105
81	0	6	4.0	200	83	10	5.0	167	176
230	67	75	11	320	74	10	6.1	166	165
260	100	81	13	320	72	9	7.0	156	167
220	120	57	18	160	61	5	5.3	92	101
312.8	--	96.4	17.5	122.0	45.1	3.0	8.2	--	161.3
9	0	2.9	.42	120	96	18	.7	164	166
5	0	1.9	.11	110	98	22	.4	154	157
21.6	--	5.9	1.7	106.0	91.0	9.9	.9	--	137.5
50.4	--	17.9	1.4	249.0	91.2	15.3	1.6	--	147.2
44.8	--	16.7	.8	117.0	84.4	7.6	1.6	--	129.3
26	0	9.0	.84	130	91	12	.9	172	182
41.6	--	14.9	1.1	125.0	86.1	8.4	1.6	--	161.9
45.2	--	15.4	1.7	134.0	85.9	8.6	1.8	--	162.8
109.6	--	41.0	1.7	105.0	67.2	4.4	1.7	--	149.5
56.8	--	20.9	1.1	154.0	84.9	8.9	2.1	--	149.0
20	0	7.3	.32	220	96	23	1.6	146	152
20.8	--	5.3	1.8	175.0	94.4	16.8	1.5	--	158.7
				Intermediate area--*Continued*					
93	0	29	5.0	30	40	1	2.7	118	126
12	0	3.7	.56	320	98	41	3.4	481	475
34	0	10	2.3	62	79	5	1.9	103	107
59	4	21	1.7	40	59	2	1.4	56	71
110	23	34	7.0	35	40	1	2.1	91	98
116.0	--	37.8	5.3	32.9	37.3	1.3	3.2	--	76.9
116.0	--	37.8	4.2	38.0	41.9	1.6	2.3	--	88.8
117.6	--	40	4.2	41.8	43	1.7	2.5	--	92.9
110	33	29	8.6	32	38	1	4.0	75	95
106.4	--	31.9	6.5	29.2	36	1.2	5.2	--	74.0
34.8	--	10.3	2.2	68.7	80.2	5.1	1.7	--	105.3
47	0	13	3.5	60	73	4	1.9	89	101
12.8	--	3.5	1.0	81.1	92.7	9.8	.8	--	116.8
188	--	47	13	--	--	--	--	--	--
133	--	50	14	--	--	--	--	--	--
5	0	2.0	.03	76	97	15	.6	97	113
10.8	--	3.5	.5	97.3	94.6	12.9	1.0	--	131.6

Supplemental Data B. Chemical analyses of ground water in Indian Wells Valley, 1987-88--*Continued*

Well number	Sulfate, dissolved (mg/L, as SO_4)	Chloride, dissolved (mg/L, as Cl)	Fluoride, dissolved (mg/L, as F)	Silica, dissolved (mg/L, as SiO_2)	Solids, residue at 180 °C, dissolved (mg/L)	Solids, sum of constituents, dissolved[12] (mg/L)	Nitrogen, NO_2+NO_3 dissolved (mg/L, as N)	Nitrogen ammonia, dissolved (mg/L, as N)	Nitrogen ammonia + organic, dissolved (mg/L, as N)
					Southeastern area--*Continued*				
26S/40E-36A1	350	360	0.5	57	--	1,310	1.40	--	--
27S/40E-1M2	38	190	1.6	52	--	599	1.10	0.020	0.30
27S/40E-2G3	100	650	1.0	9.7	--	1,350	.87	.040	.40
27S/40E-2H1	92	200	1.1	51	--	684	.85	<.010	.60
27S/40E-2J1[1]	54	480	2.9	69	--	1,130	1.30	--	--
	63	510	1.1	58	--	1,160	1.60	.010	.70
27S/40E-4B2	47	320	.7	44	--	717	2.00	--	--
	81.0	239.2	2.7	--	742.8	676	--	--	--
27S/40E-5D1	20	67	1.4	36	--	352	1.00	--	--
27S/40E-6H1	15	61	1.1	36	--	321	.53	--	--
	15.4	60.8	.9	--	317.5	274	--	--	--
27S/40E-6R2	25.8	288.4	.5	--	759.8	682	--	--	--
27S/40E-8A1	24.3	94.9	1.0	--	397.8	334	--	--	--
27S/40E-8B1	24	86	1.4	41	--	398	.11	--	--
	22.1	84.2	1.0	--	420.1	347	--	--	--
27S/40E-8B2	22.8	91.0	1.0	--	431.9	365	--	--	--
27S/40E-8C2	23.2	106.7	1.2	--	434.2	370	--	--	--
27S/40E-8F1	26.7	142.7	1.0	--	531.7	438	--	--	--
27S/40E-8Q2	34	250	1.0	--	--	645	.12	--	--
	26.6	140.8	.9	--	558.8	448	--	--	--
					Intermediate area--*Continued*				
26S/39E-13R3	5.1	24	.80	55	--	223	<.10	--	--
26S/39E-13R4	13	160	3.7	29	--	826	<.10	.510	.70
26S/39E-24P1[1]	21	21	.8	29	--	215	1.20	--	--
26S/39E-20R2	40	27	.8	27	--	199	1.50	--	--
26S/39E-25E1	47	32	.7	36	--	261	2.90	--	--
	71.3	29.0	.8	--	250.9	240	--	--	--
26S/39E-26B3	51.9	30.6	.7	--	250.8	229	--	--	--
26S/39E-27C1	50.9	30.2	.8	--	269.3	234	--	--	--
26S/40E-30E2	39	27	.6	38	--	234	2.40	--	--
	60.8	24.9	.8	--	255.1	216	--	--	--
26S/40E-30K1	22.0	28.7	1.1	--	237.1	203	--	--	--
26S/40E-30K2	27	36	.9	33	--	240	2.50	.100	1.1
	17.3	29.2	1.5	--	226.8	207	--	--	--
27S/39E-12M1	--	--	44.7	--	365	--	--	--	--
	--	45.2	--	--	340	--	--	--	--
27S/40E-6D1	21	31	.9	45	--	235	<.10	--	--
	22.3	34.3	1.2	--	278.6	239	--	--	--

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[1] Wells sampled by Warner (1975) in 1972.

[2] Water-quality analysis furnished by the China Lake Naval Air Weapons Station.

[3] Water-quality analysis furnished by the Indian Wells Valley Water District.

[4] Water-quality analysis furnished by the Inyokern Community Services District.

[5] Sample depth = 495 feet.

[6] Sample depth = 595 feet.

[7] Sample depth = 700 feet.

Nitrogen, organic, dissolved (mg/L, as N)	Nitrate, total (mg/L as NO_3)	Phosphorus, dissolved (mg/L, as P)	Boron, dissolved (µg/L, as B)	Iron, dissolved (µg/L, as Fe)	Manganese, dissolved (µg/L, as Mn)	Predominant ions	Monitored aquifer	Well use
				Southeastern area--*Continued*				
--	--	--	2,900	40	50	Na Cl SO_4	Deep	Observation
0.28	--	0.010	1,400	<3	2	Na Cl	Deep	Domestic
.36	--	.020	4,100	20	30	Na Cl	Deep	Observation
--	--	.020	1,200	99	6	Na Cl	Deep	Institutional
--	--	--	2,800	60	30	Na Cl	Deep	Institutional
.69	--	.020	2,700	20	<10	Na Cl		
--	--	--	1,200	20	7	Na Cl	Deep	Industrial
--	14.5	--	--	40	<20	Na Ca Cl		
--	--	--	1,000	59	<1	Na HCO_3	Deep	Industrial
--	--	--	760	<3	<1	Na HCO_3	Deep	Municipal
--	1.0	--	--	<100	<30	Na HCO_3		
--	9.3	--	--	200	<30	Na Cl	Deep	Municipal
--	1.5	--	--	<100	<30	Na Cl HCO_3	Deep	Municipal
--	--	--	1,100	<3	5	Na HCO_3	Deep	Municipal
--	<1.0	--	--	<100	<30	Na HCO_3		
--	<1.0	--	--	<100	<30	Na HCO_3	Deep	Municipal
--	<1.0	--	--	<100	<30	Na Cl HCO_3	Deep	Municipal
--	1.5	--	--	<100	<30	Na Cl	Deep	Municipal
--	--	--	1,400	25	11	Na Cl	Deep	Municipal
--	1.5	--	--	<100	<30	Na Cl	Deep	
				Intermediate area--*Continued*				
--	--	--	170	23	77	Ca Na HCO_3	Deep	Observation
.19	-	.170	3,500	44	1	Na HCO_3	Deep	Observation
--	--	--	260	<3	9	Na HCO_3	Deep	Military
--	--	--	150	24	9	Na HCO_3 SO_4 Cl	Deep	Military
--	--	--	220	5	10	Ca Na HCO_3 SO_4 Cl	Deep	Industrial
--	14.5	--	--	20	<20	Ca Na HCO_3 SO_4 Cl		
--	10.5	--	--	<100	<30	Ca Na HCO_3 SO_4 Cl	Deep	Municipal
--	8.9	--	--	<100	<30	Ca Na HCO_3 SO_4 Cl	Deep	Municipal
--	--	--	150	14	15	Ca Na $MgHCO_3$ SO_4Cl	Deep	Industrial
--	12.5	--	--	<20	<20	Ca $NaHCO^3$ SO^4 Cl		
--	6.0	--	--	<100	<30	Na HCO_3	Deep	Municipal
1.0	--	.010	390	20	<1	Na HCO_3	Deep	Municipal
--	3.9	--	--	<100	<30	Na HCO_3		
--	--	--	--	50	--	--	Deep	Institutional
--	5.5	--	--	50	--	--		
--	--	--	610	<3	3	Na HCO_3	Deep	Municipal
--	<1.0	--	--	100	<30	Na HCO_3		

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[8] Sample depth = 800 feet.

[9] Sample depth = 895 feet.

[10] Sample depth = 955 feet.

[11] Water-quality analysis furnished by Kerr McGee Chemical Corporation.

[12] Water-quality analysis furnished by the Kern County Public Works.

[13] Water-quality analysis furnished by the Ridgecrest Community Hospital.

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