located near Bishop, near Big Pine, north of Independence, and near Lone Pine (Hollett and others, 1991, fig. 5).

HYDROLOGIC SYSTEM

The hydrologic system of the Owens Valley can be conceptualized as having three parts: (1) an unsaturated zone affected by precipitation and evapotranspiration; (2) a surface-water system composed of the Owens River, the Los Angeles Aqueduct, tributary streams, canals, ditches, and ponds; and (3) a saturated ground-water system contained in the valley fill.

The following evaluation identifies key components of the hydrologic system, describes their interaction, and quantifies their spatial and temporal variations. Discussion of the unsaturated zone is limited to precipitation and evapotranspiration. The evaluation also includes the interaction between the hydrologic system, much of which has been altered by human activity, and the native vegetation; this interaction is the subject of recent controversy and litigation.

For purposes of organization, the surface-water and ground-water systems are presented separately. For items that have both a surface-water and a ground-water component, such as the river–aqueduct system, the discussion is presented in the section entitled "Surface-Water System"; included in this convention is the quantification of ground-water recharge and discharge. All water-budget calculations are for the area defined by Hollett and others (1991) as the aquifer system (figs. 4 and 5). Three key periods—water years 1963–69, water years 1970–84, and water years 1985–88—were used to calculate historical water budgets, to calibrate the valleywide ground-water flow model, to verify performance of the model, and to evaluate past and possible future changes in the surface-water and ground-water systems (table 4). A complete description of the ground-water flow model is included in the section entitled "Ground-Water System."

Precipitation and Evapotranspiration

Precipitation

The pattern of precipitation throughout the Owens Valley is strongly influenced by altitude, and precipitation varies in a predictable manner from approximately 4 to 6 in/yr on the valley floor to more than 30 in/yr at the crest of the Sierra Nevada on the west side of the valley (Groeneveld and others, 1986a, 1986b; Duell, 1990; Hollett and others, 1991, fig. 3). On the east side of the valley, precipitation follows a similar pattern, but with somewhat lower rates of 7 to 14 in/yr because of the lower altitude of the Inyo and the White Mountains and the rain-shadow effect caused by the Sierra Nevada. Snow, when present on the Sierra Nevada and the White Mountains, commonly is absent on the Inyo Mountains (fig. 3) and the Coso Range. Of the total average annual precipitation in the Owens Valley drainage area, about 60 to 80 percent falls as snow or rain in the Sierra Nevada, primarily during the period October to April. A lesser quantity falls during summer thunderstorms.

As shown in figure 7*A*, the pattern of average precipitation is well defined by the more than 20 precipitation and snow-survey stations that have been monitored routinely, many for more than 50 years (fig. 7*C*). Average precipitation tends to increase from south to north, much as does altitude of the land surface. The strong correlation between altitude and recent mean annual precipitation can be seen in figure 7*B* and can be described by the regression equation,

$$P_i^{RAVE} = 0.00245\ LSD_i - 3.205\ , \qquad (1)$$

where

P^{RAVE} is recent mean annual precipitation, in inches per year, on the basis of data for rain years 1963–84;

LSD is altitude of land surface, in feet above sea level; and

i is an index referring to location.

Regression equation 1 was fitted by hand from figure 7*B*, which is a graph of data presented in figure 7*C*, with an emphasis on data from the west side of the valley where the bulk of the more transmissive materials of the ground-water system are present (fig. 4). Predictably, the White Mountain Stations 1 and 2 (sites 19 and 20, fig. 7*B*) fall somewhat below the line. A similar relation that more accurately represents precipitation falling on the east side of the valley could be developed (Lopes, 1988, fig. 3). However, that relation would need to account for the difference between the quantity of precipitation falling on the White Mountains and farther south on the Inyo Mountains

(fig. 3)—only part of which seems to be attributable to a difference in altitude of the two mountain ranges.

The time period (rain years 1963–84) used to develop equation 1 was chosen on the basis of two criteria: a nearly complete record for all 20 stations and symmetry with the period selected for calibration of the ground-water flow model. Because very little precipitation occurs in the Owens Valley during July through September, precipitation values for a rain year (July 1–June 30) are virtually identical to values for the corresponding water year (October 1–September 30), which is used to summarize streamflow and ground-water pumpage data. Equation 1 can be generalized for a much longer period of record using data for the U. S. Weather Bureau station at Independence (site 10, fig. 7*C*). Long-term mean annual precipitation at this station, for the 99-year period 1886–1985, is 5.10 in/yr (M.L. Blevins, Los Angeles Department of Water and Power, written commun., 1986)—in comparison with 5.98 in/yr for rain years 1963–84. Scaling equation 1 by the ratio 5.10/5.98 produces an estimate of the long-term mean annual precipitation (P^{LTAVE}) at any location along the west side of the valley. This relation is:

$$P_i^{LTAVE} = \frac{5.10}{5.98} P_i^{RAVE}, \qquad (2)$$

where units of both P^{LTAVE} and P^{RAVE} are inches per year. Precipitation $(P_{i,j}^{AN})$ for a particular year (j) can be estimated by using annual precipitation at the Independence station $(P_{Ind,j}^{AN})$ for that same year as a weighting factor:

$$P_{ij}^{AN} = P_i^{LTAVE}\left[\frac{P_{Ind,j}}{5.10}\right], \qquad (3)$$

where

P^{AN} is annual precipitation, in inches per year;

P^{LTAVE} is long-term mean annual precipitation, in inches per year; and

P_{Ind} is annual precipitation at the U.S. Weather Bureau station at Independence, in inches per year.

Estimates of precipitation based on equations 1, 2, and 3 for locations on the valley floor need to be used cautiously because of significant local variability in precipitation (fig. 7*B*).

Although the spatial distribution of mean annual precipitation is well documented and highly correlated with altitude (fig. 7*B*), the spatial distribution of precipitation during specific years is highly variable (Hollett and others, 1991, fig. 3). For example, annual precipitation at Bishop and at Independence was compared for rain years 1935–88 (fig. 8). On average, similar quantities of precipitation fall at Bishop and at Independence (sites 2 and 10, respectively, fig. 7*C*). This similarity occurs because both sites are located on the valley floor and differ in altitude by less than 160 ft. As shown in figure 8, however, it is not uncommon for either site to have more, sometimes much more, precipitation during a particular year. C.H. Lee (1912, p. 15) noted that the high variability in precipitation in the Owens Valley is the result of the three distinct types of storms that occur in the area: (1) north Pacific storms that dominate the rainy season and provide most of the precipitation both to the mountain areas and the valley floor, (2) south Pacific storms that migrate north up the valley (usually a few times each year) generating sporadic precipitation, but favoring neither the Sierra Nevada nor the Inyo Mountains, and (3) local storms which occur during summer and which are an important contributor to total precipitation on the east side of the valley. This annual and seasonal variability makes continued monitoring of precipitation at various sites throughout the valley important—especially because both the quantity and the timing of precipitation on the valley floor play a critical role in the water use and the health of native vegetation (Sorenson and others, 1991). Ground-water recharge from precipitation is highly dependent on the quantity of water used by the overlying vegetation and is discussed in the next section on evapotranspiration.

Evapotranspiration

Evapotranspiration by the dominant native vegetation of the valley had not been measured since the detailed lysimeter studies by C.H. Lee (1912) in the early 1900's. Instead, evapotranspiration was estimated as the residual, a very large residual, in numerous water-budget studies (California Department of Water Resources, 1960, 1965, 1966; Los Angeles Department of Water and Power, 1972, 1976, 1978, 1979; Danskin, 1988). A key element of the cooperative studies begun in 1982 by the U.S. Geological Survey, Inyo County, and the Los Angeles Department of Water and Power

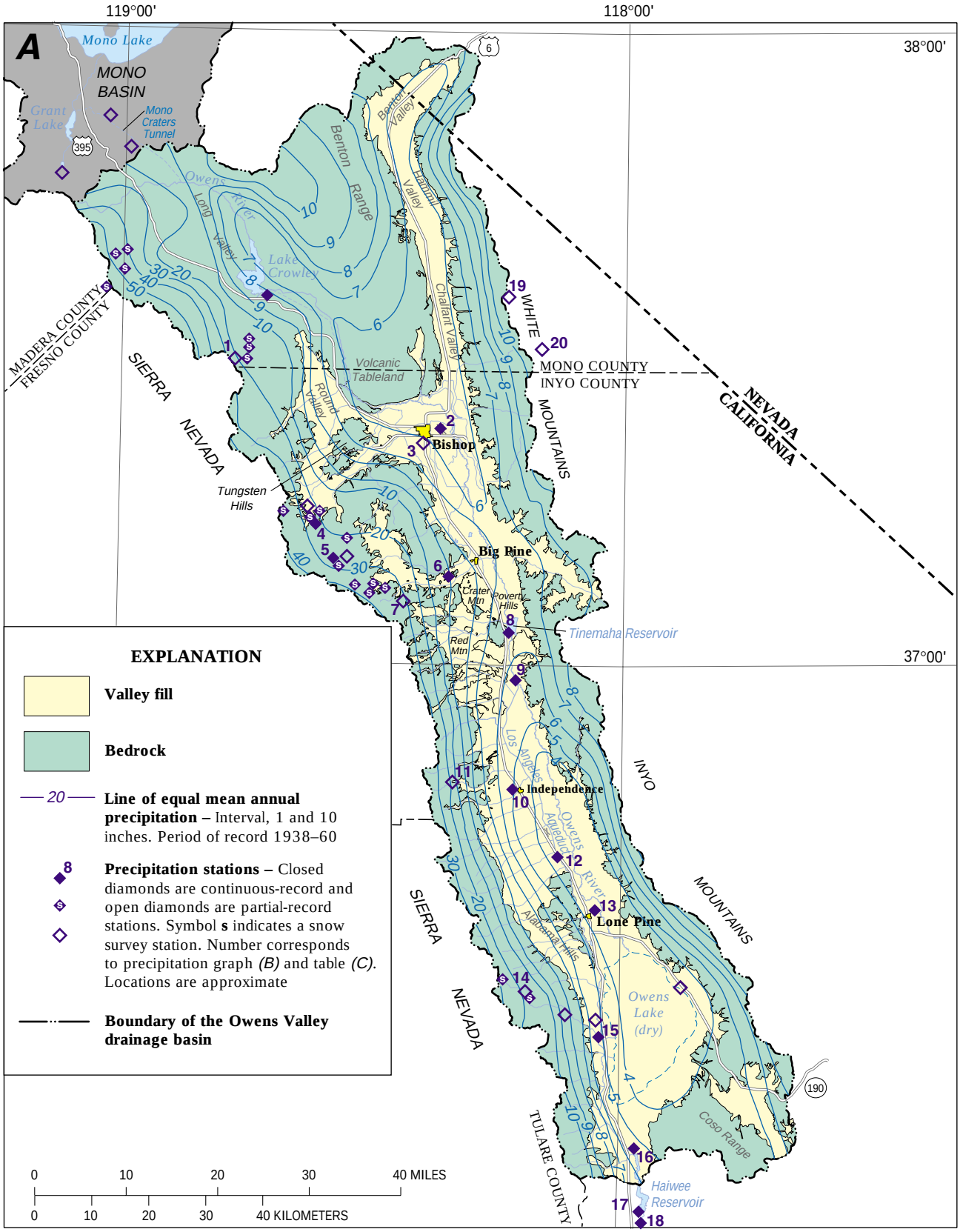
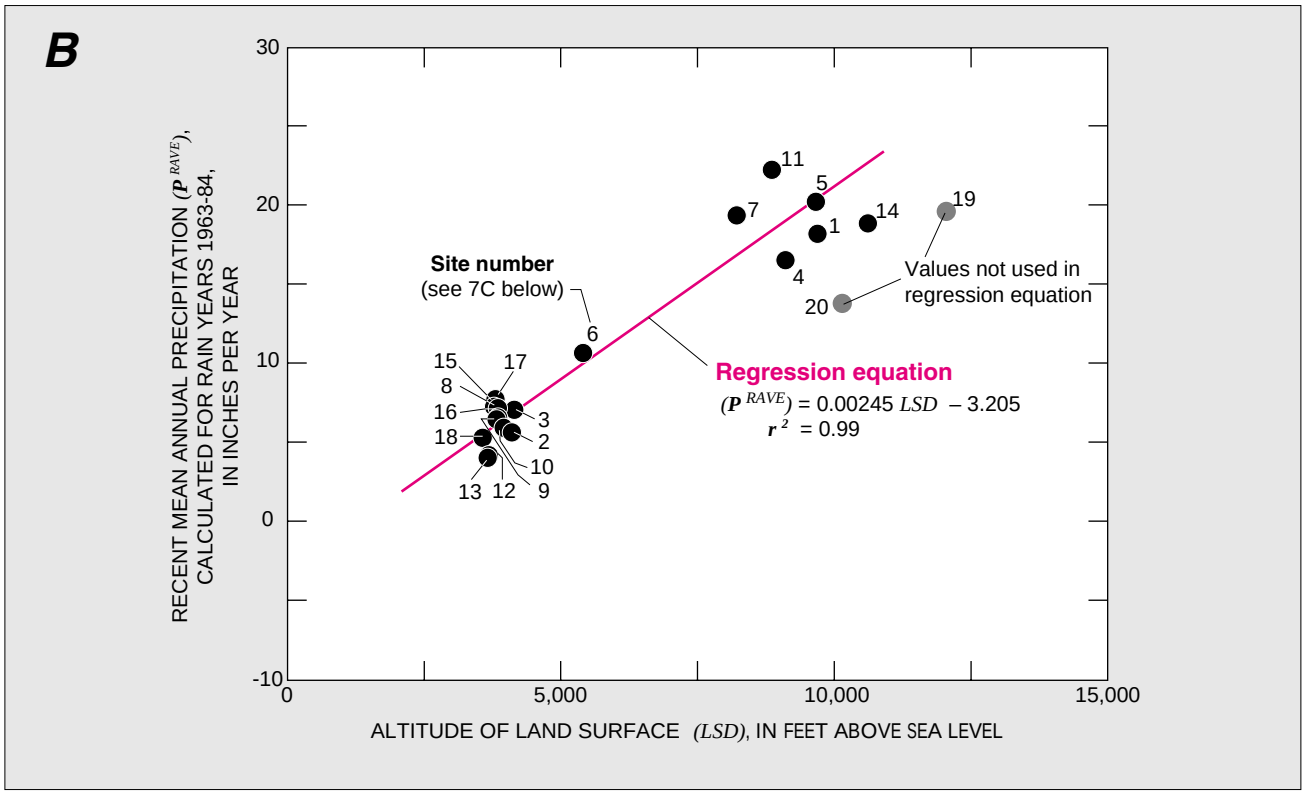


Figure 7. ***(A)*** Contours of mean annual precipitation; ***(B)*** relation between recent mean annual precipitation and altitude; and ***(C)*** data for selected precipitation stations in the Owens Valley, California. Data from E.L. Coufal, Los Angeles Department of Water and Power, written commun., 1986, and oral commun., 1989. Map modified from Stetson, Strauss, and Dresselhaus, consulting engineers, written commun., 1961.



C

Site no.	Station name	Recent mean annual precipitation for rain years 1963-84 (inches/year)	Altitude (feet)	Latitude (north)	Longitude (west)	Period of record (rain years)
1.	Rock Creek at store	18.30	9,700	37°27'	118°45'	1948-88
2.	U.S. Weather Bureau, Bishop	5.67	4,108	37°22'	118°22'	1931-88
3.	Bishop Yard	7.12	4,140	37°21'	118°24'	1931-88
4.	U.S. Weather Bureau, Lake Sabrina	16.56	9,100	37°13'	118°37'	1926-88
5.	U.S. Weather Bureau, South Lake	20.30	9,620	37°11'	118°34'	1926-88
6.	Big Pine Power House No. 3	10.72	5,400	37°08'	118°20'	1927-88
7.	Big Pine Creek at Glacier Lodge	19.45	8,200	37°06'	118°26'	1948-88
8.	Tinemaha Reservoir	7.20	3,850	37°04'	118°14'	1935-88
9.	Los Angeles Aqueduct at intake	6.49	3,825	36°58'	118°13'	1932-88
10.	U.S. Weather Bureau, Independence	5.98	3,950	36°48'	118°12'	1886-1988
11.	Onion Valley	[1]22.77	8,850	36°46'	118°20'	1950-88
12.	Los Angeles Aqueduct at Alabama Gates	4.24	3,675	36°41'	118°05'	1931-88
13.	Lone Pine	4.06	3,661	36°36'	118°04'	1919-88
14.	Cottonwood at Golden Trout Camp	[1]19.04	10,600	36°29'	118°11'	1948-81
15.	Cottonwood Gates	7.31	3,775	36°25'	118°02'	1928-88
16.	North Haiwee Reservoir	6.60	3,850	36°14'	117°58'	1931-88
17.	South Haiwee Reservoir	7.79	3,800	36°08'	117°57'	1924-88
18.	Haiwee Power House	[1]5.34	3,570	36°07'	117°57'	1930-75
19.	White Mountain No. 2	[1]19.73	12,070	37°35'	118°14'	1953-88
20.	White Mountain No. 1	[1]13.94	10,150	37°30'	118°10'	1950-77

[1] Short or discontinuous record.

Figure 7. Continued.

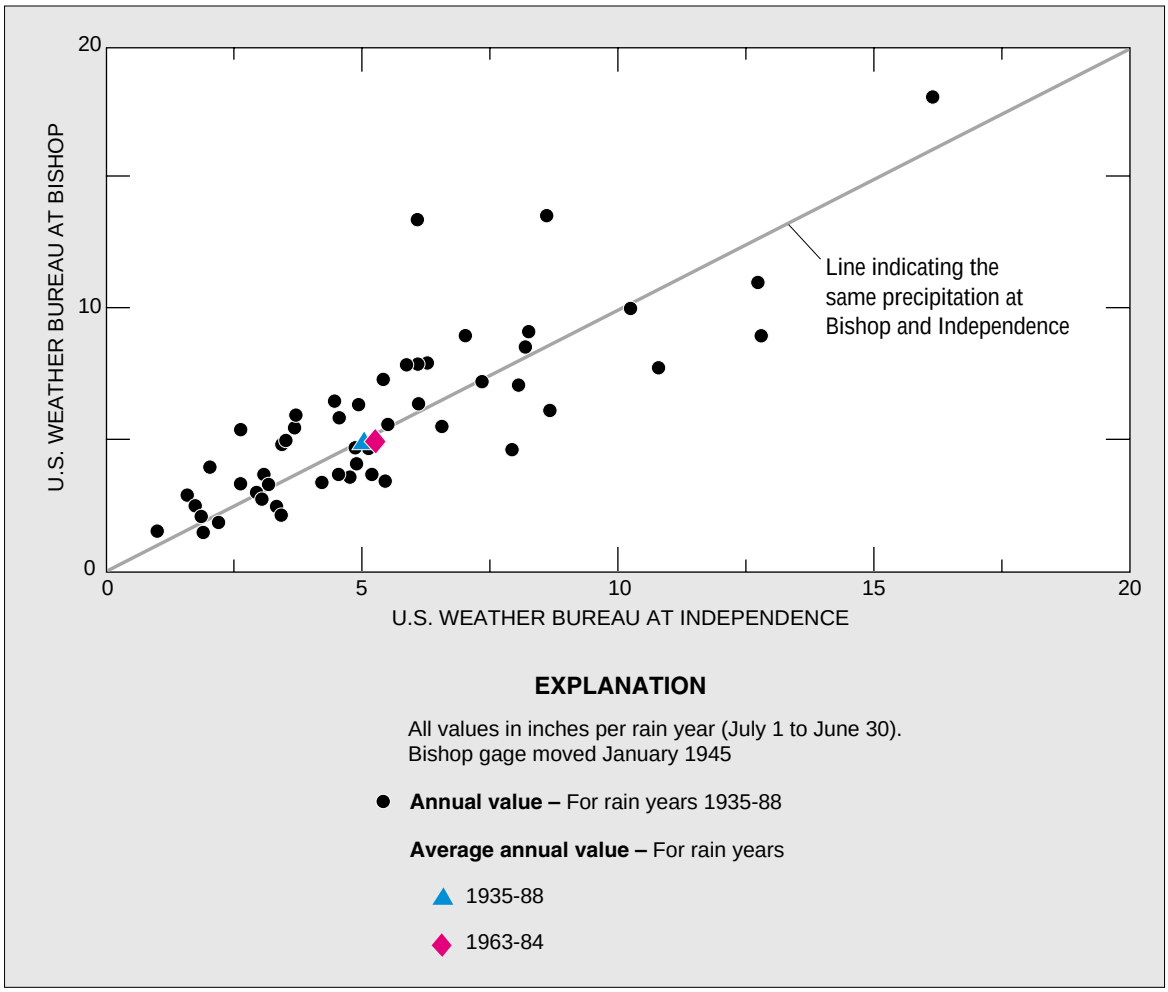


Figure 8. Annual precipitation as Bishop and Independence, California (sites 2 and 10, respectively, in figure 7).

was to measure evapotranspiration at representative vegetation study sites throughout the valley (fig. 2), to relate these data to soil and plant characteristics at the sites, to extend the relations to quantify evapotranspiration throughout the valley, and then to synthesize the results in an analysis of the overall hydrologic system.

As part of the studies of native vegetation, Duell (1990) used micrometeorologic equipment to collect detailed evapotranspiration measurements during 1984–85, a period of relatively abundant surface water and ground water in the valley. The results for high-ground-water alkali meadow and alkali scrub communities (fig. 6 and table 3), which are summarized in table 5, show that evapotranspiration rates on the valley floor ranged from about 12 in/yr to about 45 in/yr depending on the type and percentage of vegetative cover. Assuming that these rates are representative of average conditions on the valley floor where the depth to water is approximately 3 to 15 ft, then evapotranspiration is about 3 to 6 times greater than the quantity of precipitation that is available.

During the same period and at the same sites, Groeneveld and others (1986a, 1986b) collected transpiration measurements from native vegetation using a porometer, an instrument that encloses a few leaves of a plant and measures water-vapor flux (Beardsell and others, 1972). These measurements can be converted to transpiration from an entire site using measurements of total leaf area per plant and plant density per site. Results from Groeneveld and others (1986a, p.117) suggest that most of the evapotranspiration measured by Duell (1990) is transpiration from native vegetation.

Coincident monitoring of soil moisture at the same sites indicated that most of the transpired water came from the unsaturated zone, including that part just below the land surface. These findings indicate that the

plants, although originally classified as phreatophytes, might be described more accurately as facultative phreatophytes (Sorenson and others, 1991). However, one common plant on the valley floor, *Atriplex torreyi* (Nevada saltbush) (tables 3 and 5), was found to be restricted to shallow-ground-water zones. The phenology, reproductive processes, and flooding tolerance of *Atriplex torreyi* suggests that it is an obligate

Table 5. Composition of native plant communities, ground-water-level and precipitation data, and range in evapotranspiration estimates at vegetation study sites in the Owens Valley, California

[nc, not collected; —, not available; USGS, U.S. Geological Survey. Vegetation data from the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1984, 1987); evapotranspiration estimates from Duell, 1990. Estimated annual evapotranspiration from the saturated ground-water system equals average annual evapotranspiration for 1984–85 minus annual precipitation for 1984]

Site designation (figure 2 and table 1)	Well number (table 1)	Native high-ground-water plant community (table 3)	Most common plant types		Total vegetative cover (percent)	Range of ground-water levels for 1984 (feet below land surface)	Annual precipitation for 1984 (inches)	Annual evapotranspiration for 1984–85 (inches)			Estimated annual evapotranspiration from the saturated ground-water system for 1984–85 (inches)
			Common name	Percentage of total vegetation				Maximum	Average	Minimum	
A	USGS 1	Alkaline meadow.	Alkali sacaton...	43	42	10.5–15.5	nc	33.6	32.3	30.9	—
			Russian thistle ..	22							
C	USGS 2	Alkaline meadow.	Saltgrass	34	35	10.2–11.4	5.9	21.8	18.5	14.8	12.6
			Rubber rabbitbrush.	25							
E	USGS 3	Alkaline scrub.	Rubber rabbitbrush.	24	26	10.2–10.9	nc	23.6	23.6	23.5	—
			Alkali sacaton...	23							
			Mormon tea	8							
F	USGS 5	Alkaline scrub.	Saltgrass	34	24	8.0–9.0	6.3	18.9	15.2	11.9	8.9
			Greasewood.....	27							
G	USGS 6	Alkaline meadow.	Saltgrass	30	33	7.1–8.9	nc	25.8	24.3	22.8	—
			Alkali sacaton...	13							
			Rubber rabbitbrush.	9							
J	USGS 7	Alkaline meadow.	Nevada saltbush.	29	50	4.7–7.2	nc	33.0	32.0	31.0	—
			Alkali sacaton...	21							
			Rubber rabbitbrush.	16							
L	USGS 10 ..	Alkaline meadow.	Saltgrass	20	72	.1–3.9	3.1	44.8	40.5	33.1	37.4
			Alkali sacaton...	17							
			Baltic rush	15							

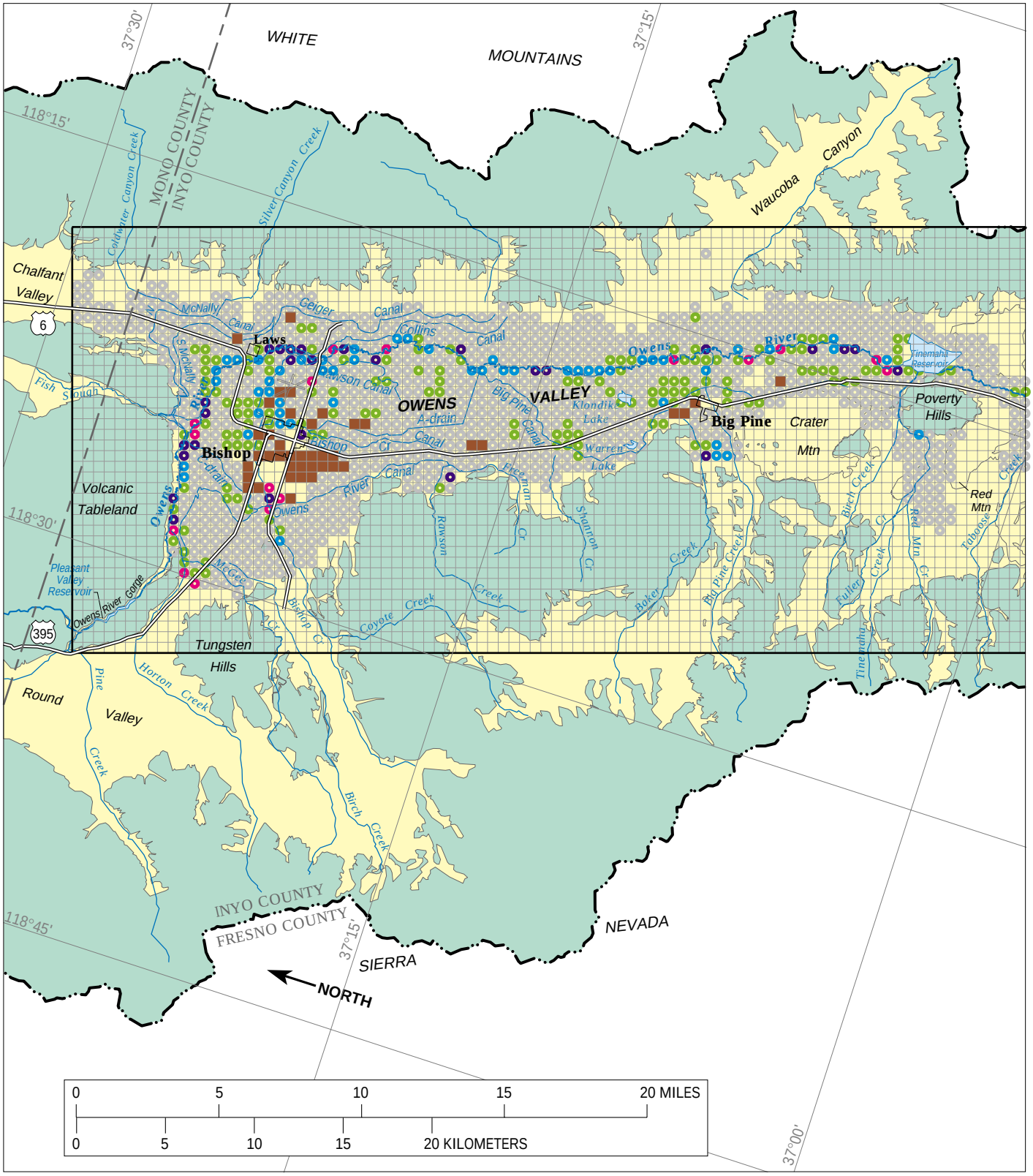


Figure 9. Estimated average annual transpiration by native vegetation during water years 1983–87 in the Owens Valley, California. Map values derived from more than 14,000 point estimates of average annual evapotranspiration obtained from the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1988).

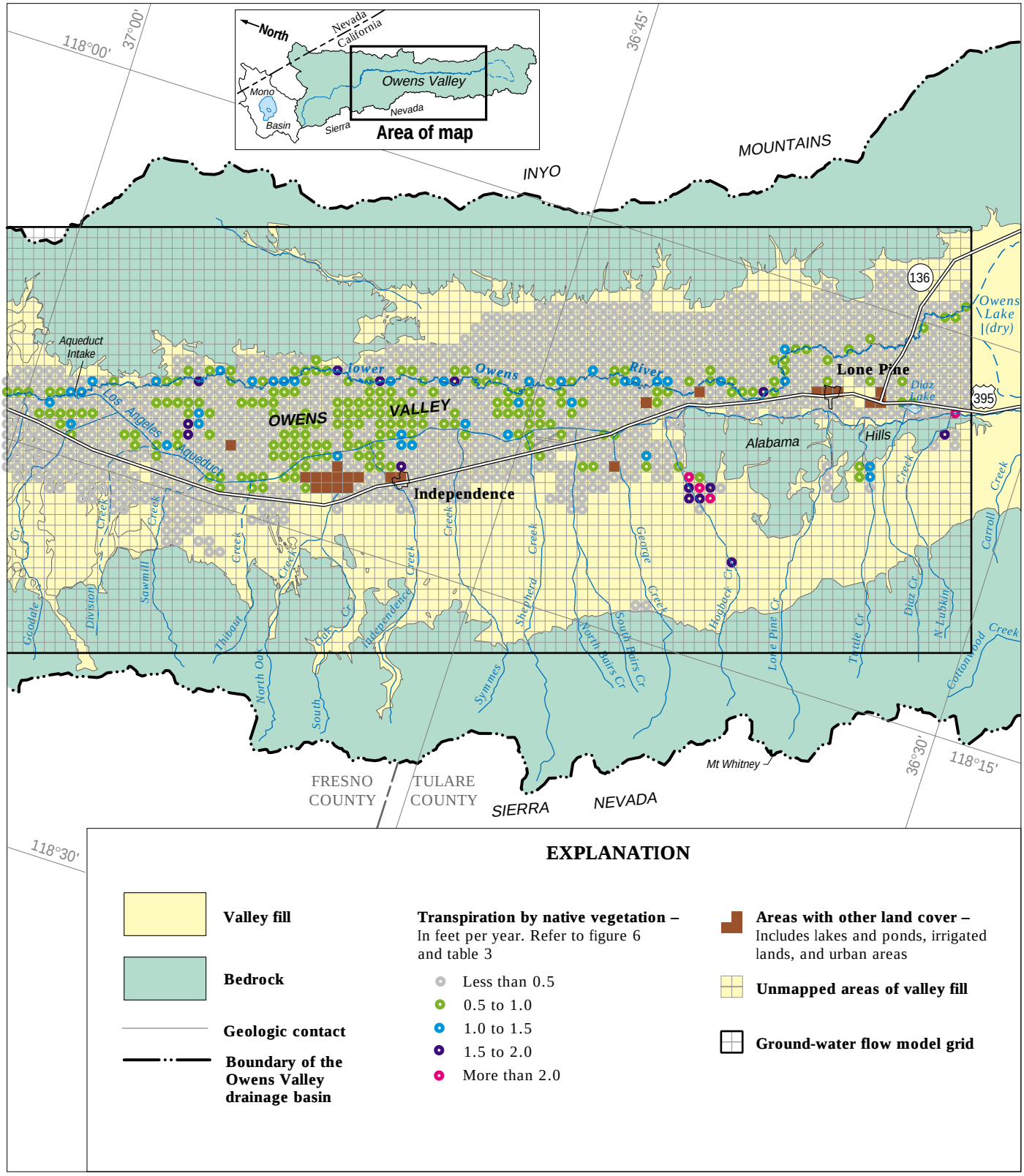


Figure 9. Continued.

phreatophyte in the Owens Valley (Groeneveld, 1985). This species also was found by Dileanis and Groeneveld (1989) to be among the most drought tolerant of the dominant species on the valley floor.

Soil-moisture monitoring also indicated that much of the precipitation that falls on the valley floor (fig. 7) percolates into the near-surface unsaturated zone and later is transpired by native vegetation (Sorenson and others, 1991). Except during brief periods of rainfall or snowmelt, or in areas where the water table is nearly at the land surface, evaporation is not a dominant part of evapotranspiration from the valley floor.

The findings of Duell (1990) and Groeneveld and others (1986a, 1986b; 1987) were combined with extensive mapping of vegetation by the Los Angeles Department of Water and Power (D.D. Buchholz, written commun., 1988) in order to produce an estimate of average annual transpiration from the valley floor (fig. 9). The mapping was done in the field using aerial photographs and land-use maps. Data collected for each mapped area (parcel) included information about plant communities, species composition, percentage of bare ground, and land use. The data were compiled on topographic maps at a scale of 1:24,000 and then digitized into data points every 250 m (820 ft) based on the Universal Transverse Mercator grid system (Synder, 1982, 1985, 1987; Newton, 1985). These individual data points of total evapotranspiration were combined with regressed values of precipitation (fig. 7) and averaged using the grid of the valleywide ground-water flow model. Evaporation from the water table was assumed to be negligible for most areas of native vegetation and to be of minor importance in the limited areas of riparian plants. To maintain consistency with analysis of the same data done by the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1988), about 50 percent of the precipitation on the valley floor was assumed to evaporate. This percentage is reasonable but has a high degree of uncertainty (D.N. Tillemans, Los Angeles Department of Water and Power, oral commun., 1987). The resulting transpiration values for native vegetation are summarized in figure 9.

Transpiration by native vegetation from most of the valley floor is less than 1.0 ft/yr, and transpiration from much of the valley floor, particularly along the east side of the valley, is less than 0.5 ft/yr. These estimates are generally lower than previous estimates of transpiration by native vegetation (R.H. Rawson, Los Angeles Department of Water and Power, written commun., 1986) and are lower than calculated values obtained by subtracting a percentage of precipitation from estimated evapotranspiration (Danskin, 1988; C.H. Lee, 1912). This reduction in transpiration is consistent with the lower values of valleywide evapotranspiration calculated by Hollett and others (1991, table 6) in comparison with values from prior studies (C.H. Lee, 1912; Los Angeles Department of Water and Power, 1974b, 1975, 1976, 1978, 1979; Danskin, 1988). These prior studies quantified transpiration or evapotranspiration for periods before the additional diversions of water from the valley in 1913 and 1970. The additional diversions reduced the quantity of water available for transpiration by native vegetation.

In a few areas of the valley floor, infiltration to the water table may occur during part of the year. For example, in meadow areas, such as east of Independence, the water table is nearly at the land surface in winter months and some precipitation likely percolates to the saturated ground-water system. However, the high annual evapotranspiration rates observed by Duell (1990) in those areas—for example, at site L (table 5 and fig. 2)—indicate that the meadow areas are net discharge points from the ground-water system. Any water that infiltrates in winter is removed in summer. In other parts of the valley floor, such as small alkali flats or patches that are almost devoid of vegetation (fig. 3), net infiltration may result during unusually wet periods when rainfall or local runoff exceeds evapotranspiration. The quantity of infiltration from such microplaya areas, however, is very small because of extremely slow infiltration rates through these characteristically fine-textured, deflocculated soils (Groeneveld and others, 1986a). As in the meadow areas, wet conditions generally are present only in winter, and all the water infiltrated (perhaps with some additional ground water) is removed in summer when evapotranspiration rates increase markedly (Duell, 1990, fig. 24). For the area of the valley fill simulated by the valleywide ground-water flow model (fig. 4), average net discharge by evapotranspiration from the saturated aquifer system was estimated to decrease from 112,000 acre-ft/yr for water years 1963–69 to 72,000 acre-ft/yr for water years 1970–84.

In the alluvial fan deposits and volcanic rocks, the depth to water ranges from many tens to many hundreds of feet. Extraction of water by plants from the

saturated ground-water system is not possible, and the plants subsist on direct precipitation. Because the precipitation rates are higher than those on the valley floor (fig. 7), some recharge to the ground-water system may occur. However, the density of vegetation also is greater at the heads of fans and may balance the increased precipitation (M.O. Smith and others, 1990a, b). Any precipitation that does infiltrate past the root zone eventually recharges the saturated ground-water system, probably at a relatively uniform rate, and flows toward the center of the valley. About 16 percent of the direct precipitation on the alluvial fan areas was estimated to recharge the ground-water system (C.H. Lee, 1912). This percentage equates to about 1.25 to 2.75 in/yr of recharge. Ground-water simulation studies suggest that these rates may be too high and that maximum values of from 0.5 to 1.0 in/yr are more likely (Danskin, 1988; Hutchison, 1988; Hutchison and Radell, 1988a, b; Los Angeles Department of Water and Power, 1988). An investigation of recharge from precipitation in other arid regions indicated that recharge did not occur until precipitation rates exceeded about 8 in/yr (Mann, 1976, p. 368). The area of valley fill in the Owens Valley that has an average precipitation of more than 8 in/yr is limited to the higher attitudes, mostly along the western alluvial fans (fig. 7*A*). On the basis of these findings, equation 2 was used to calculate 5 percent of the average annual precipitation for values greater than 8 in/yr (fig. 7*A*). For the defined aquifer system (fig. 2), the total quantity of infiltration from direct precipitation, which occurs primarily on the alluvial fan deposits and volcanic rocks, averages approximately 2,000 acre-ft/yr. Detailed evapotranspiration data on the alluvial fans will help to confirm this approximation.

These conclusions about recharge from precipitation and discharge from evapotranspiration are in general agreement with the assumptions made in previous water-budget studies by C.H. Lee (1912), Los Angeles Department of Water and Power (1972, 1976, 1978, 1979), Hutchison (1986b), and Danskin (1988) and in soil-moisture studies by Groeneveld (1986), Groeneveld and others (1986a, 1986b), and Sorenson and others (1991). All the studies assume that a minimal quantity of recharge occurs from direct precipitation on the valley floor, generally less than 10 percent of the average precipitation rate, and that a somewhat greater potential for recharge from direct precipitation is present on the alluvial fan deposits and volcanic rocks.

An important difference between this study and those done prior to 1983, when the fieldwork and model simulations for this study were begun, is the assumption of a lower infiltration rate from direct precipitation on the alluvial fan and volcanic areas. The lower infiltration rate multiplied by the large size of the affected area results in a substantially lower value of recharge to the saturated ground-water system. This decrease in recharge is matched by a similar decrease in discharge by evapotranspiration from the valley floor. In general, average evapotranspiration rates measured by Duell (1990) and transpiration rates measured by Groeneveld and others (1986a, 1986b) are lower than previous estimates and support the assumption of lower recharge rates from direct precipitation. Because of the recent collection of detailed evapotranspiration data on the valley floor, recharge from direct precipitation on the alluvial fan deposits and volcanic rocks is now the least quantified part of a valleywide ground-water budget. Additional evapotranspiration measurements or soil-moisture studies in these areas would help to confirm present water-budget estimates.

Surface-Water System

The primary source of surface water in the Owens Valley is precipitation that falls on the slopes of the Sierra Nevada. Rivulets from the resulting runoff form tributary streams that flow down mountain canyons, across the alluvial fans, and out onto the valley floor. In the Bishop Basin, the tributary streams are captured by the trunk stream of the valley, the Owens River, which has its headwaters in the Long Valley (fig. 1). In the Owens Lake Basin, approximately 5 mi downstream (south) from the Tinemaha Reservoir, the Los Angeles Department of Water and Power diverts nearly all flow in the Owens River into the Los Angeles Aqueduct. The upstream end of the Los Angeles Aqueduct is referred to as the “intake” (fig. 1). Any water not diverted into the aqueduct continues to flow east of the aqueduct in the natural channel of the lower Owens River. South of the intake, additional tributary streams along the west side of the valley are diverted into the aqueduct. The combined flows of the river–aqueduct system and the diverted tributary streams are routed south out of the valley through the Haiwee Reservoir. Any water

remaining in the lower Owens River flows into the Owens Lake (dry) and evaporates. The entire Owens Valley drainage basin area is shown in figure 1, and photographs of major surface-water features in the Owens Valley are shown in figure 10. The river–aqueduct system, major tributaries, and selected gages within the area of concentrated study are shown in figure 11.

Surface-water monitoring in the Owens Valley is much more complete than in most basins in the United States. More than 600 continuous gaging stations are monitored by the Los Angeles Department of Water and Power in order to measure inflow to the valley from tributary streams and to document water use within the valley. Most of the continuous gages monitor minor flows in canals and ditches in the Bishop area to ensure that sufficient water is delivered to ranching operations. Many of the gages are on the tributary streams and are used to monitor inflow to the valley and to schedule diversions to the river–aqueduct system.

Monitoring of the river–aqueduct system and the lower Owens River is less well documented. Discharge in the river–aqueduct system is gaged routinely at only three locations (the Pleasant Valley Reservoir, the Tinemaha Reservoir, and near the Alabama Hills); discharge in the lower Owens River is gaged routinely at only two locations (immediately below the intake to the aqueduct and at Keeler Bridge) (fig. 11). For other locations, “calculated” discharge values are made by using measured and estimated inflow, outflow, and water use. These calculated values are subject to a large roundoff error as a result of the addition and subtraction of many numbers.

Tributary Streams

Tributary streams provide nearly 50 percent of the surface-water inflow to the Owens Valley; the Owens River and ungaged runoff provide the rest (M.L. Blevins, Los Angeles Department of Water and Power, written commun., 1988; Hollett and others, 1991, tables 2 and 3). Many of the natural channels of tributary streams have been modified by the Los Angeles Department of Water and Power for operation of the river–aqueduct system. Diversion structures have been installed in nearly all streams, and the natural channels of some streams, such as Goodale Creek, have been straightened. Other streams, namely Bishop Creek, Thibaut Creek, Division Creek, and Coldwater Canyon Creek, are diverted to pipes for much of their length (fig. 11). In the Bishop Basin, most of the tributary streamflow that reaches the valley floor is diverted to canals that distribute water for agricultural uses, wildlife habitat, or ground-water recharge. Excess water is returned to the canals and eventually to the Owens River.

Since 1913, little or no tributary streamflow in the Owens Lake Basin has reached the lower Owens River in average-runoff years. During wet years when surface water is abundant, however, tributary streamflow exceeds the capacity of the river–aqueduct system, and some of the tributary streamflow either is diverted onto the alluvial fans to recharge the ground-water system or is conducted in pipes over the top of the aqueduct and then flows across the valley floor toward the lower Owens River.

Tributary streamflow in the Owens Valley is gaged continuously by the Los Angeles Department of Water and Power at more than 60 sites on 34 tributaries. The sites, many constructed originally during prior investigations by the U.S. Geological Survey in the early 1900's (W.T. Lee, 1906; C.H. Lee, 1912), are equipped with concrete channel controls, stilling wells, and automatic data recorders. On most of the tributaries, at least two sites are gaged. Typically, one gage is located near the base of the mountains, and the other is located close to the river–aqueduct system. The location of these gages is shown in figure 11. The station names and abbreviations are given in table 6. A complete record at the sites, except for occasional short gaps, is available for water years 1935–88 (M.L. Blevins, Los Angeles Department of Water and Power, written commun., 1988).

Mean annual discharge for tributaries measured at base-of-mountains gaging stations ranged from 51 to 67,748 acre-ft (Hollett and others, 1991, table 2). Tributaries having the greatest flow include Bishop, Big Pine, Cottonwood, Independence, and Lone Pine Creeks (fig. 11). Mean annual discharge for most streams was about 6,000 acre-ft. Annual flow is highly variable, and maximum and minimum mean annual discharge values for individual streams typically differ by a factor of 10 or more. Although useful as a guide, annual values (Hollett and others, 1991, table 2) tend to mask periods of even higher or lower flows occurring within a single year. Variability in streamflow among tributaries results from differences in size of the drainage basin, quantities of precipitation per basin, and

Figure 10. Major surface-water features in the Owens Valley, California. ***A,*** Owens River just north of Bishop looking west toward the Tungsten Hills and Round Valley (photograph taken winter 1988). ***B***, Los Angeles Aqueduct looking north toward the Sierra Nevada (photograph taken winter 1985). ***C,*** lower Owens River east of the Alabama Hills (photograph taken summer 1988). ***D,*** Owens Lake viewed from alluvial fan south of the Alabama Hills (photograph taken spring 1986).

rates of infiltration. In general, tributary streamflow increases from south to north much as precipitation does (fig. 7).

As expected from precipitation patterns (fig. 7*A*), discharge from tributary streams on the east side of the valley is much less than discharge on the west. Only two streams produce a reliable source of water each year—Coldwater Canyon and Silver Canyon Creeks (fig. 11), and these streams typically discharge less than 2,000 acre-ft/yr. Farther south, Mazourka Creek was monitored by the U.S. Geological Survey continuously during 1961–72 (Mazourka Creek near Independence, USGS station 10282480). Zero flow was recorded all days except during two brief periods in 1967 and 1969. During these periods, discharge peaked at more than 1,300 and 600 ft^3/s, respectively. This type of large, infrequent runoff is characteristic of other basin-and-range valleys (Fenneman, 1931, p. 329) and probably is typical of most stream drainages along the east side of the Owens Valley south of Silver Canyon Creek (fig. 11).

Percent Valleywide Runoff

Total runoff for the Owens Valley is highly correlated with flow in individual tributary streams and has been calculated by the Los Angeles Department of Water and Power (M.L. Blevins, written commun., 1988; table 5) for water years 1935–88. Total runoff is defined as the sum of inflow from the Owens River at the Pleasant Valley Reservoir, measured and estimated inflow from tributary streams, and estimated mountain-

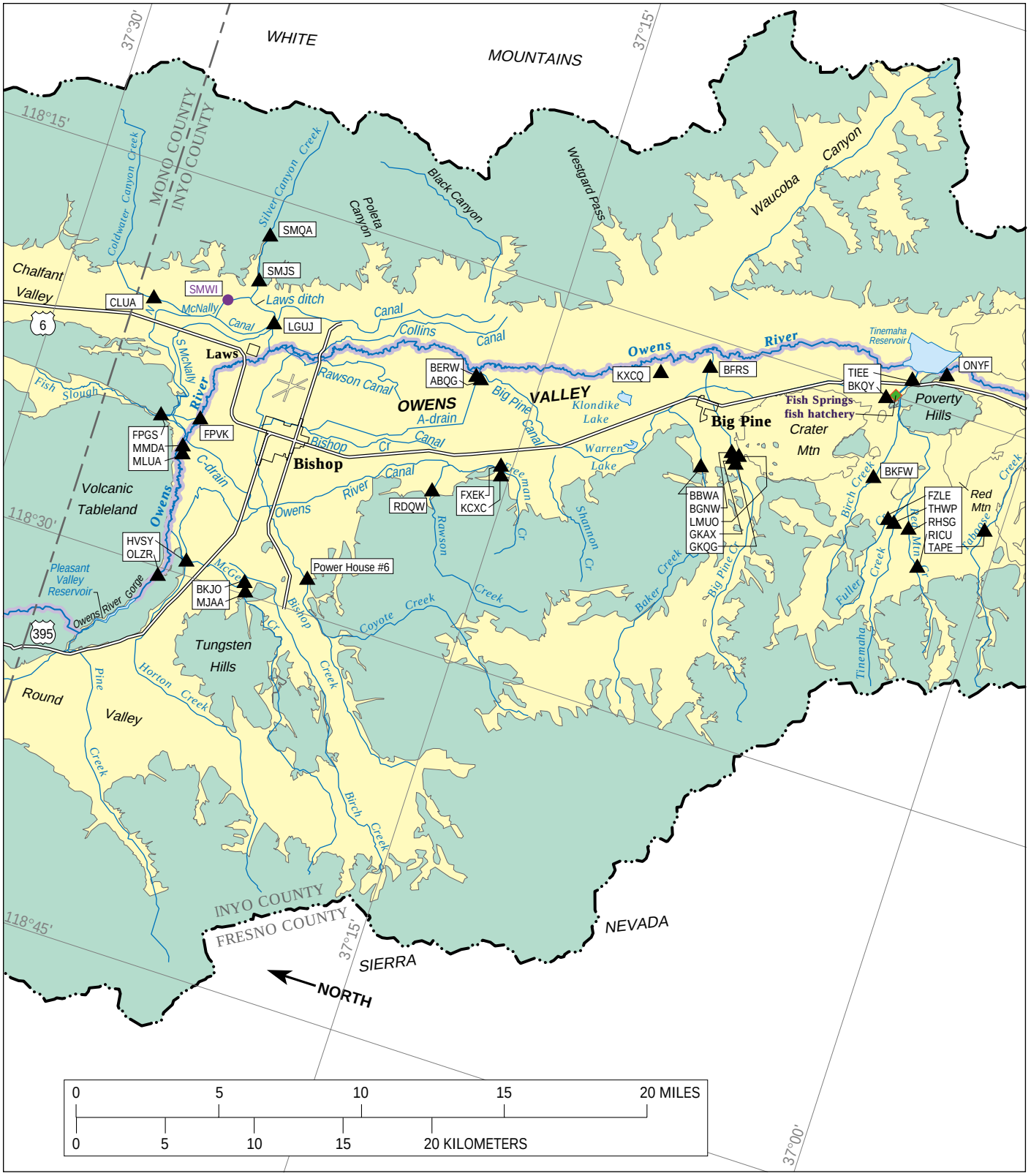


Figure 11. Location of the Owens River–Los Angeles Aqueduct system, the lower Owens River, tributary streams, lakes, reservoirs, spillgates, major gaging stations, and selected pumped wells in the Owens Valley, California.

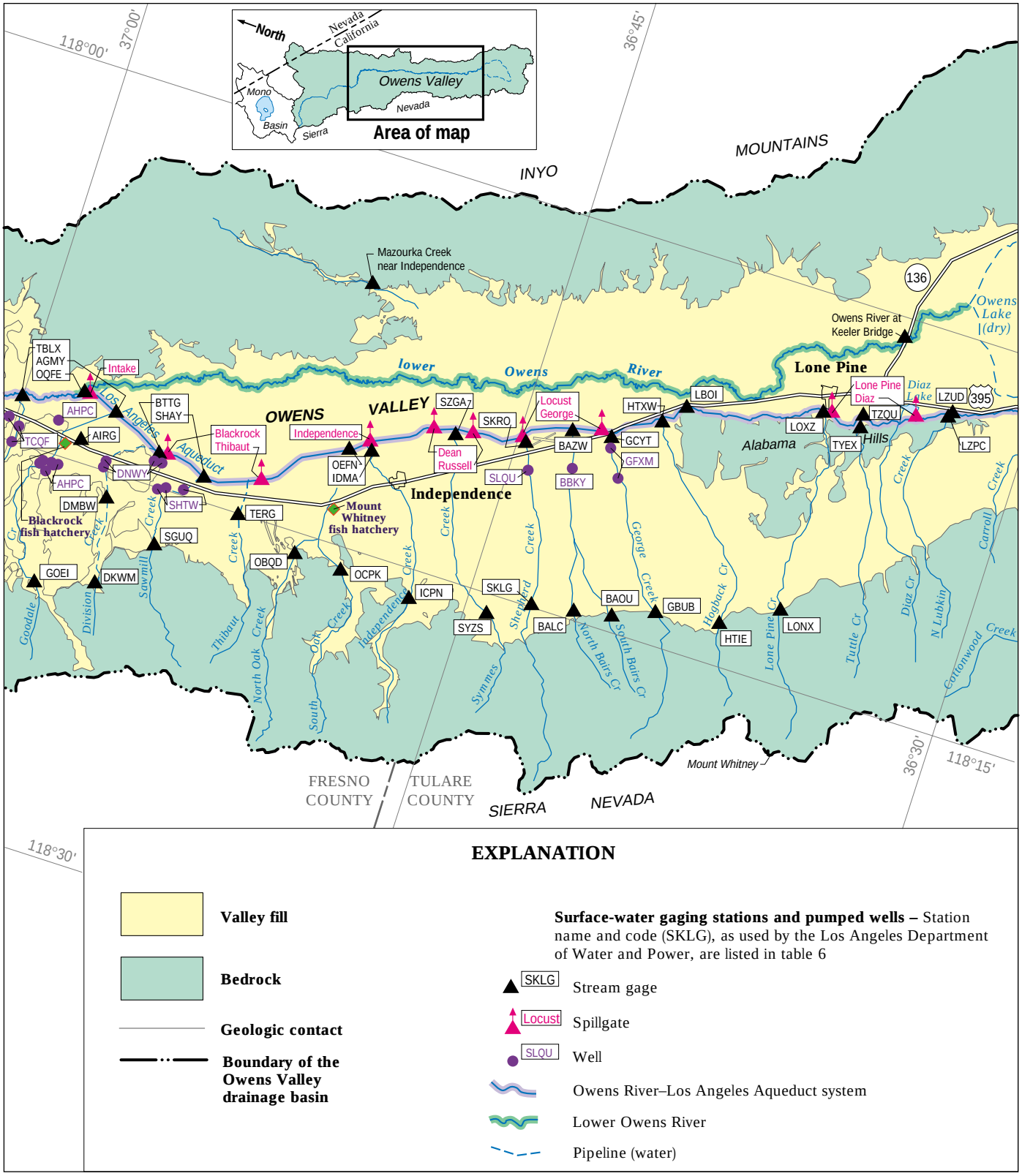


Figure 11. Continued.

front runoff between tributary streams. From annual values of total valleywide runoff, the percent of long-term average annual valleywide runoff for a specific year, referred to locally as the "percent runoff year," is calculated and used extensively by the Los Angeles Department of Water and Power to guide water-

Table 6. Selected surface-water gaging stations and pumped wells in the Owens Valley, California

[Station code and name used by the Los Angeles Department of Water and Power; pumped wells are assigned a station code if well discharge affects a surface-water discharge measurement]

Station code	Station name	Station code	Station name
ABQG	A Drain above Big Pine Canal.	LONX	Lone Pine Creek at base of mountains.
AGMY	Aberdeen Ditch at Los Angeles Aqueduct.	LOXZ	Lone Pine Creek at overhead no. 19.
AHPC	Aberdeen Ditch wells 106, 110–114, 355.	LZPC	Lubkin Creek at Los Angeles Aqueduct.
AIRG	Aberdeen–Blackrock bypass ditch at intake.	LZUD	Lubkin Creek over Los Angeles Aqueduct.
BALC	Bairs Creek (north fork) at base of mountains.	MJAA	McGee Creek at Aberlour Ranch.
BAOU	Bairs Creek (south fork) at base of mountains.	MLUA	South (lower) McNally Canal at O.V.P.A. (Owens Valley Protective Association).
BAZW	Bairs Creek at Los Angeles Aqueduct.	MMDA	North (upper) McNally Canal at O.V.P.A. (Owens Valley Protective Association).
BBKY	Bairs Creek well 353.	OBQD	Oak Creek (north fork) at base of mountains.
BBWA	Baker Creek at Los Angeles Aqueduct Station (4-foot flume).	OCPK	Oak Creek (south fork) at base of mountains.
BERW	Big Pine Canal at intake.	OEFN	Oak Creek at Los Angeles Aqueduct.
BFRS	Big Pine Creek at Cartmell well.	OLZR	Owens River at Pleasant Valley Reservoir, total.
BGNW	Big Pine Creek at U.S. Geological Survey.	ONYF	Owens River at Tinemaha Reservoir.
BKFW	Birch Creek above mill site.	OQFE	Owens River below intake spillgates.
BKJO	Birch Creek at Tungsten City Road.	OUKR	Owens Valley runoff.
BKQY	Birch Creek below highway.	PXHU	Owens River transit loss, Pleasant Valley Reservoir to Tinemaha Reservoir.
BTTG	Blackrock Ditch at Los Angeles Aqueduct.	RDQW	Rawson Creek at base of mountains.
CLUA	Coldwater Canyon Creek at end of pipeline.	RHSG	Red Mountain Creek at Forest Service boundary.
DKWM	Division Creek below intake (overflow).	RICU	Red Mountain Creek diversion above station.
DMBW	Division Creek powerhouse no. 1.	SGUQ	Sawmill Creek at base of mountains.
DNWY	Division Creek wells 108, 109, 351, 356.	SHAY	Sawmill Creek at Los Angeles Aqueduct.
FPGS	Fish Slough at Los Angeles station no. 2.	SHTW	Sawmill Creek wells 155, 159, 339.
FPVK	Fish Slough at Owens River.	SKLG	Shepherd Creek at base of mountains.
FXEK	Freeman Creek at Keough.	SKRO	Shepherd Creek at Los Angeles Aqueduct.
FZLE	Fuller Creek at Forest Service boundary.	SLQU	Shepherd Creek well 345.
GBUB	George Creek at base of mountains.	SMJS	Silver Canyon Creek at base of mountains.
GCYT	George Creek at Los Angeles Aqueduct.	SMQA	Silver Canyon Creek at base of mountains, site no. 2.
GFXM	George Creek wells 76, 343.	SMWI	Silver Canyon Creek at old Clark Ranch (at well 251).
GKAX	Giroux Ditch (lower).	SYZS	Symmes Creek at base of mountains.
GKQG	Giroux Ditch (upper).	SZGA	Symmes Creek at Los Angeles Aqueduct.
GOEI	Goodale Creek at base of mountains.	TAPE	Taboose Creek at base of mountains.
HCKU	North Haiwee Reservoir inflow.	TBLX	Taboose Creek at Owens River.
HTIE	Hogback Creek at base of mountains.	TCQF	Taboose Creek wells 116, 342, 347.
HTXW	Hogback Creek at Los Angeles Aqueduct.	TERG	Thibaut Creek at intake.
HVSY	Horton Creek above Owens River Canal.	THWP	Tinemaha Creek at Forest Service boundary.
ICPN	Independence Creek at Junction Station.	TIEE	Tinemaha Creek at railroad crossing.
IDMA	Independence Creek at Los Angeles Aqueduct	TLRC	Tinemaha Reservoir evaporation, including precipitation.
KCXC	Keough Hot Springs above diversions.	TLYR	Tinemaha Reservoir evaporation pan.
KXCQ	Klondike Drain at Owens River.	TYEX	Tuttle Creek at Canyon Road.
LBOI	Los Angeles Aqueduct at Alabama Gates.	TZQU	Tuttle Creek flow into Los Angeles Aqueduct.
LGUJ	Laws Ditch at railroad.		
LMUO	Little Pine Creek at McMurray Meadows Road.		

management decisions. Values for water years 1935–88 are given in table 7.

Using the percent runoff year for various analyses has two major advantages over other methods: (1) it provides a simple, unifying theme to many complex calculations, and (2) it is relatively independent of the specific method and values used by different individuals and agencies to calculate valleywide runoff. As a result, this key parameter was used extensively in this study, particularly in the analysis of recharge from tributary streams and in the evaluation of selected water-management alternatives.

The probability distribution of the percent runoff year for the Owens Valley for water years 1935–84 is shown in figure 12. This graph and the related best-fit line identify the likely occurrence of a particular percent runoff year. For example, a runoff year having 70 percent or less of the average annual runoff (a 70-percent runoff year) will occur about 15 percent of the time, or about 1 out of 7 years. Water years 1976 and 1977 fall into this category.

The method of developing the probability plot uses the technique of Weibull (1939), as described by Chow (1964, p. 8–28). The 50 annual values for water years 1935–84 (table 7) were assumed to be independent and follow a lognormal distribution. The values were ranked in order (r) and plotted on lognormal probability paper using the relation r/(n + 1), where in this case n equals 50. A general trend line was fitted by hand. Although skewness in the data was recognized (mean equals 100, median equals 94), no other evaluation of the probability distribution was made.

Runoff during the detailed period of analysis chosen for this study, water years 1963–88, slightly exceeded (106 percent) the long-term average runoff. Thus, despite two periods of exceptionally dry conditions (1976–77 and 1987–88) (table 7), the overall period was wetter than normal. In addition, unusually high runoff years—1967, 1969, 1978, 1980, 1982, and 1983—all occurred during this period (fig. 12).

Table 7. Percent of long-term average annual runoff for the Owens Valley, California, water years 1935–88

[Data for station OUKR (table 6) (M.L. Blevins, Los Angeles Department of Water and Power, written commun., 1988). Average runoff (469,604 acre-feet per year equals 100 percent) was calculated for base period, water years 1935–84]

Water year	Percent of average annual runoff	Water year	Percent of average annual runoff
1935	78	1962	94
1936	94	1963	107
1937	110	1964	69
1938	156	1965	96
1939	92	1966	73
1940	94	1967	141
1941	131	1968	80
1942	114	1969	196
1943	108	1970	99
1944	89	1971	79
1945	114	1972	69
1946	111	1973	106
1947	86	1974	107
1948	67	1975	88
1949	70	1976	64
1950	72	1977	55
1951	80	1978	134
1952	132	1979	98
1953	82	1980	142
1954	80	1981	89
1955	77	1982	143
1956	115	1983	189
1957	91	1984	132
1958	122	1985	98
1959	74	1986	158
1960	58	1987	78
1961	53	1988	68

Tributary Stream Recharge

Tributary streams generally lose water as a result of streambed leakage, diversions of streamflow onto the alluvial fans, and, to a lesser extent, evapotranspiration from areas along the stream channel. Several streams also receive water from pumped wells just upstream from the river–aqueduct site (fig. 11), and a few streams receive water from springs, canals, or diversions from other streams. Some streams may gain water in lower reaches because of local seepage of ground water caused by faults, shallow bedrock, or changes in the hydraulic characteristics of the depositional material. Although discharge at the base-of-mountains and river–aqueduct sites is gaged continuously and pumpage from wells is metered, other gains to or losses from tributary streams generally are not measured or are not measured continuously.

The basic technique used to estimate tributary stream recharge is similar to that of C.H. Lee (1912) and uses the following general equation:

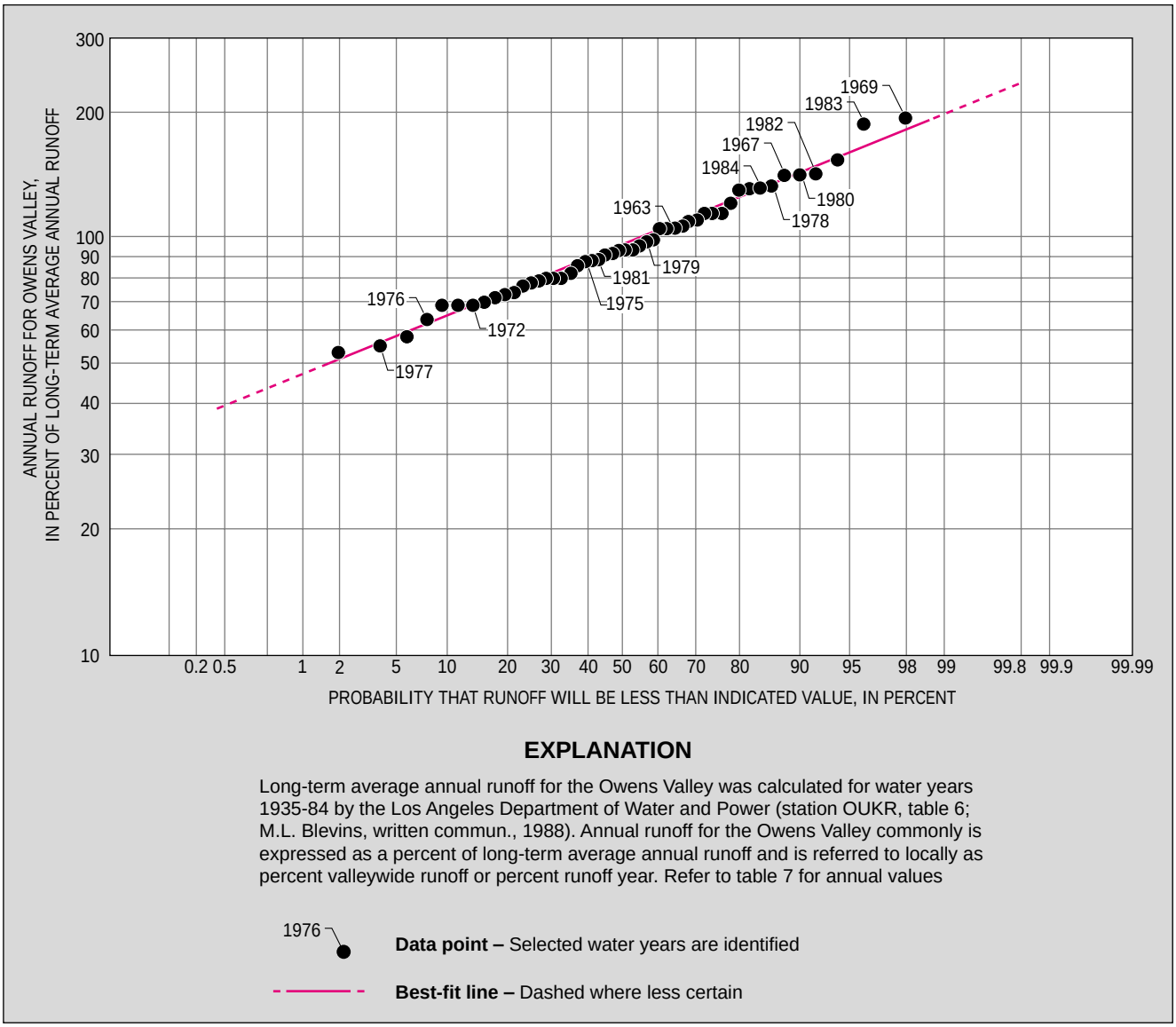


Figure 12. Annual-runoff probability for the Owens Valley, California.

$$R^G = (S^{BM} - S^{RA}) + W^G - ET^G \quad , \qquad (4)$$

where

R^G is stream recharge to the aquifer system for the reach between the base-of-mountains and river–aqueduct gages, in acre-feet per year;

S^{BM} is measured stream discharge at the base-of-mountains gage, in acre-feet per year;

S^{RA} is measured stream discharge at the river–aqueduct gage, in acre-feet per year;

W^G is measured well discharge that flows into the stream between the base-of-mountains and river–aqueduct gages, in acre-feet per year; and

ET^G is the estimated evapotranspiration between the two gages in the immediate vicinity of the stream channel, in acre-feet per year.

Streamflow data for a 50-year period, water years 1935–84, were used to determine the loss for each tributary stream, defined as the sum of R^G and ET^G. Because all other values in equation 4 are

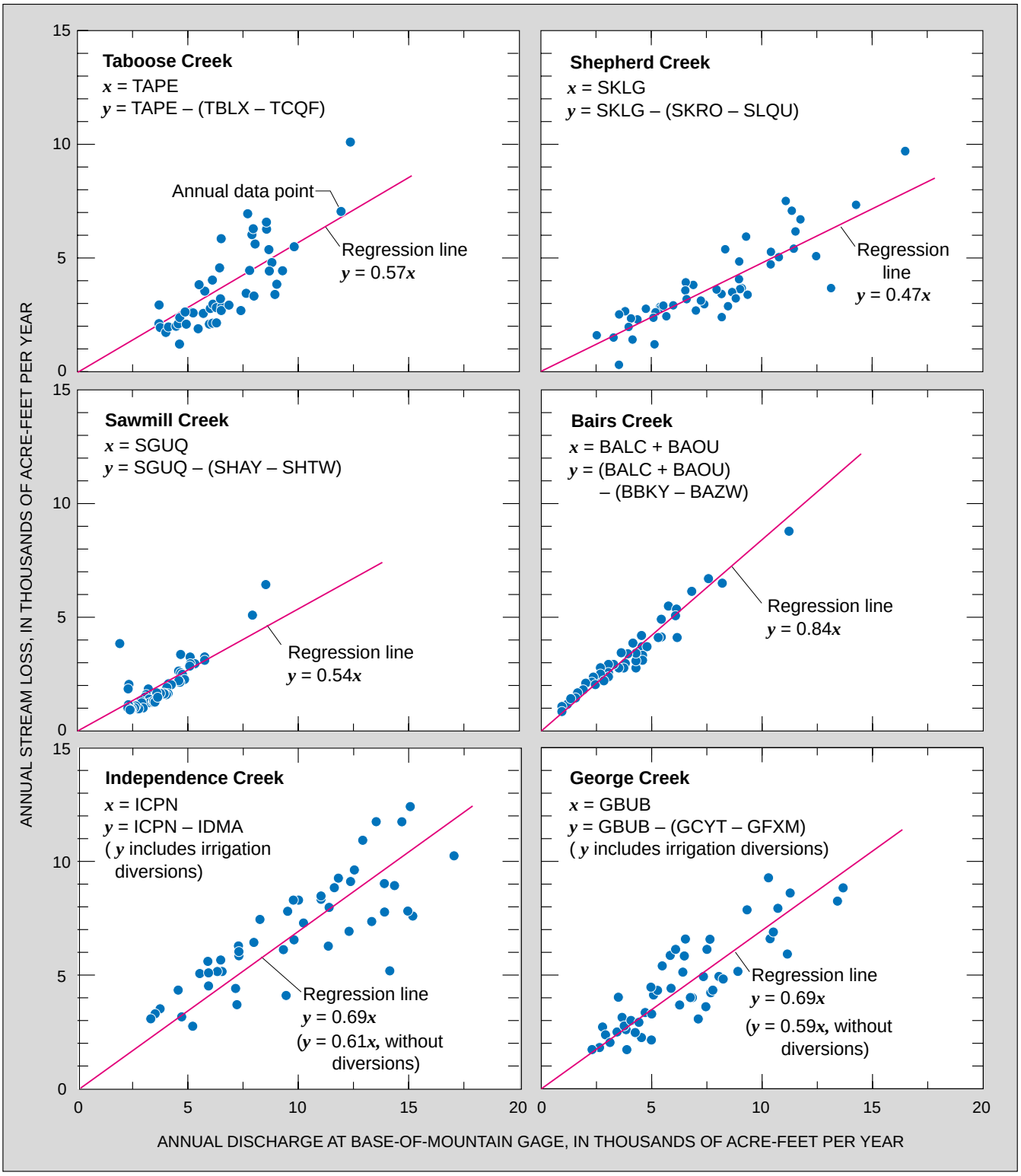


Figure 13. Streamflow relations for selected tributary streams in the Owens Valley, California. Annual data are for water years 1935–84. Station codes, such as TAPE, are shown in figure 11 and described in table 6.

measured, the quantity of stream loss between the base-of-mountains and river–aqueduct gages is well documented. As shown in figure 13, stream loss for each stream is fairly predictable if the quantity of discharge at the base-of-mountains gage (S^{BM}) is known. From the regression equation for each stream (fig. 13), the quantity of stream loss between the gages can be calculated for any known or estimated discharge at the base-of-mountains gage. Similar graphical relations were evaluated, and linear regression equations were developed, for each of the 34 tributary streams using data from the discharge gages identified in figure 11 and listed in table 6.

The average stream loss rates (coefficient *a* in the regression equations in figure 13 with the general form $y = ax$) calculated from the 50 years of discharge data generally are higher than those reported by C.H. Lee (1912, pl. 9), who used about 4 years of record. The cause of the increase is not known, but it may result from the slightly greater length of the gaged section, additional diversions of water from the streams, or changes to the channels.

Tributary stream recharge between the gages (R^G) was calculated from stream loss by estimating evapotranspiration for each stream using the equation,

$$ET^G = \frac{ET^O SL_i^G SW^G SV^G}{43,560}, \qquad (5)$$

where

ET^G is estimated evapotranspiration between the two gages in the immediate vicinity of the stream channel, in acre-feet per year;

ET^O is the average annual evapotranspiration rate for high-water-use species, in feet per year;

SL^G is the length of the stream channel between the two gages, in feet;

SW^G is the width of vegetation near the stream channel, in feet; and

SV^G is the percent of vegetative cover near the stream, expressed as a decimal fraction.

Because detailed data were not available for most variables in equation 5, estimates were made on the basis of limited field observations of Bishop, Independence, Oak, Taboose, and Lone Pine Creeks, and measurements of vegetative conditions on the valley floor (table 5) (D.P. Groeneveld, Inyo County Water Department, written commun., 1986; Duell, 1990). Constant values were chosen for SW^G (50 ft), ET^O (47 in/yr), and SV^G (0.30). Stream length was measured by digitizing 1:24,000-scale topographic maps. For each of the tributary streams, evapotranspiration was found to be minimal, ranging from about 10 to less than 100 acre-ft/yr (Hollett and others, 1991, table 8). This quantity generally is less than about 2 percent of the discharge at the base-of-mountains gage and less than about 5 percent of the estimated recharge between the two gages.

For selected water years, such as the ground-water simulation period (water years 1963–88), annual discharge at each base-of-mountains gage was estimated by multiplying the 50-year average discharge at the base-of-mountains gage (water years 1935–84) by the percent runoff year for individual years (table 7). Recharge above or below the gaged section of the stream was determined from gaged records of diversions and by comparing respective lengths of stream channels in the gaged and ungaged sections. The relation for total recharge for a stream *(i)* in water year *(j)* can be expressed as:

$$R_{ij}^T = R_{ij}^G + R_{ij}^A + R_{ij}^B, \qquad (6)$$

where

R^T is the total stream recharge between the surrounding bedrock and the river–aqueduct system, in acre-feet per year;

R^G is stream recharge that occurs between the base-of-mountains and river–aqueduct gages, in acre-feet per year;

R^A is the stream recharge that occurs above the base-of-mountains gage, in acre-feet per year; and

R^B is the stream recharge that occurs below the river–aqueduct gage, in acre-feet per year.

Within the gaged section of a specific stream (*i*), stream loss during a particular year (*j*) can be estimated as,

$$SLQ_{ij}^G = SLR_i^G [S_i^{BM} RO_j], \qquad (7a)$$

and stream recharge estimated as,

$$R_{ij}^G = SL_{ij}^G - ET_i^G, \qquad (7b)$$

where

SLQ^G is the quantity of water lost from the stream between the base-of-mountains and river–aqueduct gages, in acre-feet per year;

SLR^G is the average loss rate (a), determined from the regression equation $y = ax$ (fig. 13) expressed as a decimal fraction;

S^{BM} is the long-term mean annual discharge at the base-of-mountains gage (Hollett and others, 1991, table 2), in acre-feet per year;

RO is the percent runoff year (table 7), expressed as a decimal fraction; and

ET^G is estimated evapotranspiration between the two gages in the immediate vicinity of the stream channel, in acre-feet per year.

For most streams with standard channels,

$$R_{ij}^{A} = R_{ij}^{G}\left[\frac{SL_i^A}{SL_i^G}\right], \qquad (8a)$$

and

$$R_{ij}^{B} = R_{ij}^{G}\left[\frac{SL_i^B}{SL_i^G}\right], \qquad (8b)$$

where

SL^A is stream length above the base-of-mountains gage, in feet;

SL^G is the stream length between the base-of-mountains and river–aqueduct gages, in feet; and

SL^B is stream length below the river–aqueduct gage, in feet.

From these relations, total recharge for each stream can be estimated both for historical periods and for hypothetical situations, such as those evaluated as possible water-management alternatives.

Several of the tributary streams could not be evaluated using this approach because only a single gaging station was operated on the stream, because unquantified diversions were made from one stream to another, or because a spring between the two gages added an unknown quantity of water to the stream. In these cases, an average recharge rate per foot of stream channel was calculated for streams with two gages (Hollett and others, 1991, table 8). These recharge rates were applied to streams that have similar annual discharge rates and that flow over similar types of materials.

For a few streams, the long length of channel above the base-of-mountains gage *(SL^A)*, such as for Independence Creek (fig. 11), produced an unrealistically high quantity of recharge, indicating that the stream may have been flowing on top of a narrow, fully saturated, alluvial fan or glacial deposit that was not capable of receiving additional water from the stream. For these sections of streams, recharge estimates were scaled downward on the basis of a shorter recharge length for the stream and on recharge values for similar nearby streams. Diversion of flow from Big Pine Creek and Oak Creek for domestic use and irrigation on nearby Indian reservations decreased recharge rates for those streams in comparison with the total loss rate calculated from equation 4. Using these methods, the average annual recharge for all tributary streams within the area of the defined aquifer system (fig. 2) was estimated to be 106,000 acre-ft/yr for water years 1963–69 and 103,000 acre-ft/yr for water years 1970–84.

Ungaged Runoff

Mountain-Front Runoff Between Tributary Streams

Most runoff from precipitation falling on the mountains surrounding the Owens Valley is measured at the base-of-mountains gaging stations on the major tributary streams (fig. 11). Some runoff, however, occurs from precipitation falling on ungaged drainage areas between gaged tributary streams. Precipitation in these small, triangular-shaped areas—commonly referred to as intermountain slopes (C.H. Lee, 1912)—runs off as sheet flow, in rivulets, or in small intermittently flowing streams. The intermountain slopes along the southwest side of the basin were mapped and described by C.H. Lee (1912, p. 13 and pl. 1). Most of the runoff from these areas disappears into the alluvial fans a short distance from the edge of the mountains. This water, referred to as “hidden recharge” by Feth (1964a) because it is not measured, either is transpired by nearby plants or contributes recharge to the ground-water system. The increase in vegetation along the upper part of the alluvial fans observed by M.O. Smith and others (1990a, b) may result not only from increased precipitation, related to the increase in altitude (fig. 7*B*), but also from runoff between tributary streams.

The abundance of springs in many bedrock areas along both sides of the valley (shown on USGS 1:62,500-scale topographic maps) indicates that the quantity of water contributed to the basin might be significant. For example, discharge from Scotty Springs near Division Creek (Mt. Pinchot quadrangle) has been measured at greater than 2 ft^3/s (C.H. Lee,

1912, p. 44). Except for spring discharge, the total quantity of ungaged surface-water inflow is difficult or impossible to measure.

Instead, estimates of the quantity of ungaged surface-water inflow and resulting ground-water recharge typically are made using precipitation records, runoff coefficients calculated for gaged drainage areas, and assumptions about the percentage of runoff that percolates to the ground-water system. Using this approach in the southwestern part of the Owens Valley, C.H. Lee (1912, p. 66–67 and table 61) estimated that as much as 75 percent of the total volume of precipitation on the ungaged drainage areas recharged the ground-water system. Lee noted that the high rate resulted from steep mountain slopes and rapid melting of snow, both of which minimize losses from evapotranspiration and percolation through the extremely transmissive alluvial fan deposits.

In the present study, recharge for each of the ungaged drainage areas was estimated in a similar manner, but using different percolation rates depending on the part of the valley being analyzed. Recharge for each area along the southwest side of the valley was calculated using the average annual precipitation from figure 7 and the 75-percent percolation rate suggested by C.H. Lee (1912). Recharge for areas along the northwest side of the valley was somewhat less because of smaller drainage areas, lower precipitation values, or an abundance of mountain meadows that discharge the ungaged water as evapotranspiration before it can reach the valley ground-water system. Recharge for the Volcanic Tableland was significantly less than for areas on the west side of the valley because precipitation rates are much lower (fig. 7), potential evaporation is much higher because of the higher average temperature, and percolation is restricted by the impermeable capping member of the Bishop Tuff (figs. 4 and 5). Recharge for areas on the east side of the basin was almost zero because virtually no runoff has been observed between the intermittently flowing tributary streams, particularly those south of Coldwater Canyon Creek (figs. 3 and 11).

A few of the larger ungaged streams flow far enough down the alluvial fans to join a major tributary stream below the base-of-mountains gage (fig. 3). This addition of water to the gaged tributaries is not accounted for in the estimates of tributary streamflow or tributary stream recharge described earlier in the section “Tributary Streams.” This recharge, however, is accounted for using the method described above for ungaged runoff.

Recharge to the defined aquifer system (fig. 2) contributed from all ungaged areas was estimated to average approximately 26,000 acre-ft/yr for both water years 1963–69 and water years 1970–84. In order to estimate ungaged recharge for different water years, the long-term average recharge rates were multiplied by the annual percent of valleywide runoff (table 7). Although a high degree of uncertainty is associated with the values of recharge between tributary streams, recharge from ungaged areas for most of the valley is a relatively small component of the ground-water budget. Significant refinement in the quantity of runoff or ground-water recharge is unlikely because of the difficulty of measurement. However, a comprehensive surface-water/ground-water budget for the entire valley, as suggested by Danskin (1988), might improve the confidence limits for ungaged runoff and the related ground-water recharge.

Runoff from Bedrock Outcrops Within the Valley Fill

A small quantity of precipitation falls on the bedrock outcrops within the valley fill, in particular on the Tungsten Hills, the Poverty Hills, and the Alabama Hills (fig. 7). Most of the precipitation probably is evaporated or transpired by the sparse native vegetation covering the hills. Some runoff can occur during longer duration, high-intensity storms. This quantity is not important either for local uses or for export from the valley.

Springs visible on the north and west sides of the Alabama Hills (Lone Pine and Union Wash quadrangles, USGS 1:24,000-scale topographic maps) indicate that precipitation does exceed evapotranspiration and that some local infiltration occurs into the soil and fractured rocks. During longer duration storms, some recharge to the ground-water system in the immediate vicinity of the bedrock outcrops probably occurs. Also, some additional recharge probably occurs from the minor spring discharges along the sides of the bedrock outcrops. A likely range of recharge values was determined using estimates of average precipitation (fig. 7) and a range of possible runoff coefficients (C.H. Lee, 1912). The total quantity of recharge to the aquifer system (fig. 2) from runoff from bedrock outcrops for average conditions of precipitation and evaporation probably is less than 1,000 acre-ft/yr.

Table 8. Mean annual discharge at selected gaging stations on the Owens River–Los Angeles Aqueduct system in the Owens Valley, California.

[—, not available. Measured discharge data in acre-feet per year from the Los Angeles Department of Water and Power (M.L. Belvins, written commun., 1988). Values for the Los Angeles Aqueduct at the North Haiwee Reservoir are estimates]

Station name	Station code (table 6)	Water years			
		1935–69	1945–69	1953–69	1970–84
Owens River at the Pleasant Valley Reservoir.	OLZR	250,000	260,000	260,000	330,000
Owens River at the Tinemaha Reservoir.	ONYF	—	—	320,000	390,000
Los Angeles Aqueduct at the Alabama Gates.	LBOI	—	320,000	330,000	450,000
Los Angeles Aqueduct at the North Haiwee Reservoir.	HCKU	320,000	340,000	350,000	480,000

Owens River and the Los Angeles Aqueduct

The river–aqueduct system within the study area extends from the Mono Basin to the Haiwee Reservoir (fig. 1). At the northernmost point of the river–aqueduct system in the Mono Basin, streams flowing out of the Sierra Nevada are diverted into a concrete-box conduit. The diverted water is routed to Grant Lake in the Mono Basin and eventually is conveyed to the Owens River in the Long Valley through the 11.3-mile-long Mono Craters Tunnel (fig. 1). The mean annual discharge through the tunnel is about 72,000 acre-ft. At the end of the Mono Craters Tunnel, water from the Mono Basin joins the upper reach of the Owens River and together flows about 12 mi to Lake Crowley, also known as the Long Valley Reservoir. Lake Crowley, which is the largest reservoir in the river–aqueduct system, regulates the flow of water through a 96- to 108-inch pipeline (penstock) that connects Lake Crowley in the Long Valley with the Pleasant Valley Reservoir in the Owens Valley. The natural channel of the Owens River through the Volcanic Tableland is used infrequently to convey floodwaters or to divert water during maintenance of the pipeline. Three hydroelectric plants located along the pipeline generate electricity as a result of a drop in altitude of about 1,600 ft from the Long Valley to the Owens Valley. The mean annual discharge of the Owens River at the Pleasant Valley Reservoir increased from about 250,000 acre-ft for water years 1935–69 to about 330,000 acre-ft for water years 1970–84 (table 8). This increase resulted from additional diversion of water from the Mono Basin, as well as from greater runoff during the latter, wetter period (106 percent runoff in comparison with 97 percent).

The Pleasant Valley Reservoir regulates flow to the natural channel of the Owens River downstream from the outlet tower at the Pleasant Valley Dam. Between the Pleasant Valley Reservoir and the Haiwee Reservoir at the south end of the Owens Valley, discharge in the river–aqueduct system is constantly altered by gains of water from streams, springs, pumped wells, flowing wells, and seepage from the ground-water system, as well as by losses of water to irrigation and to the ground-water system. Emerging from the Pleasant Valley Reservoir, the Owens River continues south, gaining water primarily from tributary streams and from pumped and flowing wells before discharging into the Tinemaha Reservoir at the south end of the Bishop Basin. A photograph (fig. 10*A*) taken just north of Bishop near the Five Bridges area (Fish Slough quadrangle, USGS 1:24,000-scale topographic map) shows the general character of the Owens River in the Bishop Basin. The natural, meandering channel of the Owens River is generally about 20 to 50 ft wide and about 3 to 6 ft deep, and has a silt, sand, and clay bottom. The mean annual discharge of the Owens River at the Tinemaha Reservoir was about 390,000 acre-ft for water years 1970–84, or about 60,000 acre-ft/yr greater than the discharge at the north end of the Bishop Basin at the Pleasant Valley Reservoir (table 8).

Flow in the Owens River resumes south of the Tinemaha Reservoir and continues for approximately 5 mi until virtually all water is diverted into the unlined, trapezoidal channel of the Los Angeles Aqueduct (fig. 10*B*). Flowing along the toes of the western alluvial fans, the aqueduct gains additional water from streams and wells. In the Owens Lake Basin, tributary streams are generally smaller, although

more numerous than in the Bishop Basin, and there are fewer diversions for agricultural uses. At the Alabama Gates (fig. 11), on the north side of the Alabama Hills, the aqueduct changes to a concrete-lined channel. The mean annual discharge at the Alabama Gates was about 450,000 acre-ft for water years 1970–84, or about 60,000 acre-ft/yr greater than the discharge at the Tinemaha Reservoir (table 8). At the Haiwee Reservoir at the southern boundary of the study area, mean annual discharge is about 1.5 times mean annual discharge at the Pleasant Valley Reservoir (table 8). The Haiwee Reservoir regulates and temporarily stores water before releasing it into the two channels of the dual-aqueduct system that conveys the water to the Los Angeles area. After completion of the second aqueduct, discharge to Los Angeles increased approximately 160,000 acre-ft/yr both as a result of changes in management practices and greater average runoff (tables 4, 7, and 8).

Since the early 1900's, successive changes in water management have altered the role of the Owens River in the Owens Valley hydrologic system. Prior to development of the river–aqueduct system, the natural channel of the Owens River was the primary drain of both the surface-water and ground-water systems. Tributary streams flowed across the valley floor to merge with the river, and ground water flowed upward under pressure to augment discharge in the perennially flowing Owens River. After operation of the Los Angeles Aqueduct was begun in 1913, the hydrologic system of the valley remained dominated by the Owens River in the Bishop Basin, but the system became dominated by the Los Angeles Aqueduct in the Owens Lake Basin. The diversion of tributary streams at the edge of alluvial fans into the aqueduct prevented the lower Owens River from acting as a major surface-water collector. The river–aqueduct system drained the surface-water system, and the Owens River in the Bishop Basin and the lower Owens River in the Owens Lake Basin drained the ground-water system.

After 1970, increased ground-water pumping began to change these conditions. What had been a relatively simple hydrologic system began the transition to a more complex system with dynamically changing surface-water/ground-water interactions. In at least one area of the valley near Big Pine, the Owens River began losing water to the ground-water system. Water-level data collected from nearby wells show a hydraulic gradient from the Owens River to production wells along the edge of Crater Mountain (fig. 11). In other parts of the valley with high ground-water pumpage, such as near Laws, the quantity of water gained by the Owens River from the ground-water system probably was reduced.

The Los Angeles Aqueduct, because it is elevated topographically above the center line of the valley, never acted as a major ground-water collector. However, for most of its unlined length, the aqueduct is at an altitude at which it can exchange water readily with the ground-water system. The local hydraulic gradient between the aqueduct and the ground-water system, as described above for the Owens River, determines the direction and rate of flow. Hydrogeologic sections developed by Hollett and others (1991, pl. 2), Griepentrog and Groeneveld (1981), and the Los Angeles Department of Water and Power (1978) indicate the general areas where the aqueduct gains or loses water for different ground-water conditions. Under average conditions, most sections of the aqueduct continue to gain water from the ground-water system. However, during periods of significant ground-water withdrawals, such as 1971–74, ground-water levels near the aqueduct decline and the rate of gain decreases; the decline can be sufficient to change the direction of flow, resulting in a loss of water from the aqueduct. This condition likely occurred in areas with numerous production wells, such as between Taboose and Thibaut Creeks (fig. 11). South of George Creek, the altitude of the aqueduct is generally above even the highest ground-water levels; therefore, the aqueduct loses water to the ground-water system. The concrete-lined section of the aqueduct adjacent to the Alabama Hills also is elevated above the nearby ground-water system and has the potential to lose water; however, the loss through the concrete and related joints probably is minimal.

Estimates of the quantity of loss (or gain) for the river–aqueduct system typically are calculated as the residual of a mass balance for a gaged section of the stream. This is the same method used to calculate recharge for the tributary streams. When the loss is a small fraction of the measured flows, however, large residual errors can result, masking the actual loss or gain. For this reason, estimates of the likely range of loss or gain for the river and aqueduct were developed using loss studies on canals that flow over similar materials, but have a much smaller discharge.

Analysis of several canals in the Laws area indicates that a 15-foot-wide canal with a mean discharge of 2 to 10 ft^3/s typically loses 0.3 to

1.1 (ft^3/s)/mi (R.H. Rawson, Los Angeles Department of Water and Power, oral commun., 1988). Similar loss rates were calculated for tributary streams (Hollett and others, 1991, table 8). If vertical conductivity for the canals, river, and aqueduct are similar, then these rates equate to approximately 1 to 3 (ft^3/s)/mi for the wider Owens River or the Los Angeles Aqueduct. Because the rate of exchange (either loss or gain) between the river or aqueduct and the ground-water system is dependent on the physical characteristics of the stream channel, which are fairly constant, and on the local hydraulic gradient between the stream and the ground-water system, which generally varies over a small range of values, the exchange rates probably are similar for both the gaining and losing reaches of the river and aqueduct.

If bed material of the river–aqueduct system is finer grained than bed material of the tributary streams and selected canals, the exchange rates probably are less for the river–aqueduct than for streams or canals. To accommodate this uncertainty, ground-water recharge or discharge (river–aqueduct loss or gain) was determined by applying a range of estimated rates of gain or loss to the respective gaining or losing sections of the river–aqueduct system and then comparing these values with results from the valleywide ground-water flow model. For the area of the aquifer system (fig. 4), the river–aqueduct system during water years 1963–69 and water years 1970–84 was estimated to gain approximately 16,000 acre-ft/yr and 3,000 acre-ft/yr, respectively.

As part of an extensive surface-water monitoring network, the Los Angeles Department of Water and Power computes mass balances for various sections of the river–aqueduct system. These calculations are given stations identifiers, such as those in table 6, and are listed in a monthly report, “Uses and Losses” (L. Lund, Los Angeles Department of Water and Power, written commun., 1988). The mass-balance values for several years suggest that the Owens River gains about 33,000 acre-ft/yr from the ground-water system between the Pleasant Valley Reservoir and the Tinemaha Reservoir (station PXHU, table 6). This value is equivalent to a rate of gain of about 1.5 (ft^3/s)/mi of river channel. Although this value is physically realistic, the calculated gain for the river–aqueduct system in this reach is much higher than the values estimated using the technique described above or values derived from the ground-water flow model described later. A detailed water budget linking the surface-water and ground-water systems as suggested by Danskin (1988), or development of a surface-water/ground-water model, might help solve this discrepancy.

The specific interactions of the river–aqueduct system with the ground-water system are difficult to measure or estimate. Further improvements in knowledge may require taking advantage of water-quality and temperature measurements of the river–aqueduct and of ground water. These analyses may be useful in confirming concepts and quantities of interactions that are less clearly defined by water-use calculations and water-level mapping, particularly in the complex water-distribution area near Bishop (fig. 3).

Spillgates.—Ten spillgates are located along the aqueduct and are used at various times throughout the year to clean the aqueduct of debris and, during high-runoff years, to discharge excess water onto the valley floor. Discharge from the spillgates is measured and is relatively constant in average-runoff years. During most years, total discharge from the 10 spillgates averages about 22,000 acre-ft/yr, but during high-runoff years such as 1967, 1969, and 1983 (fig. 12), total discharge can be several times that quantity. Nine spillgates are shown in figure 11; an additional spillgate is located near Cottonwood Creek, just south of the focused area of study. The Cottonwood spillgate was not included in the analysis presented in this report.

Some ground-water recharge occurs as a result of discharge from the spillgates. Although the quantity of discharge is measured, the quantity that infiltrates to the ground-water system is not known. Some of the discharge, especially in high-runoff years, may flow across the valley floor to the channel of the lower Owens River. In a regression analysis of discharge in the lower Owens River, Hutchison (1986d) attributed much of the measured discharge in the lower Owens River at Keeler Bridge (fig. 11) to releases from the spillgates.

Discharge of surface water from the spillgates is limited to some extent by litigation (*Natural Soda Products Co.* v. *Los Angeles*, 23 California 193) that restricts discharge to the Owens Lake (dry). Occasional wetting of the dry lakebed is believed to contribute to air-quality degradation in the valley caused by dust storms (Saint-Amand and others, 1986; Lopes, 1988). In high-runoff years, these restrictions are difficult or impossible to meet because of the large quantity of water in the valley and the limited capacity of the river–aqueduct system. For example, in the exceptionally wet

water years 1969 and 1983 (fig. 12), there was water, quite literally, everywhere in the valley and the spillgates were used extensively. Surface water that could not be exported out of the valley was diverted onto the valley floor, primarily through the Blackrock spillgate (fig. 11).

During such exceptionally-high-runoff years, infiltration into the unsaturated zone and recharge to the underlying water table may be so great that the infiltration restores the unsaturated zone to field capacity and the recharge reequilibrates shallow ground-water levels from any previous decline caused by nearby pumping or drought. Massive releases from the several spillgates likely play an important role in doing this. Areas of the valley that historically have been inundated with water during high-runoff years are shown on maps compiled by Boyle Engineering and by the Los Angeles Department of Water and Power (M.L. Blevins, written commun., 1986) for 1952, 1967, and 1969.

In this present study, the quantity of infiltration from spillgates was estimated by subtracting the likely losses from evapotranspiration and an estimate of the return flow to the lower Owens River from the measured discharge. Because the discharge channels were observed to have a greater abundance of vegetation than nearby areas on the valley floor, a relatively high evapotranspiration rate of 40 in/yr (Duell, 1990) was used in the calculations. The total recharge to the defined aquifer system (fig. 4) from spillgates was estimated to average approximately 6,000 acre-ft/yr.

Lower Owens River

Prior to substantial surface-water diversions in 1913, both surface and ground water migrated to the lower Owens River and eventually discharged into the Owens Lake. As of 1988, nearly all water flowing out of the Tinemaha Reservoir is diverted into the river–aqueduct system, and the lower Owens River has become relatively isolated from other surface-water features of the valley. A photograph of the lower Owens River (fig. 10*C*) taken in summer 1988 shows an abundance of riparian vegetation, especially bulrush and cattails, within the river channel. Typically, the riverbed itself is moist almost to the land surface. Although in some places the lower Owens River has flowing water that continues for several hundred feet, most of the river channel is occupied by this type of riparian vegetation (fig. 3).

In average-runoff years, most discharge reaching the Owens Lake (dry) via the lower Owens River is surface water returned to the river from ditches and undiverted tributary streamflow or ground water that seeps into the river channel (Hutchison, 1986d). During extremely wet years, runoff exceeds the capacity of the river–aqueduct system and not all flow in the Owens River is diverted into the Los Angeles Aqueduct. For example, annual discharge in the lower Owens River measured just below the aqueduct intake (station OQFE, table 6; fig. 11) for water years 1945–84 was typically 0 acre-ft, but annual discharge for water years 1969 and 1983 exceeded 75,000 acre-ft (L. Lund, Los Angeles Department of Water and Power, written commun., 1988).

Discharge in the lower Owens River also is measured continuously at the Keeler Bridge east of Lone Pine (fig. 11). For water years 1927–86, mean annual discharge was about 17,000 acre-ft (Hollett and others, 1991, table 3). Using regression techniques, Hutchison (1986d) evaluated the river-discharge record at the Keeler Bridge for runoff years 1946–86 and concluded that most streamflow at the bridge resulted either from operational releases to the river from the river–aqueduct system or from ground-water discharge. He noted that ground-water discharge in the lower Owens River was affected significantly by bank storage. Sediment along the bank of the river becomes saturated with river water as stage of the river rises, and the stored water then is gradually released back to the river as stage of the river falls. This hydraulic buffering dampens fluctuations in stage and discharge. By separating the various components of discharge, Hutchison (1986d) estimated that the ground-water contributions to the lower Owens River for runoff years 1946–86 ranged from 3,000 to 11,000 acre-ft/yr and averaged about 3,600 acre-ft/yr.

In years of much greater than average runoff (fig. 12 and table 7), the lower Owens River probably changes from a gaining stream to a losing stream, thereby recharging the nearby ground-water system, particularly on the east side of the valley. This change is most likely a temporary one; water that is lost will be regained by the river over the next few months or couple of years as the stage in the river channel returns to almost zero. This is essentially the same bank-storage process noted by Hutchison (1986d).

In order to more accurately identify interaction of the lower Owens River with the ground-water system, the Los Angeles Department of Water and

Power measured instantaneous discharge during 1986–87 at 10 sites along the river from the aqueduct intake to the Keeler Bridge (Hollett and others, 1991, fig. 22). River reaches between the measurement sites were defined as either gaining- or losing-water reaches—although only three of the reaches were found to act in a consistent manner during the period of observations. The first section, a few miles south of the aqueduct intake (Hollett and others, 1991, fig. 22), generally lost water to the ground-water system. As discussed in later sections of this report, this loss may correlate with pumpage from wells between Taboose and Thibaut Creeks (fig. 11). Gaining reaches near Independence and Lone Pine may result from abundant recharge in the vicinity of Oak Creek, discharge from spillgates (fig. 11), and a fining of aquifer materials near Lone Pine. Some of the water gained by the river is discharged as evapotranspiration by the abundant riparian vegetation in the natural channel of the lower Owens River (fig. 10*C*).

Areas surrounding the lower Owens River are shown as having transpiration values ranging from about 0.5 to 1.5 ft/yr (fig. 9). These intermediate values are attributed to transpiration by riparian vegetation that has high transpiration rates, often exceeding 3.5 ft/yr (D.P. Groeneveld, Inyo County Water Department, written commun., 1984), mixed with other native vegetation that has lower rates (table 5). In the immediate vicinity of the lower Owens River, transpiration from dense riparian vegetation, such as occupies the river channel (figs. 3 and 10*C*), probably consumes much of the rising ground water that would otherwise flow down the river.

Reservoirs and Small Lakes

Reservoirs

The Pleasant Valley and the Tinemaha Reservoirs are impounded by earth-filled dams and are used to regulate flow in the river–aqueduct system (fig. 11). The Pleasant Valley Reservoir is at the mouth of the Owens River gorge, which cuts deeply through the Volcanic Tableland. Nearly all water that normally flowed through the gorge has been diverted into a 96- to 108-inch pipeline (penstock) that passes through three power-generation plants. Water is discharged from the third power plant into the adjacent reservoir, which is about 20 ft deep and covers about 1,700 acres. The reservoir is used primarily as an afterbay for the power-generation facilities and to stabilize flow into the Owens River. Since 1970, when the additional diversions of water from the Mono Basin began, annual inflow to the Pleasant Valley Reservoir has increased by more than 60,000 acre-ft (table 8).

Seepage through the earthen dam that impounds the Pleasant Valley Reservoir undoubtedly occurs although the rate is not known. Any seepage through the dam probably is regained by the Owens River a short distance downstream from the dam. More important, the bottom of the reservoir may contact the more transmissive members of the Bishop Tuff (fig. 5; Hollett and others, 1991). If this contact is present and the normal siltation in the reservoir has not restricted direct hydraulic connection between reservoir water and these well-sorted sands, then significant seepage may occur from the reservoir to the ground-water system.

The Tinemaha Reservoir is at the south end of the Bishop Basin, about 5 mi upstream from the intake to the aqueduct (fig. 11). The reservoir, which was built in 1929, covers between 0 and 16,000 acres depending on runoff during the particular year (table 7) and is less than 25 ft deep. The reservoir is underlain by moderately transmissive fluvial deposits composed primarily of silt, clay, and sand (fig. 4).

Mass-balance calculations for the Tinemaha Reservoir are made each day using gaged outflow (station ONYF, table 6; fig. 11) and nearby measurements of pan evaporation. Evaporation from the reservoir in excess of precipitation for water years 1945–84 was estimated to be about 300 acre-ft/yr (station TLRC, table 6). Mean annual pan evaporation for the same period was 92.6 in. (station TLYR, table 6). Measurements were not made that permit a calculation of ground-water recharge from the reservoir. This recharge is caused by the elevated stage of the reservoir in comparison with nearby ground-water levels. Some of the recharge, particularly seepage through the face of the earthen dam, may be gained back into the Owens River just downstream (south) of the reservoir, as in the case of the Pleasant Valley Reservoir. Because of the large values of river inflow and outflow (about 450 ft^3/s), any value of ground-water recharge calculated as a residual in a mass-balance equation has a high degree of uncertainty.

To gain a better understanding of the interaction of reservoirs with the ground-water system, detailed maps of surface-water and ground-water contours near each reservoir were developed. Water-level data for 1984 were plotted at a scale of 1:62,500 using a 10-foot

contour interval. In the area near the Pleasant Valley Reservoir, few ground-water-level data points were available and, therefore, the contouring was inconclusive. The elevated stage of the reservoir, however, indicates that it was recharging the nearby ground-water system. In the area surrounding the Tinemaha Reservoir, the water-level data clearly indicate a hydraulic gradient from the Owens River, and possibly from the northern part of the Tinemaha Reservoir, to the northwest toward production wells along the edge of Crater Mountain (fig. 1). This gradient indicates that, as suggested by T.E. Griepentrog (Buckhorn Geotech, written commun., 1985), surface water from the reservoir was moving into and through the ground-water system in a northwest direction. This direction of movement is just opposite of the natural flow direction prior to increased pumpage in the Big Pine area. Although qualitatively helpful, the contouring methods did not yield reliable estimates of the quantity of recharge.

Water quality of outflow from the Tinemaha Reservoir was sampled bimonthly during 1974–85 as part of the USGS National Stream Quality Accounting Network. The principal ions found in the samples were calcium (the predominant cation), sodium, bicarbonate (the predominant anion), and sulfate. Total concentration of dissolved solids ranged from 66 to 274 mg/L, with a mean of 181 mg/L (Hollett and others, 1991, table 4). This particular sampling point indicates the quality of water emanating from the reservoir and may reflect some changes in chemical and physical properties because of residence time in the reservoir. Comparison of these data with data from nearby ground water may aid in understanding the dynamics of flow between the reservoir and the ground-water system. However, it is likely that additional surface-water and ground-water samples would be needed for the comparison. A similar analysis of water quality in and around the Pleasant Valley Reservoir would help answer similar questions of seepage rates and flow directions in that area.

Small Lakes

Several small lakes, including Klondike, Warren, and Diaz Lakes (figs. 3 and 11), are present in the Owens Valley. Diaz Lake and, more recently, Klondike Lake have been used for recreation, including fishing and the use of motor boats. To accommodate this usage, water levels in Klondike and Diaz Lakes have been maintained within a fairly narrow range by the diversion of water from nearby tributary streams and canals.

Prior to being used and managed for recreation in 1986, Klondike Lake functioned much as does Warren Lake. Under unmanaged conditions, water levels in both lakes fluctuate markedly from one season to another and from one year to another depending on the quantity of runoff and the altitude of nearby ground-water levels. During above-average runoff years (fig. 12 and table 7), the lakes fill; during drier periods, the lakes empty as a result of local withdrawals and evapotranspiration.

Because the lakes are topographically low points, they most likely are natural ground-water discharge areas under unmanaged conditions. During wet periods, the lakes receive an influx of water and probably act as localized recharge points to the ground-water system. In general, this type of recharge will be temporary—as the water level in the lake falls, the hydraulic gradient from the ground-water system to the lake is reestablished, and the ground-water system resumes draining. This cyclical process is similar to that observed for the lower Owens River.

Detailed analysis of the small lakes and the surrounding ground-water system is beyond the scope of the present study. However, as an aid in determining local recharge and discharge relations, water-level data were plotted at a scale of 1:62,500 using a 10-foot contour interval as was done in analyzing the reservoirs. No indications of recharge from or discharge to the lakes were evident. The absence of a noticeable hydraulic gradient suggests that the rates of exchange with the ground-water system probably are small and localized in comparison with the more dominant controls on ground-water flow, such as recharge from tributary streams and discharge to the Owens River.

Although the small lakes do not seem to have a major effect on the valleywide hydrologic system, they can be locally important. For example, Klondike Lake is north of production wells near Big Pine and may buffer the effects of pumping, much as the Tinemaha Reservoir does to the south. As pumpage increases and ground-water levels decline, additional recharge will be induced from Klondike Lake, thereby minimizing ground-water-level declines and increasing recharge to the ground-water system. The presence of fine-grained, lake-bottom sediment will inhibit, but not prevent, recharge. Similarly, Diaz Lake may provide an important source of ground-water recharge for the Lone Pine area, including the Lone Pine town-supply wells.

Canals, Ditches, and Ponds

Canals and Ditches

A complex network of canals and ditches, particularly near Bishop, have been used to convey water for irrigation, livestock, and ground-water recharge (figs. 3 and 11). The canals and ditches range in length from tens of feet to tens of miles and, although some channels are lined with broken rock or concrete, most have sides and bottom composed of native earth. The original purpose of many of the ditches in the Bishop area was to drain the soil so that the land could be farmed. Agricultural activities, begun in the late 1800's, increased rapidly and by 1920 there were about 24,000 acres of cultivated crop land and 51,000 acres of flood-irrigated pasture land (D.E. Babb, Los Angeles Department of Water and Power, written commun., 1988).

By 1978, irrigated farmlands had declined to about 17,000 acres, largely as a result of land purchases by the Los Angeles Department of Water and Power and subsequent retirement of land from irrigated use. Over the past 75 years in the Owens Valley, the net result of many separate changes in land use has been a general shift toward less local consumption of water (table 4; Hollett and others, 1991, fig. 5).

Changes in land use, beginning about 1968, affected the operation of canals and ditches. Although less land was being farmed, the allocation of water to the remaining farms and ranches was more certain. The few canals and ditches that remained in operation had a more constant flow rate during each year, and from year to year (R.H. Rawson, Los Angeles Department of Water and Power, oral commun., 1988). With more uniform conditions, recharge from the canals and ditches to the ground-water system probably also was more uniform.

As of 1988, most of the canals and ditches in the Owens Valley are used conjunctively for purposes of flood control, irrigation, stockwater, recreation, wildlife habitats, and spreading of water for recharge. The Bishop area has the highest density of canals and ditches, and most of the larger ones are operated during most of the year (fig. 11). South of Bishop, canals and ditches are concentrated in agricultural areas near the towns of Big Pine and Lone Pine, and in the vicinity of Oak Creek near Independence (fig. 3).

Parts of the Owens Valley that no longer have active farms or ranches, such as east of Independence, still have remnant canals and ditches. Some of the canals and ditches are marked by occasional trees. The ditches typically are the lowest point of the local land surface and determine the highest altitude of ground-water levels. Ground water rising to a higher altitude is drained. In extremely-high-runoff years, such as 1969 and 1983 (table 7), dormant canals and ditches in the areas south of Bishop and east of Independence are used by the Los Angeles Department of Water and Power to disperse excess surface water.

The complex and confusing array of canals and ditches in the Bishop area (fig. 3) makes detailed analysis difficult. Computations of surface-water and ground-water budgets are probably less reliable than those made for other parts of the valley. To help overcome this complexity, the Los Angeles Department of Water and Power maintains more than 500 continuously recording gaging stations on the canal and ditch system. The stations generally are equipped with a Parshall flume and recording float (R.H. Rawson, Los Angeles Department of Water and Power, oral commun., 1987). Most of the stations are used to document the quantity of water delivered to individuals who lease lands from the Los Angeles Department of Water and Power.

The specific interaction of each canal and ditch with the ground-water system is not documented, but estimates can be made by comparing measurements of discharge at the different gages and subtracting estimates of water use between the gages. Using this approach, the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1988) concluded that most of the canals lose water to the ground-water system. This interaction is just the opposite from that observed when the valley was first developed for farming in the late 1800's, when many of the canals were built to drain the soil. Some localized sections of canals, particularly in the Bishop area, may still operate as drainage ditches.

The quantity of ground-water recharge from canals and ditches varies from one year to the next depending on operating conditions. Data for the larger canals and ditches, such as the North (upper) McNally and the Big Pine Canals (fig. 11), indicate that loss rates of as much as 1.1 $(ft^3/s)/mi$ can be sustained over a period of several months. These larger conveyances typically have water flowing in them continuously except for brief periods of maintenance. Most of the water flowing in them and the related recharge is from diversions of tributary streams and the Owens River.

However, during some periods, ground-water pumpage is the only source of water routed into some sections of the canals. Recharge under these conditions is a localized recycling of ground water. This condition is most common for the South (lower) McNally Canal, which has a series of wells spaced along its banks (fig. 11).

Riparian vegetation growing in and along the canals and ditches withdraws water from the soil-moisture zone and reduces the quantity of seepage that actually enters the ground-water system. This reduction in actual recharge was found to be minimal [less than 0.02 (ft^3/s)/mi] using calculations based on estimates of the width of vegetation (5 to 20 ft), percentage of vegetation cover (30 to 100 percent), and evapotranspiration (40 to 60 in/yr).

An estimate of recharge was made for each of the 19 larger canals and ditches, which have individual names such as the Owens River Canal. The largest of these are shown in figure 11; all 19 canals and ditches are shown on USGS 1:24,000-scale topographic maps compiled by the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1987). Recharge was calculated using measured and estimated loss rates, the measured length of the channel, and the average period of operation. Typically, the canals and ditches lost about 0.7 (ft^3/s)/mi and were operated all year. Total recharge from the named canals and ditches within the defined aquifer system (fig. 4) was estimated to average about 20,000 acre-ft/yr.

Many smaller, unnamed canals and ditches have a lower loss rate because of a smaller wetted perimeter and lesser depth of water. The recharge from these conveyances was lumped into the values of ground-water recharge from irrigation and watering of livestock discussed in later sections of this report.

The effect on native vegetation from operation of the canals and ditches is not well documented. In general, however, when a canal or ditch is taken out of service, as was the Owens River Canal (fig. 11) after 1969, recharge to the ground-water system is reduced and the quantity of water available for evapotranspiration in the immediate vicinity of the canal is less. This change may be visible as a reduction in the quantity of leaves or possibly the number of plants (Groeneveld and others, 1986b) in the immediate vicinity of the canal or ditch. If the canal or ditch is elevated above the water table, then similar effects can be expected to occur toward the center of the valley where the water table is closer to the rooting depth of native vegetation.

Ponds

Several ponds are operated in the valley, usually in conjunction with canals and ditches, for wildlife habitat and as areas to contain operational releases of surface water or to purposefully recharge the ground-water system. Some of the pond-like areas are referred to as sloughs, although the distinction generally is not important. Sloughs, which are referred to as ponds in this report, tend to be areas with a more undulating topography and a less-well-defined shoreline. The primary areas of ponds are Farmer's Ponds north of Bishop; Buckley Ponds, Arkansas Flats, Runkle Slough, and Partridge Slough south of Bishop; Thibaut Ponds near Thibaut Creek; Calvert Slough near Taboose Creek; and Billy Lake east of Independence. The location of these areas is shown on USGS 1:24,000-scale topographic maps and on land-use maps compiled by the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1987). The quantity of discharge to these areas varies with the quantity of runoff in the valley (table 7). In years with below-normal runoff, little or no water is diverted except to the few migratory-bird habitat areas, such as Farmer's Ponds. In years with unusually high quantities of runoff, the ponds are flooded with tens of thousands of acre-feet of water.

After operation of the second aqueduct was begun in 1970, purposeful recharge operations were emphasized in order to help balance the increased quantity of ground water pumped. Whenever extra surface water is available, in excess of the demands for wildlife habitat, it is diverted to areas with the most favorable ground-water-recharge characteristics. During high-runoff years, such as 1978, just the purposeful ground-water recharge from those areas has been estimated to be as much as 25,000 acre-ft (R.H. Rawson, Los Angeles Department of Water and Power, written commun., 1988). During average and below-average runoff years (fig. 12 and table 7), the total quantity of recharge from ponds is much less.

Annual recharge from each pond was estimated from an annual water-use summary obtained from the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1988). In this unpublished summary, water use is tabulated by area of the basin (Laws, Bishop, Big Pine, Tinemaha–Haiwee) and by category of water use (operational, ground-water recharge, recreation and wildlife, enhancement and mitigation). In general, operational use is defined as water that is released from the river–aqueduct system

for safety or maintenance reasons; ground-water recharge is defined as water used to purposefully maximize recharge of the aquifer system; recreation and wildlife is defined as surface water released to meet the needs of wildlife, primarily birds; enhancement and mitigation is defined as water designed to meet the needs of vegetation in selected areas.

With the considerable aid of R.H. Rawson, percentages were chosen to split the summary values for each area into values for individual ponds (or pond-like areas). For example, water used in the Laws area for operational purposes is distributed to three ponds: south of the North (upper) McNally Canal, south of the South (lower) McNally Canal, and near the Laws Ditch (fig. 11). The average percentage distribution to each pond was estimated to be 40 percent, 40 percent, and 20 percent, respectively.

Also with the aid of R.H. Rawson, a recharge rate was estimated for each pond and use of water. For example, recharge from an operational release of water to the pond near the Laws Ditch was estimated to be about 20 percent of the total water released. In contrast, recharge from water designated as ground-water recharge in the same pond was estimated to be about 75 percent. This large difference in recharge rates for the same physical area results from the specific conditions, timing, and volume of the release of water. The extensive gaging-station records maintained by the Los Angeles Department of Water and Power aided in confirming the reasonableness of the estimates for water distribution and recharge. From these estimates, annual recharge was calculated for 28 different combinations of ponds and water use for water years 1970–88.

Tabulated summaries for years prior to 1970 were not available from the Los Angeles Department of Water and Power. Therefore, correlations between the 1970–88 data and the percent valleywide runoff were used to determine values of water distribution and recharge for water years 1963–69. Because changes in definitions and categories occurred during the period 1970–88, such as between “operational releases” and “ground-water recharge,” some judgement was required in assigning the earlier values. Average recharge from all ponds within the defined aquifer system (fig. 4) was estimated to be 12,000 acre-ft/yr during water years 1963–69 and 11,000 acre-ft/yr during water years 1970–84.

Owens Lake

The Owens Lake is the terminus for the natural surface-water system (figs. 1, 3, 10*D*, and 11). Runoff that is not diverted into the Los Angeles Aqueduct, recharged to the ground-water system, or evapotranspired eventually flows onto the Owens Lake playa and is evaporated.

Historically, the Owens Lake was as much as 20 ft deep, and steam-powered ferry boats crossed it. As of 1988, the lake was dry, except for a small area near the northwestern side. Spring discharge into the lake is visible along the northwestern shore—presumably ground-water discharge from the area west of the Alabama Hills. During the high-runoff year of 1983 (fig. 12), the lake occupied nearly the entire area of the playa shown in figures 1 and 10*D*, but it evaporated almost entirely within a single year. Not surprisingly, lake water and nearby ground water have exceptionally high concentrations of dissolved solids (Hollett and others, 1991; Lopes, 1988).

Although not a part of the detailed study area for this investigation, the Owens Lake remains a major factor in water-management operations within the Owens Valley. The restriction on the Los Angeles Department of Water and Power from discharging water into the lake and the occurrence of huge dust storms, which are believed to be related to rewetting of the playa and which occasionally extend from the area of the Owens Lake to north of Independence, are ongoing topics of investigation (Saint-Amand and others, 1986; Lopes, 1988).

Ground-Water System

The ground-water system of the Owens Valley is unusual in comparison with that of other basin-and-range valleys in eastern California. The abundant precipitation in the Sierra Nevada and resulting runoff fills the basin to nearly overflowing each year. Historically, this abundance of water has eroded the surrounding mountains, filled the graben with highly transmissive deposits, and created a shallow water table beneath much of the valley, a water table which in turn supports a great density of native vegetation not found in other similarly formed basins. In nearby basin-and-range valleys, such as Indian Wells Valley to the south (Dutcher and Moyle, 1973) and Death Valley to the southeast (Hunt and others, 1966), the quantity of runoff is much less and most of the sparse native vegetation must subsist solely on precipitation.

As a result of the abundant runoff into the Owens Valley, the surface-water and ground-water systems are strongly linked. Much of the valley floor is characterized by surface-water conveyances that are in contact with the ground-water system (figs. 3 and 10), and this connection facilitates a ready exchange of water. Native vegetation on the valley floor is dependent on a combination of water obtained from precipitation, subirrigation from surface-water conveyances, and ground water. Since 1970, when export of water from the valley was expanded to include ground water, the two systems have become linked even more closely politically as well as physically. Water management of one system typically has a noticeable effect on the other.

The following sections describe the hydrogeologic framework of the ground-water system; the hydraulic characteristics of the hydrogeologic units that compose the system; the source, occurrence, and movement of water through the system; and the valley-wide ground-water flow model used to simulate the system and evaluate selected water-management alternatives. The hydrogeologic history of the ground-water system and related aquifer materials is described in detail by Hollett and others (1991). Many of the major components of the ground-water system are strongly linked to a surface-water feature, such as the river–aqueduct system. For these components, the primary description, including quantification of ground-water recharge and discharge, is presented in an earlier section entitled "Surface-Water System."

Geometry and Boundary Conditions

Nearly all the recoverable ground water in the valley is in the unconsolidated to moderately consolidated sedimentary deposits and intercalated volcanic flows and pyroclastic rocks that fill the basin. Where saturated, these sedimentary deposits and volcanic rocks make up the ground-water system. The primary part of the ground-water system, defined by Hollett and others (1991) as the "aquifer system," is capable of yielding significant quantities of ground water to wells (Lohman and others, 1972). The defined aquifer system delineated in figure 14 is also the part of the ground-water system that was simulated with the valleywide ground-water flow model documented later in this report.

The aquifer system is a three-dimensional body of valley fill that is saturated with ground water. This saturated volume of valley fill is bounded on all sides by a "boundary surface" (Franke and others, 1987). The boundary surface allows water to either flow in or out of the system, such as at the water table, or acts as a flow barrier, which allows little or no water to enter or leave the system across the boundary surface, such as at a bedrock contact.

The upper boundary surface of the aquifer system is the water table and the lower surface is either a bedrock contact, the top of moderately consolidated valley fill, or an arbitrary depth based on the depth of pumped wells. The sides of the aquifer system are either bedrock or a part of a lateral boundary surface that allows ground water to flow in or out of the aquifer system, termed a "flow boundary." Thus, water can flow in (recharge) or out (discharge) of the aquifer system only through a flow boundary.

Flow also occurs into or out of the Owens Valley aquifer system at wells, springs, rivers, or as underflow through a cross section of the aquifer system. Lateral inflow boundaries (underflow) include sections along the southeast end of Round Valley, south end of Chalfant Valley, and that part of the two valleys overlain by the Volcanic Tableland (figs. 4, 5, and 14). Underflow also enters the aquifer system from the drainages of Bishop and Big Pine Creeks and from Waucoba Canyon. The lateral outflow boundary from the system is a section that crosses the valley approximately east to west at the south end of the Alabama Hills.

Hydrogeologic Units and Subunits

The hydrogeologic framework of the aquifer system controls the vertical and horizontal flow of ground water in the system. The complex framework of the actual system was simplified by Hollett and others (1991) into a vertical series of units that represent either ground-water-producing zones or major zones of confinement to vertical flow. These units are referred to as "hydrogeologic units" and are numbered 1 to 3, from top to bottom in the aquifer system. Saturated valley fill that lies below the defined aquifer system and in contact with the bedrock is referred to as hydrogeologic unit 4 and is not part of the aquifer system. The primary purpose for simplifying the heterogeneous sedimentary and volcanic materials into hydrogeologic units was to be able to discretize the aquifer system for the three-dimensional, ground-water flow model. Shown in figure 5 are typical hydrogeologic sections representing the major structural and depositional areas of the aquifer system and the division into hydrogeologic units. Additional sections and descriptions are

presented by Pakiser and others (1964), Bateman (1965), Griepentrog and Groeneveld (1981), and Hollett and others (1991).

The criteria for dividing the aquifer system into hydrogeologic units are described in detail by Hollett and others (1991); only a summary is presented here. The first criterion used to divide the aquifer system is a method that defines the hydrogeologic units on the basis of uniform hydraulic properties, commonly represented by geologic or stratigraphic units. This method worked well for some parts of the aquifer system, such as the thick clay beds near Big Pine (section *B–B'*, fig. 5), but not for most of it. The second criterion defines hydrogeologic units on the basis of the distribution of vertical head. This method enabled the definition of units in the thick sequences of valley fill where interfingering and lateral discontinuity cause complex heterogeneity, such as beneath much of the valley floor. The third criterion defines hydrogeologic units on the basis of the depth at which significant recharge or discharge can occur. In areas of the Owens Lake Basin where little information is present to differentiate between hydrogeologic units 3 and 4 (section *C–C'*, fig. 5), the base of hydrogeologic unit 3 was chosen arbitrarily at 1.5 times the depth of the deepest production well in the area. The following is a brief description of the geologic, stratigraphic, and hydraulic characteristics of each of the hydrogeologic units.

Hydrogeologic Unit 1.—Hydrogeologic unit 1 represents the unconfined part of the aquifer system and includes the water table as the upper boundary surface. Unconfined conditions are areally pervasive throughout the aquifer system, although the depth of significant confinement varies with local conditions. Typically, the upper 100 ft of saturated deposits displays minimal restriction to the vertical movement of water, and differences in hydraulic head usually are less than 2 to 3 ft. In some parts of the aquifer system, confined conditions near the water table can be created by the less transmissive layers of the olivine basalt flows or by a fine-grained fluvial or lacustrine deposit (figs. 4 and 14). This type of local confinement near the land surface is not typical of most conditions in the valley, and hydrogeologic unit 1 can be considered generally to have a saturated thickness of about 100 ft.

Hydrogeologic Unit 2.—Hydrogeologic unit 2 is the material, where present, that separates hydrogeologic unit 1 from hydrogeologic unit 3. In the middle of the valley, this material typically consists of fine-grained silt and clay beds that restrict the vertical movement of ground water. Near Big Pine, hydrogeologic unit 2 is composed of a massive, readily identifiable clay bed with a total thickness of more than 80 ft—referred to as the “blue-green clay” by Hollett and others (1991, p. 31 and fig. 12). Vertical groundwater flow also is restricted by the volcanic materials of the Big Pine volcanic field even though they are depositionally much different from the fine-grained silt and clay beds. The volcanic material in the aquifer system near Bishop, in contrast, consists mostly of unconsolidated pumice (the lower member of the Bishop Tuff), which has hydraulic properties similar to sand and offers minimal restriction to vertical flow. Along the margins of the valley, the alluvial fan deposits are relatively homogeneous, displaying no dominant horizontal layering. In these areas, hydrogeologic unit 2 is virtually absent.

Hydrogeologic Unit 3.—Several confined zones that are present in the aquifer system have been combined into hydrogeologic unit 3. The confined part of the aquifer system generally extends from the toes of the alluvial fans along the Sierra Nevada to the toes of the alluvial fans along the Inyo and the White Mountains and extends along nearly the full length of the valley (fig. 14). Confinement is created by a number of lenticular-to-continuous, flat-lying fluvial and lacustrine clay and silty-clay beds (hydrogeologic unit 2). Confinement also can be created by fine-grained material deposited by mudflows. These confining beds thin to extinction along the margins of the valley. Additional areas of confinement may be formed by the upper member of the Bishop Tuff, where present (fig. 5), and by volcanic flows of the Big Pine volcanic field (fig. 4), but an absence of data in these areas prevents a more detailed analysis. Saturated thickness of hydrogeologic unit 3 ranges from tens of feet along the margins of the basin to about 500 ft beneath most of the valley floor.

Hydrogeologic Unit 4.—Although not part of the defined aquifer system, hydrogeologic unit 4 occupies a large part of the valley fill (fig. 5). Despite its large volume, the quantity of ground water flowing through or extractable from hydrogeologic unit 4 probably is minimal. Deep test drilling during 1988 by the Los Angeles Department of Water and Power (E.L. Coufal, oral commun., 1988) showed that most materials at depths greater than about 700 ft do not yield significant quantities of water to wells, generally less than 0.2 ft^3/s. Deep volcanic deposits penetrated by drilling near Taboose Creek (fig. 14) may yield greater quantities, although no aquifer testing was done. Except at the location of these deep test borings and a

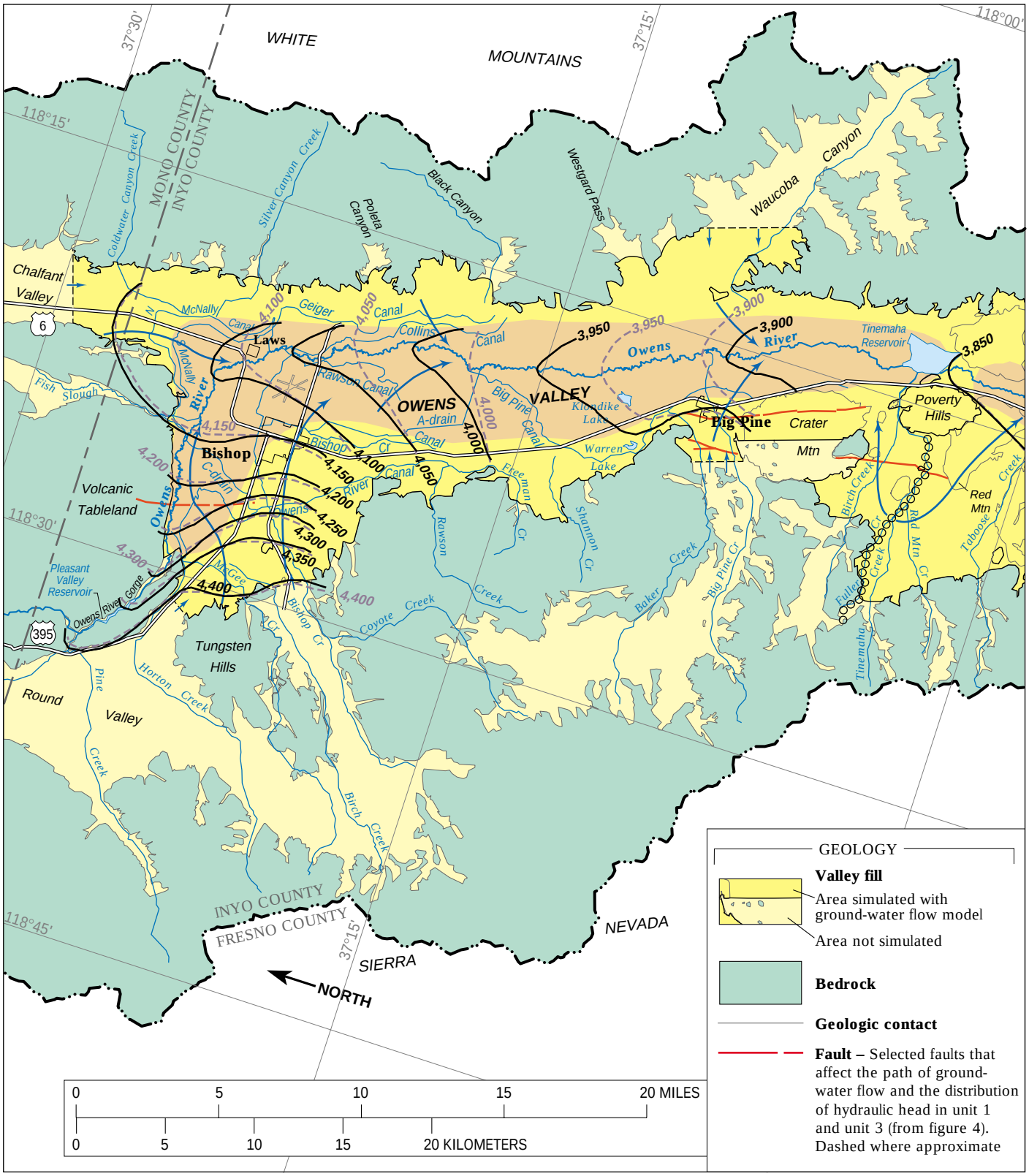
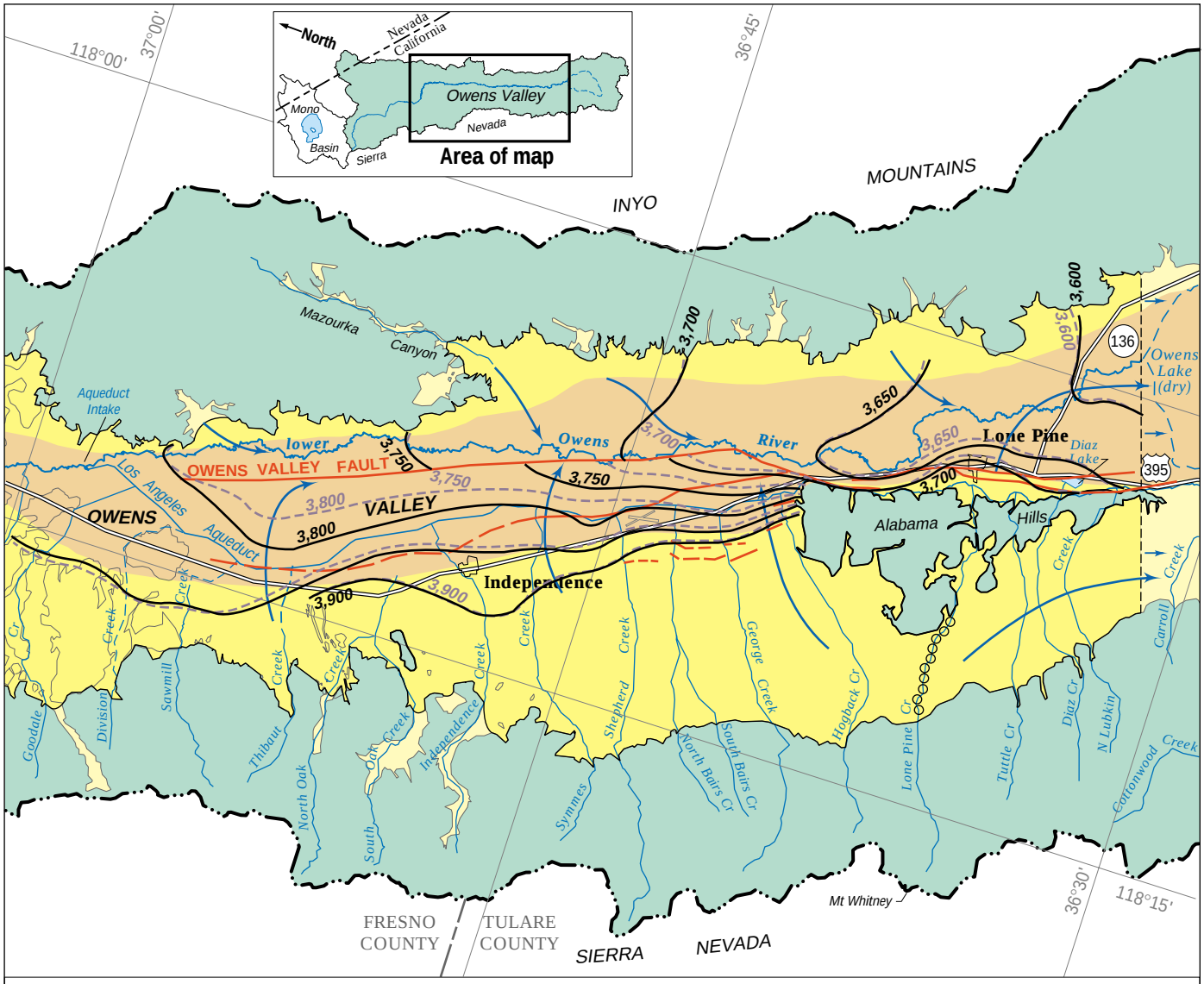


Figure 14. Ground-water conditions in the defined aquifer system of the Owens Valley, California, spring 1984. Shown area areal extent of the defined aquifer system, occurrence of unconfined and confined conditions, boundary conditions, configuration of potentiometric surface in hydrogeologic units 1 and 3, and generalized direction of ground-water flow (from Hollett and others, 1991, fig. 17).



EXPLANATION

HYDROLOGY

Approximate extent of the confined part of the aquifer

3,900 — **Potentiometric contour –** Shows approximate altitude of the water table in hydrogeologic unit 1, represented by the upper layer of the ground-water flow model, spring 1984. Contour interval 50 feet. Datum is sea level

3,900 — **Potentiometric contour –** Shows approximate altitude of the hydraulic head in hydrogeologic unit 3, represented by the lower layer of the ground-water flow model, spring 1984. Contour interval 50 feet. Datum is sea level

Generalized direction of ground-water flow – Combined direction of ground-water flow in hydrogeologic units 1 and 3

Ground-water divide – Approximately located

Boundary of the aquifer system – As defined in this report. Arrows indicate the direction of ground-water flow to or from adjacent permeable materials

Boundary of the Owens Valley drainage basin

Figure 14. Continued.

few previously drilled deep wells, the chemical and hydraulic characters of hydrogeologic unit 4 are largely undocumented.

Hollett and others (1991) further divided the hydrogeologic units into subunits on the basis of the type of geologic deposit (fig. 4). For example, hydrogeologic unit 1 in section *C–C'* (fig. 5) has subunits 1a representing alluvial fan deposits and 1c representing undifferentiated fluvial deposits. Hydrogeologic unit 3 in the same section has subunit 3a representing alluvial fan deposits; subunit 3t representing transition-zone deposits; and subunit 3c representing undifferentiated fluvial deposits. Additional subunits were defined for volcanic deposits and massive clay-bed deposits (figs. 4 and 5). The combination of hydrogeologic units and subunits formed the basis of ground-water "model zones" discussed later.

Hydraulic Characteristics

The hydraulic characteristics of the aquifer system—transmissivity, saturated thickness, horizontal and vertical hydraulic conductivities, specific yield, and storage coefficient—were estimated from pumped-well and aquifer tests, drill-hole data, and geophysical data. Detailed descriptions of the methods used to define the hydraulic characteristics and a general range of horizontal hydraulic conductivity and specific yield for different types of aquifer materials in the Owens Valley are presented by Hollett and others (1991, table 1). Additional confirmation of these values was obtained from preliminary ground-water flow models (Yen, 1985; Danskin, 1988; Hutchison, 1988; Hutchison and Radell, 1988a, b; Los Angeles Department of Water and Power, 1988) and from development and calibration of the final valleywide ground-water flow model documented in this report.

The areal distribution of aquifer characteristics was determined by analyses of all known pumped-well and aquifer tests, at more than 130 wells, in the valley. A complete list of the transmissivity, average horizontal hydraulic conductivity, and storage coefficient obtained from these analyses and the method of calculation (aquifer-test method) are given in table 9 (p. 155). In some cases, several calculations were made for a single well. Values calculated by the Los Angeles Department of Water and Power (M.L. Blevins, written commun., 1984–87) for some wells also were obtained. The values given in table 9 are those most representative of transmissivity unaffected by leakance and of a longer-term storage coefficient that reflects drainage of the aquifer system. These criteria were chosen in part to ensure consistency with the valleywide ground-water flow model. Leakance, if not taken into account in aquifer-test analysis, will tend to increase calculated transmissivity values. Storage coefficient, which is specific yield for water-table conditions, was difficult to calculate from the available tests. None of the values reach the 0.10–0.15 range that is characteristic of a true specific yield of these aquifer materials (Hollett and others, 1991; S.N. Davis, 1969). Much longer aquifer tests probably are required to achieve more representative values of specific yield. Calculation of storage coefficients for confined conditions was somewhat more successful; values typically ranged from 0.0005 to 0.005. Average horizontal hydraulic conductivity was calculated using an estimate of the total saturated thickness of transmissive deposits affected by the well—calculated as the depth of the well below the water table minus the total thickness of clay layers or, if data were available, as the total length of perforations.

The areal distributions of transmissivity and average horizontal hydraulic conductivity are shown in figures 15 and 16, respectively. Both sets of values are well correlated with the distribution of depositional materials (figs. 4 and 5). Values for many of the wells near the Los Angeles Aqueduct in the Owens Lake Basin reflect the buried, more transmissive, transition zone deposits (fig. 5) rather than the overlying, less transmissive, alluvial fan deposits.

In some cases, the transmissivity values in figure 15 and table 9 represent only a part of the transmissivity of the aquifer system. Some wells are not open to all of the transmissive aquifer materials, especially shallow materials, or the wells may not penetrate the entire depth of the aquifer system, especially in the volcanic areas. For these reasons, extrapolation of transmissivity values to the entire aquifer needs to be done cautiously. Alternatively, average horizontal hydraulic conductivity values (fig. 16) multiplied by an estimate of the saturated thickness of the aquifer system may yield more reliable values of transmissivity. Gross estimates of saturated thickness in the center of the valley are 100 ft and 500 ft for hydrogeologic units 1 and 3, respectively. The thickness of hydrogeologic unit 2 is minimal, generally less than 15 ft, except near Big Pine.

Movement of Ground Water

Virtually all the ground water in the Owens Valley aquifer system is derived from precipitation that falls within the Owens Valley drainage basin area (fig. 1). Ground-water recharge (deep infiltration) occurs primarily through the alluvial fans as water runs off the Sierra Nevada as a result of snowmelt or rainfall. Most of the runoff infiltrates through the heads of the alluvial fans and through the tributary stream channels. Lesser quantities of recharge result from seepage of water flowing in canals and ditches, from direct precipitation on the sparsely vegetated volcanic rocks, from runoff from bedrock areas within the valley fill, by leakage from the river–aqueduct system, and as underflow from Chalfant and Round Valleys. Underflow to the Bishop Basin from Chalfant Valley also includes water moving south from Hammil and Benton Valleys. Most of the ground water from Chalfant, Hammil, and Benton Valleys is believed to enter the Bishop Basin near Fish Slough beneath the southeastern part of the Volcanic Tableland (Hollett and others, 1991, p. 63). Recharge to the aquifer system is minimal from percolation of water that moves through fractures in the surrounding bedrock to the zone of saturation or, because of the high evapotranspiration, from water that percolates directly to the water table from rainfall on the valley floor.

Ground water moves along permeable zones of the ground-water system from areas of higher head to areas of lower head. The direction of ground-water flow is approximately perpendicular to lines of equal head. The areal pattern of ground-water flow in the valley is shown in figure 14. The vertical flow directions in hydrogeologic units 1, 2, and 3 are shown in figure 5 and can be inferred from the relative position of equal-head contours for hydrogeologic units 1 and 3 in figure 14. The Darcian rate of flow along the illustrated flow paths is determined by the hydraulic gradient, the hydraulic conductivity, and the cross-sectional area of flow. Typical rates in the valley range from less than a foot per year in clay and silt to hundreds of feet per year in the more permeable basalt. Rates of horizontal flow of water in hydrogeologic units 1 and 3 generally range from 50 to 200 ft/yr. Additional studies of ground-water quality, particularly the analysis of hydrogen and oxygen isotopes, which can be used to determine the relative age of water, would help to confirm these rates of flow.

Ground water flows from areas of recharge to areas of discharge. Discharge can be from springs, wells, evapotranspiration, or seepage to the river–aqueduct system and the lower Owens River. In general, ground-water flow is from the margins of the valley, mainly the west margin, toward the center of the valley and then southward toward the Owens Lake (fig. 14). As ground water flows downgradient to the toes of the alluvial fans and the transition-zone deposits, the flow is primarily horizontal rather than vertical (fig. 5). This horizontal flow of ground water is split by the confining beds of hydrogeologic unit 2 that interfinger with the alluvial fan and the transition-zone deposits and direct the flow of water into hydrogeologic units 1 and 3. Discharge from hydrogeologic unit 3 is generally upward through hydrogeologic unit 2 to unit 1, from pumped or flowing wells, or through the valley fill to the south end of the valley. Discharge from hydrogeologic unit 1 is principally to evapotranspiration, pumped wells, springs, the river–aqueduct system, and the lower Owens River.

In the Bishop Basin, ground water that originates as underflow from Round and Chalfant Valleys and as underflow from the lower member of the Bishop Tuff enters hydrogeologic units 1 and 3. This water mixes with water recharged through alluvial fans and through the Big Pine volcanic rocks and moves southward along the center line of the valley (fig. 14). In the Big Pine area, however, the direction of ground-water flow has changed, at least during some periods, since 1970. Increased pumpage from wells near Crater Mountain has shifted the ground-water gradient and caused ground water to flow northwest from the Tinemaha Reservoir and west from the section of the Owens River just north of the reservoir toward Crater Mountain.

In the Owens Lake Basin, water that enters the aquifer system as underflow through the narrows or as recharge through the alluvial fans moves south to the Owens Lake (dry). Most of the water is discharged to evapotranspiration, wells, or the lower Owens River. What happens to the remaining ground water that reaches the south end of the ground-water system at the Owens Lake (dry), however, is not known with certainty. The bulk of the ground water probably flows vertically upward and is discharged as evaporation from the dry lake. Minor quantities of water may flow at depth through the fractured bedrock beneath the Haiwee Reservoir to Rose Valley, which is south of the Owens Valley. Berenbrock and Martin (1991) estimated total underflow from Rose Valley south to Indian

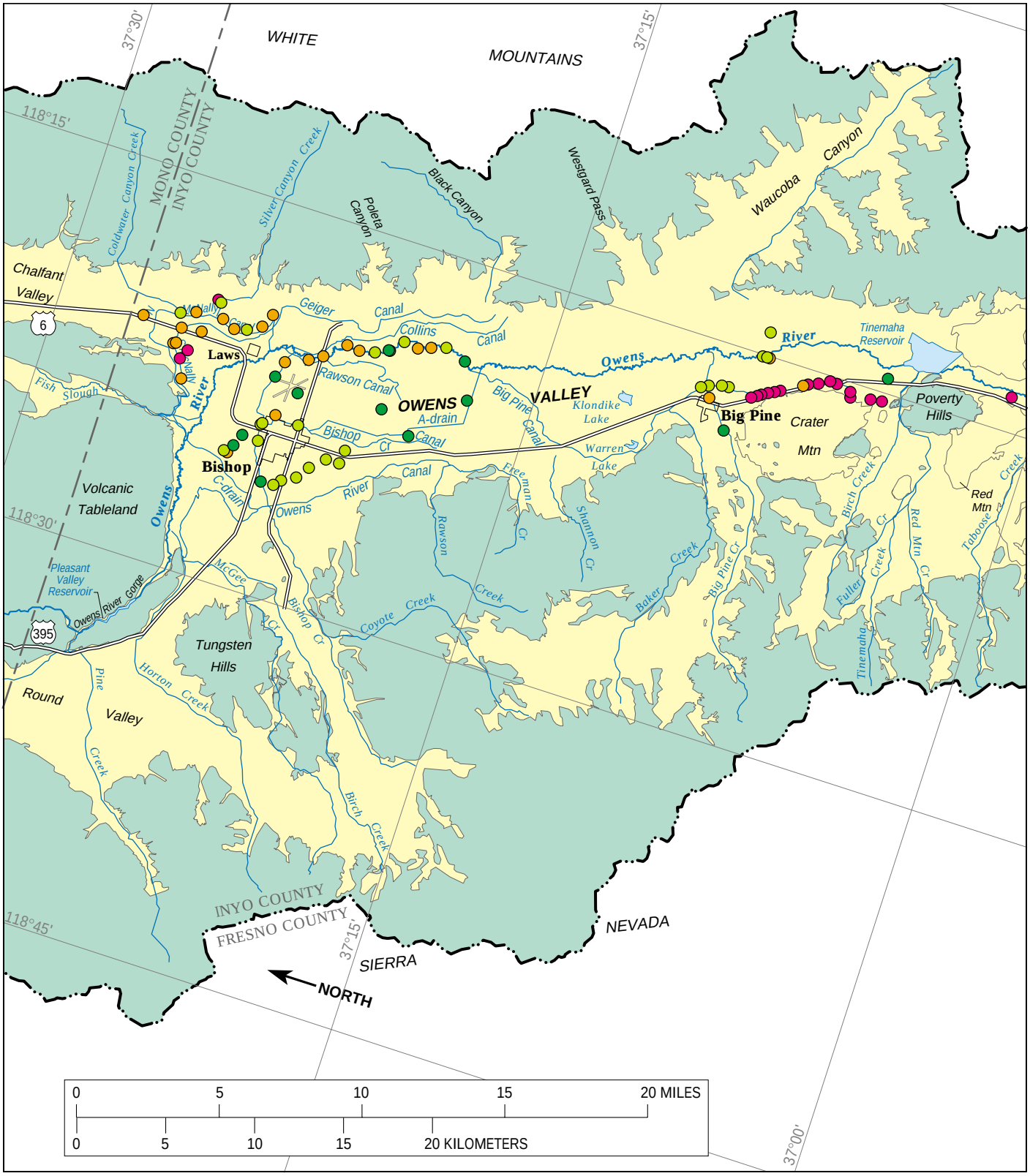


Figure 15. Transmissivity of valley-fill deposits as determined from aquifer tests in the Owens Valley, California.

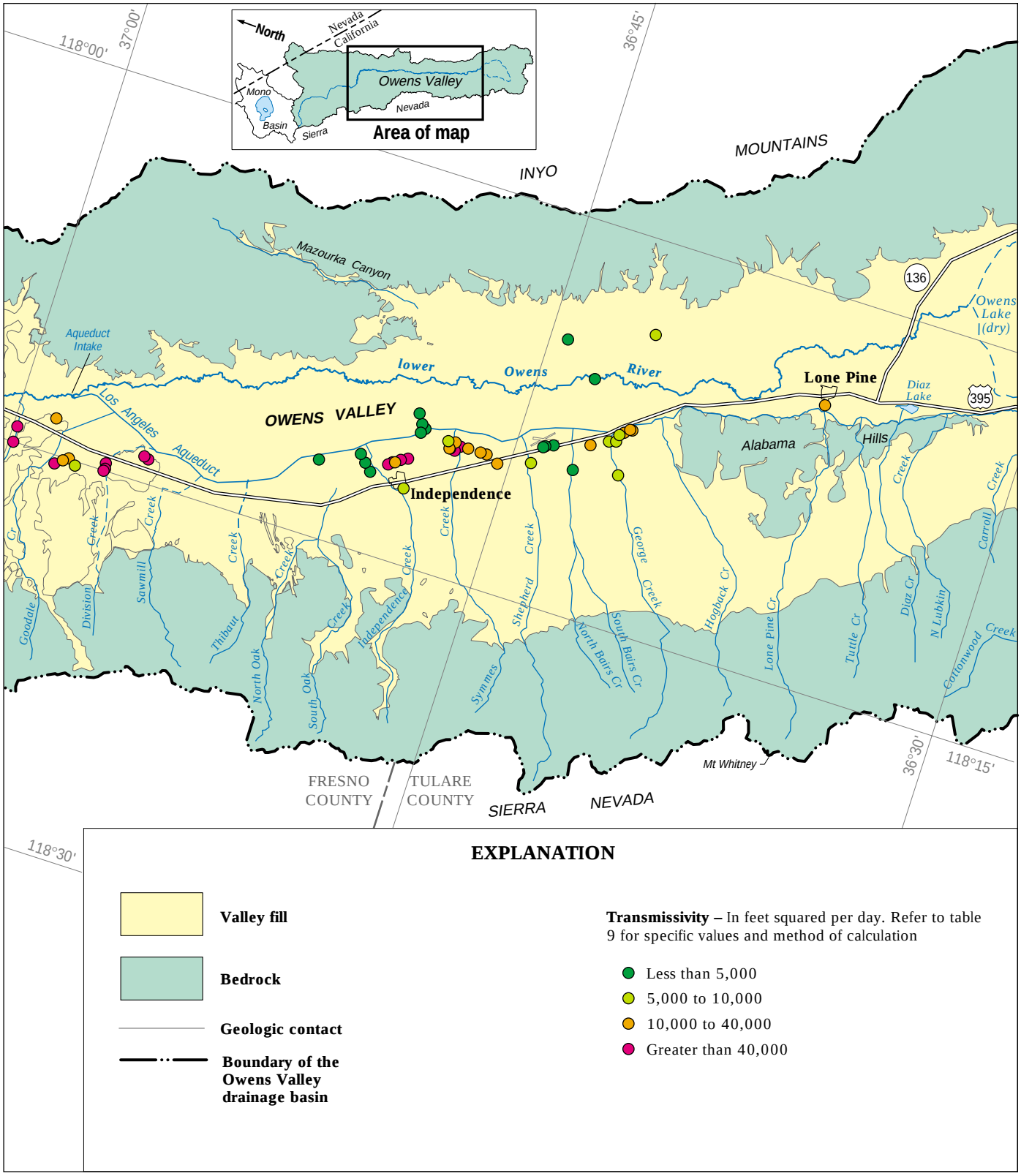


Figure 15. Continued.

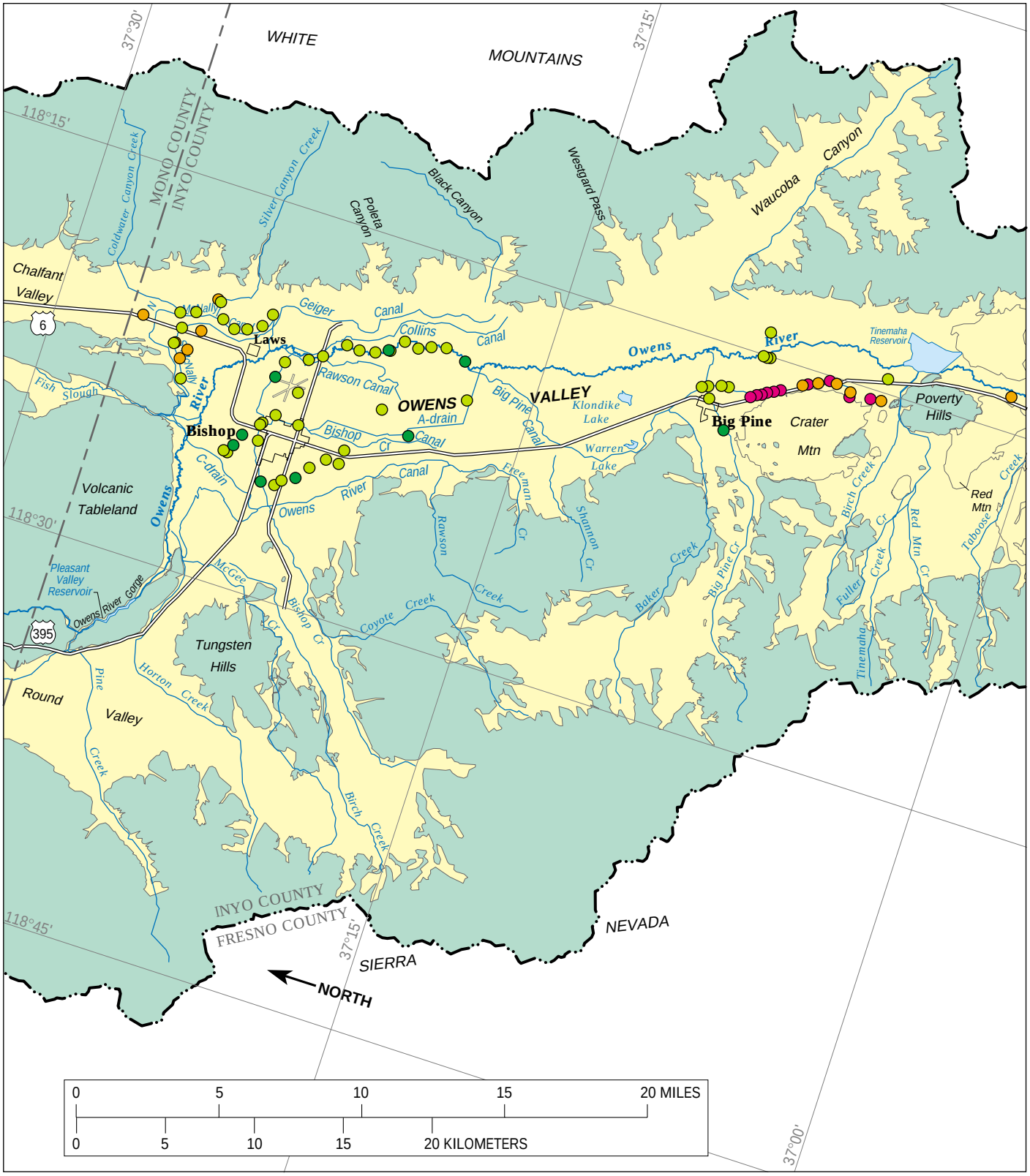


Figure 16. Average horizontal hydraulic conductivity of valley-fill deposits in the Owens Valley, California.

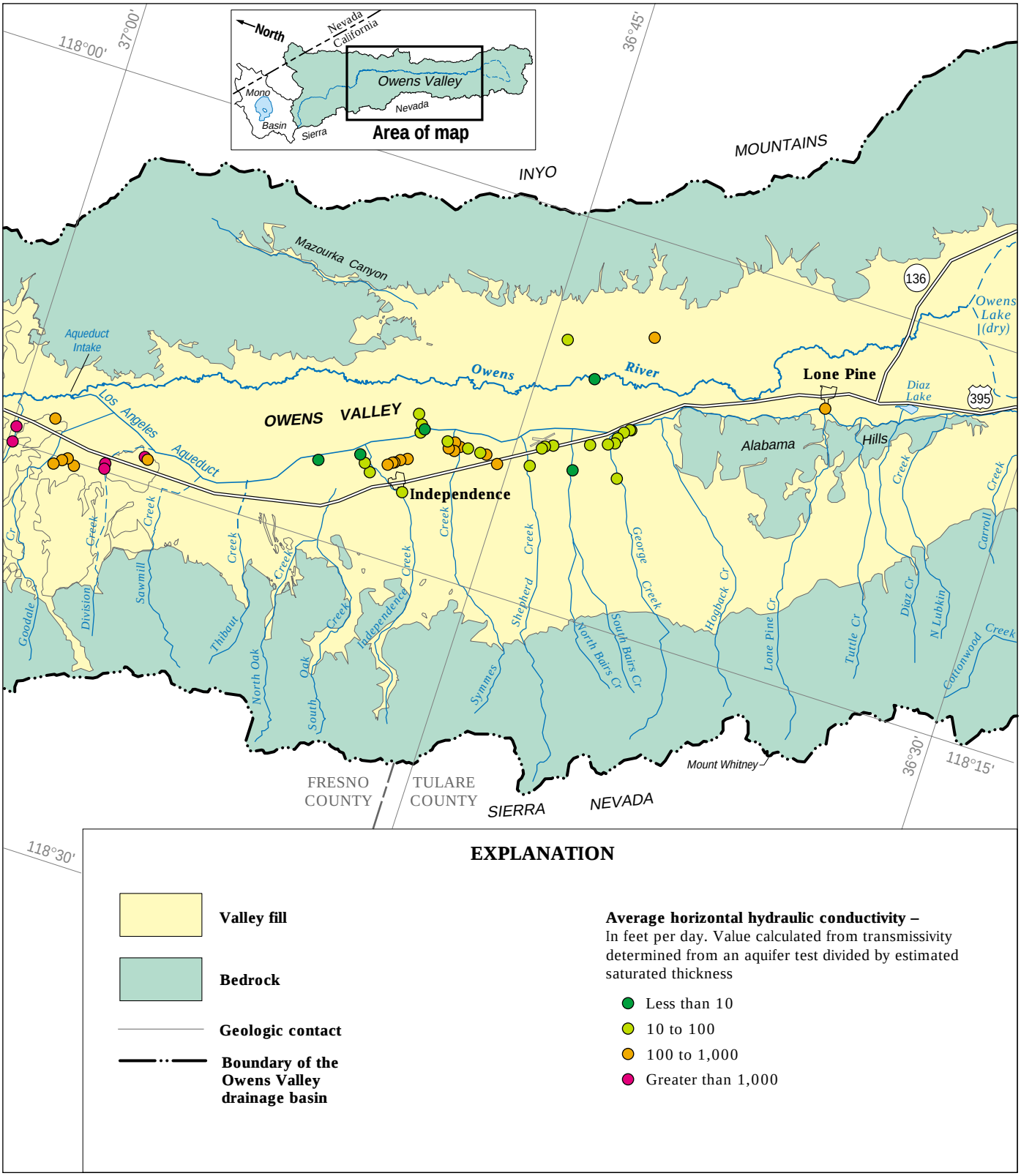


Figure 16. Continued.

Wells Valley to be less than 50 acre-ft/yr, part of which is seepage from the Haiwee Reservoir (Danskin, 1988).

The presence of faults within the aquifer system (fig. 4) may affect the movement of ground water, depending on the transmissive characteristics of the individual faults. The physical and chemical processes that cause one fault to retard ground-water movement more than another are discussed by Schaefer (1978), Freeze and Cherry (1979, p. 474) and Hollett and others (1991). Some faults in the Owens Valley, most notably the Owens Valley Fault (figs. 4 and 14), significantly retard and deflect ground-water movement. For example, the Owens Valley Fault effectively splits the Owens Lake Basin into two halves. Most ground water flows southward down the west side of the fault; lesser quantities slowly seep over and through the fault to the east side of the basin. The effects of both recharge and pumping on the west side of the basin are isolated to a large extent from the east side of the basin—except in the northern part of the Owens Lake Basin, where the Owens Valley Fault does not appear to impede ground-water movement (compare figs. 4 and 14).

Other faults that have a significant regional effect on ground-water flow were noted by Hollett and others (1991, p. 74). Additional water-retarding faults identified since that study was completed include a fault through Red Mountain (figs. 3 and 14), en echelon sliver faults near Lone Pine (figs. 4 and 14), and a probable, unexposed fault in the vicinity of west Bishop (figs. 4 and 14).

Northwest-trending faults along the east side of Crater Mountain (Hollett and others, 1991, fig.15) have created additional fractures in the highly transmissive volcanic deposits. Calibration of the ground-water flow model required much higher transmissivities in this area than for other volcanic deposits in order to maintain the unusually flat water table along the edge of Crater Mountain. These fracture conduits appear to provide an enhanced pathway for ground water recharged in the Big Pine Creek drainage to move southward through Crater Mountain to the vicinity of Fish Springs.

Some of the water-retarding faults force ground water to rise to land surface, producing noticeable seeps and springlines. Many of these features can be identified readily by an increase in vegetation (Meinzer, 1927) and are indicated by linear red zones (false color) in figure 3. An excellent example is the sequence of faults just north of the Alabama Hills (figs. 3, 4, and 14) described by D.E. Williams (1970).

In some parts of the Owens Valley, water-retarding en echelon faults have created flow compartments that are relatively isolated from the rest of the aquifer system. Areas with closely spaced faults near Lone Pine and just north of the Alabama Hills are typical of this phenomenon (fig. 4). Recharge to the compartments typically is localized, such as from a stream. Discharge may be to a spring or well. Underflow into and out of the compartment depends on the retarding effect of the fault, which may vary with depth. Simulation of these areas, as discussed later, was difficult and not particularly successful.

Hollett and others (1991, fig. 6) mapped numerous other fault traces, some of which may be locally important in affecting ground-water movement. Additional site-specific aquifer tests could be used to detect any significant retardation of ground-water flow caused by known or suspected faults in the Owens Valley. Ground-water-level data from an aquifer test show an unexpected change in the rate of drawdown if a flow-retarding fault is within the area of influence of the pumped well (Driscoll, 1986, p. 562).

The movement of ground water in the Owens Valley is controlled to a large extent by springs, seeps, evapotranspiration by native vegetation, and seepage to the river–aqueduct system and the lower Owens River. Each of these features acts as a “hydraulic buffer” on nearby ground-water levels in hydrogeologic unit 1. As the altitude of the water table increases, discharge from the springs and seeps, by native vegetation, and to the river–aqueduct system and the lower Owens River increases, thereby restricting the rise in water-table altitude. As the water table declines, discharge from each feature is reduced, thereby reducing the decline in water-table altitude. Without the broad areal distribution of these hydraulic buffers, which cover most of the valley floor, fluctuations in ground-water levels in response to changes in recharge and discharge would be much greater. The action of hydraulic buffers on ground-water levels and on recharge to and discharge from the aquifer system is a recurring theme that is exceptionally important in understanding the operation of the hydrologic system in the Owens Valley and in evaluating the effect of different water-management alternatives.

Ground-Water Budget

A ground-water budget is an accounting of the inflow to and outflow from a ground-water system (in

this case, the defined aquifer system) and the changes in the volume of ground water in storage. If inflow equals outflow and if the change in the volume of ground water is zero, then the aquifer is in equilibrium or a steady-state condition. Equilibrium is reflected by nearly constant ground-water levels or by even fluctuations of levels with no long-term rise or decline. If total inflow does not equal total outflow, then the aquifer is in nonequilibrium or a transient condition, and the change in the volume of ground water in storage is reflected in the changing ground-water levels.

In several previous investigations, water budgets have been summarized for the whole hydrologic system in the Owens Valley. The investigators include C.H. Lee (1912), Conkling (1921), California Department of Water Resources (1960), D.E. Williams (1969), Los Angeles Department of Water and Power (1972, 1974b, 1975, 1976, 1978, and 1979), Griepentrog and Groeneveld (1981), and Hutchison (1986b).

Each of the water budgets, except that of Hutchison (1986b), was reviewed by Danskin (1988). In comparing the respective components of inflow and outflow, he noted that comparisons were difficult because each of the studies covered different areas or different periods of time. In addition, some of the water budgets used the same components of inflow and outflow, but with different definitions. A complete analysis of the hydrologic system of the Owens Valley, he concluded, would require at least three interrelated water budgets for the valley-fill part of the drainage basin area—a total budget for both saturated and unsaturated materials, including all precipitation and evapotranspiration; a budget for the surface-water system; and a budget for the ground-water system. To facilitate verification and comparisons, the budgets would need to cover the same area and time period and use similarly defined components.

The synthesis of three complex, interrelated water budgets was outside the scope of this study; however, significant progress in that direction has been made by development of a detailed ground-water budget (tables 10 and 11) [table 11 in pocket]. In addition, data have been collected and summarized and predictive relations have been developed for precipitation, evapotranspiration, and tributary streamflow. Eventual development of the three interrelated budgets would be needed to further refine the ground-water budget presented in this report.

The ground-water budget for the defined aquifer system shown in figure 14 is summarized in table 10. Each component of the ground-water budget is defined and discussed more fully by Hollett and others (1991). The values in table 10 are revised slightly from those presented by Hollett and others (1991, table 6), but they were developed using identical concepts and methods. Development of the ground-water budget involved using data from previous studies, new evapotranspiration and stream-loss data collected during this 6-year study, and results of simulation of the aquifer system described later in this report.

Average values for each component are given in table 10 for two time periods, water years 1963–69 and water years 1970–84. The first period represents average conditions in the aquifer system prior to increased pumpage and additional export of water from the valley (table 4). The second period represents conditions after pumpage and exports increased. The uncertainty of each value for the second period was estimated, and the likely range of values is given.

Ground-water budgets, such as the two given in table 10, can be useful in making semi-quantitative evaluations of an aquifer system, but budgets can be misinterpreted or misused quite easily (Bredehoeft and others, 1982). For example, the approximation of equilibrium is rarely satisfied over an entire system that has been modified by human activity. Localized areas in the Owens Valley likely will be undergoing change for years or decades as a result of human intervention. Changes in recharge or discharge, such as occurred in 1913 and 1970, are reflected in changes in the magnitude of several different components of the water budget (compare tables 4 and 10). In general, the interaction between the components is complex and the magnitude of the changes to the hydrologic system cannot be estimated from the budget alone. For this reason, numerical simulation is a critical part of understanding the operation of the aquifer system and the potential effects of water-management decisions.

The following components of the ground-water budget are not linked to a specific surface-water feature and were not discussed in previous sections of this report.

Discharge from Pumped and Flowing Wells

Discharge from wells includes discharge from both pumped and flowing wells, although the quantity from flowing wells is much less and is limited to a few wells along the Owens River south of Bishop and a few

Table 10. Ground-water budget for the aquifer system of the Owens Valley, California [1]

[Values in acre-feet per year. Positive numbers indicate recharge to the aquifer system; negative numbers ( ) indicate discharge from the aquifer system]

Component	Average values		Likely range of average values for water years 1970–84	
	Water years 1963–69	Water years 1970–84	Minimum	Maximum
Precipitation	2,000	2,000	0	5,000
Evapotranspiration	(112,000)	(72,000)	(50,000)	(90,000)
Tributary streams	106,000	103,000	90,000	115,000
Mountain-front recharge between tributary streams	26,000	26,000	15,000	35,000
Runoff from bedrock outcrops within the valley fill	1,000	1,000	0	2,000
Owens River and Los Angeles Aqueduct system:				
Channel seepage	(16,000)	(3,000)	0	(20,000)
Spillgates	6,000	6,000	3,000	10,000
Lower Owens River	(5,000)	(3,000)	(1,000)	(8,000)
Reservoirs and small lakes	1,000	1,000	(5,000)	5,000
Canals, ditches, and ponds	32,000	31,000	15,000	60,000
Irrigation and watering of livestock	18,000	10,000	5,000	20,000
Pumped and flowing wells	(20,000)	(98,000)	(90,000)	(110,000)
Springs and seeps	(26,000)	(6,000)	(4,000)	(10,000)
Underflow:				
Into the aquifer system	4,000	4,000	3,000	10,000
Out of the aquifer system	(10,000)	(10,000)	(5,000)	(20,000)
Total recharge	196,000	184,000	170,000	210,000
Total discharge	(189,000)	(192,000)	(175,000)	(225,000)
Change in ground-water storage[2]	7,000	(8,000)	(5,000)	(15,000)

[1] Values of water-budget components for individual years may vary considerably from the average values presented in this table. Uncertainties in the measurement and estimation of each water-budget component for water years 1970–84 are reflected in the likely range of average values. The likely ranges for total recharge, total discharge, and change in ground-water storage are estimated separately for the overall aquifer system and are somewhat less than what would be computed by summing the individual ranges for respective water-budget components.

[2] Positive change in storage indicates water going into ground-water storage; negative ( ) change in storage indicates water coming out of ground-water storage.

wells in the Independence area near the aqueduct. Several of the flowing wells also are equipped with pumps, and thus discharge sometimes is free-flowing ground water and sometimes is pumped ground water. In this report, all discharge from pumped and flowing wells is referred to informally as “ground-water pumpage.”

Nearly all ground-water pumpage is from production wells owned and operated by the Los Angeles Department of Water and Power. Most of these wells provide water for export; a few wells supply water for ranching operations and to the four major towns; and four large-capacity wells supply water to two fish hatcheries. Some additional pumpage is from private domestic and agricultural wells. Distribution of the wells (fig. 17) generally follows the river–aqueduct system. In fact, a few of the present production wells were installed in the early 1900's for dewatering and water supply during construction of the first aqueduct. Division of the wells into well fields shown in figure 17 was done on the basis of general location of the wells and included all wells with production during water years 1963–88, as reported by the Los Angeles Department of Water and Power (M.L. Blevins, written commun., 1988; table 11). The well fields identified in figure 17 and used elsewhere in this report are similar to those defined by the Los Angeles Department of

Water and Power (1979, fig. 4-4; Hollett and others, 1991, fig. 18).

Annual pumpage for individual wells for water years 1963 through 1988 was obtained from the Los Angeles Department of Water and Power (M.L. Blevins, written commun., 1988). Pumpage for water years 1963–69 was copied from typed summary sheets of well discharge per month. Pumpage for water years 1970–71 was estimated by interpolating between instantaneous discharge readings for each well. Pumpage for water years 1972–88 was obtained directly from computerized files.

Average pumpage in most areas of the Owens Valley changed dramatically after 1970, as shown by the inset graphs of well-field discharge in figure 17. Within the defined aquifer system (fig. 14), total pumpage averaged about 20,000 acre-ft/yr during water years 1963–69 and about 98,000 acre-ft/yr during water years 1970–84 (table 10). Much of this increase was caused by the switching from surface to ground water by two major fish hatcheries. The fish hatcheries, Fish Springs and Blackrock, are located near Fish Springs and Big Blackrock Springs, respectively (fig. 17). Average pumpage changed again in 1987 with the addition of new "enhancement and mitigation" wells, which were used to provide water for selected recreation and wildlife projects throughout the Owens Valley (table 4; Los Angeles and Inyo County, 1990a, p. 5–20).

The total quantity of ground-water pumpage varies each year with the quantity of runoff. In years of greater runoff, less pumpage is required for in-valley uses or for export. Pumpage also depends on the quantity of runoff in the preceding year, as shown in figure 18. When antecedent conditions are wet, the river–aqueduct system is full, and pumpage is less.

Discharge from different hydrogeologic units was investigated by analyzing each well. The first significant clay layer, as identified from the lithologic well log, was used to mark the separation between hydrogeologic units 1 and 3. Discharge from each well then was apportioned as withdrawal from hydrogeologic units 1 and 3 (upper and lower model layers) on the basis of length of perforations and estimated hydraulic conductivity of the adjacent material in hydrogeologic units 1 and 3, respectively. In most parts of the valley, well withdrawals are primarily from hydrogeologic unit 3 (fig. 17). Near the Big Pine volcanic field, many wells tend to be shallow, and most water is withdrawn from the highly transmissive volcanic deposits near the land surface (figs. 4 and 5).

Springs and Seeps

Most springs in the Owens Valley are near the toes of alluvial fans and along the edge of volcanic deposits near the Poverty Hills (fig. 17). A few springs are caused by faulting as indicated by an obvious surface trace (fig. 3; Hollett and others, 1991, fig. 15). Historically, springs have discharged a large quantity of water, most of which eventually flowed into the river–aqueduct system. For example, Fish Springs near Crater Mountain discharged as much as 22 ft^3/s prior to 1970. When ground-water pumpage increased in 1970, discharge at springs dropped dramatically, to zero at some. Average discharge from major springs within the defined aquifer system was about 33,000 acre-ft/yr during water years 1963–69 and about 8,000 acre-ft/yr during water years 1970–84. About 20 percent of this discharge was estimated to return to the aquifer system as recharge in the immediate vicinity of the springs (Hollett and others, 1991). Net discharge from the aquifer system was about 26,000 and 6,000 acre-ft/yr for the two periods, respectively (table 10).

Seeps occur along some faults where ground-water flow is forced to the land surface and along the toes of alluvial fans where ground water flows out onto the valley floor. The major seeps (shown in figures 3 and 17) discharge an unknown quantity of water, nearly all of which is evapotranspired by nearby vegetation.

Springs and to a lesser extent seeps, such as the Independence "springfield" (fig. 17), act as hydraulic buffers and exert a strong local influence on the aquifer system. The maximum altitude of the water table, particularly near the Poverty Hills, is controlled by the altitude of nearby springs and the transmissive properties of the adjacent deposits (figs. 14, 15, and 17). Fish Springs, for example, prior to an increase in nearby pumpage in 1970, was exceptionally effective at dampening fluctuations in nearby ground-water levels [well 224, pl. 1 (in pocket)]. In the Big Pine area, an increase in recharge to the aquifer resulted in an increase in discharge from Fish Springs and only a minimal rise in ground-water levels near the spring; a decrease in recharge to the aquifer resulted in a decrease in discharge from Fish Springs and only a minimal decline in ground-water levels near the spring. After 1970, the buffering effect of springs near the Poverty Hills (fig. 17) was reduced, and changes in

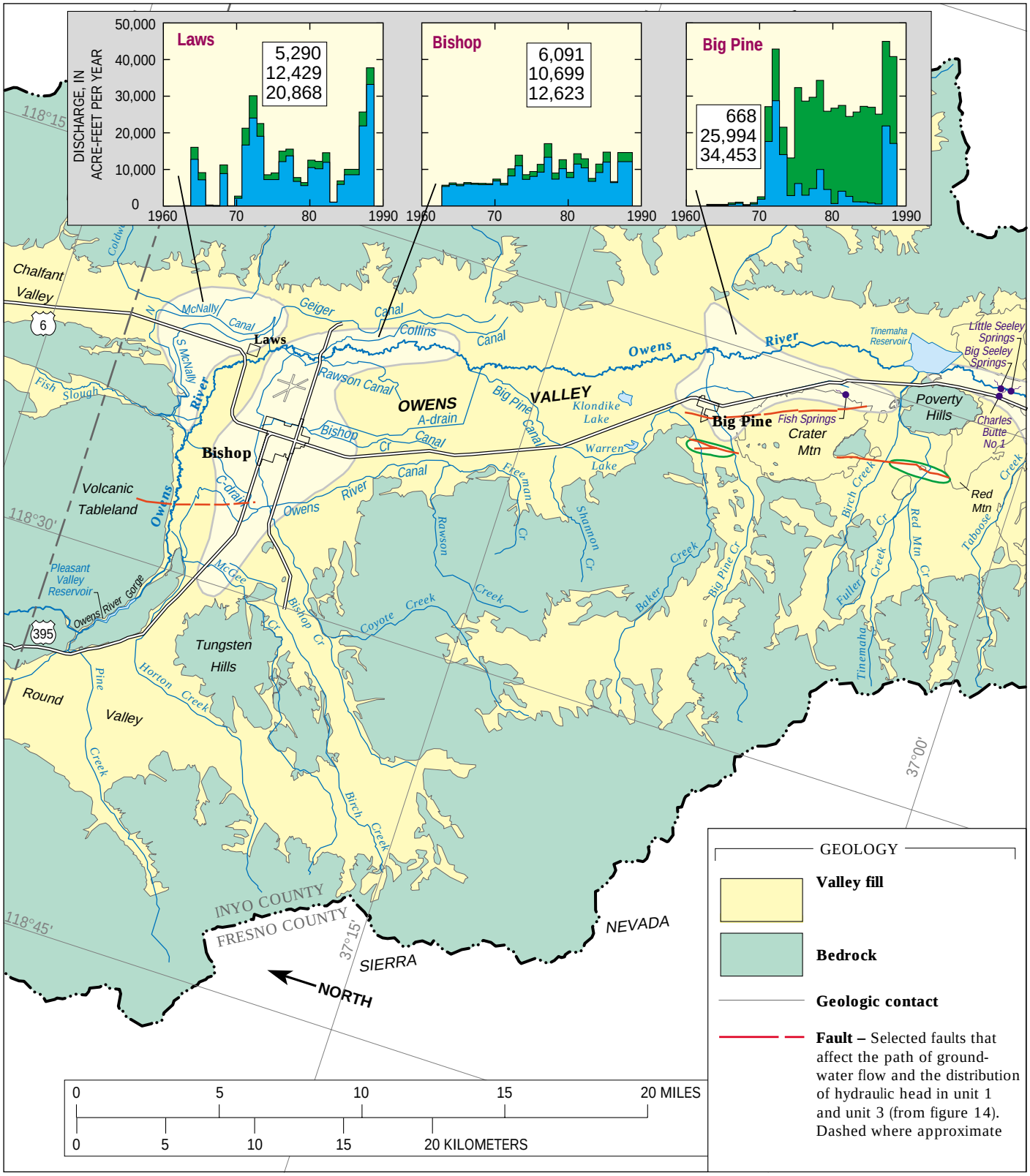


Figure 17. Location of springs, seeps, pumped or flowing wells, and approximate area of well fields in the Owens Valley, California. Inset graphs show annual discharge from each well field for water years 1963–88.

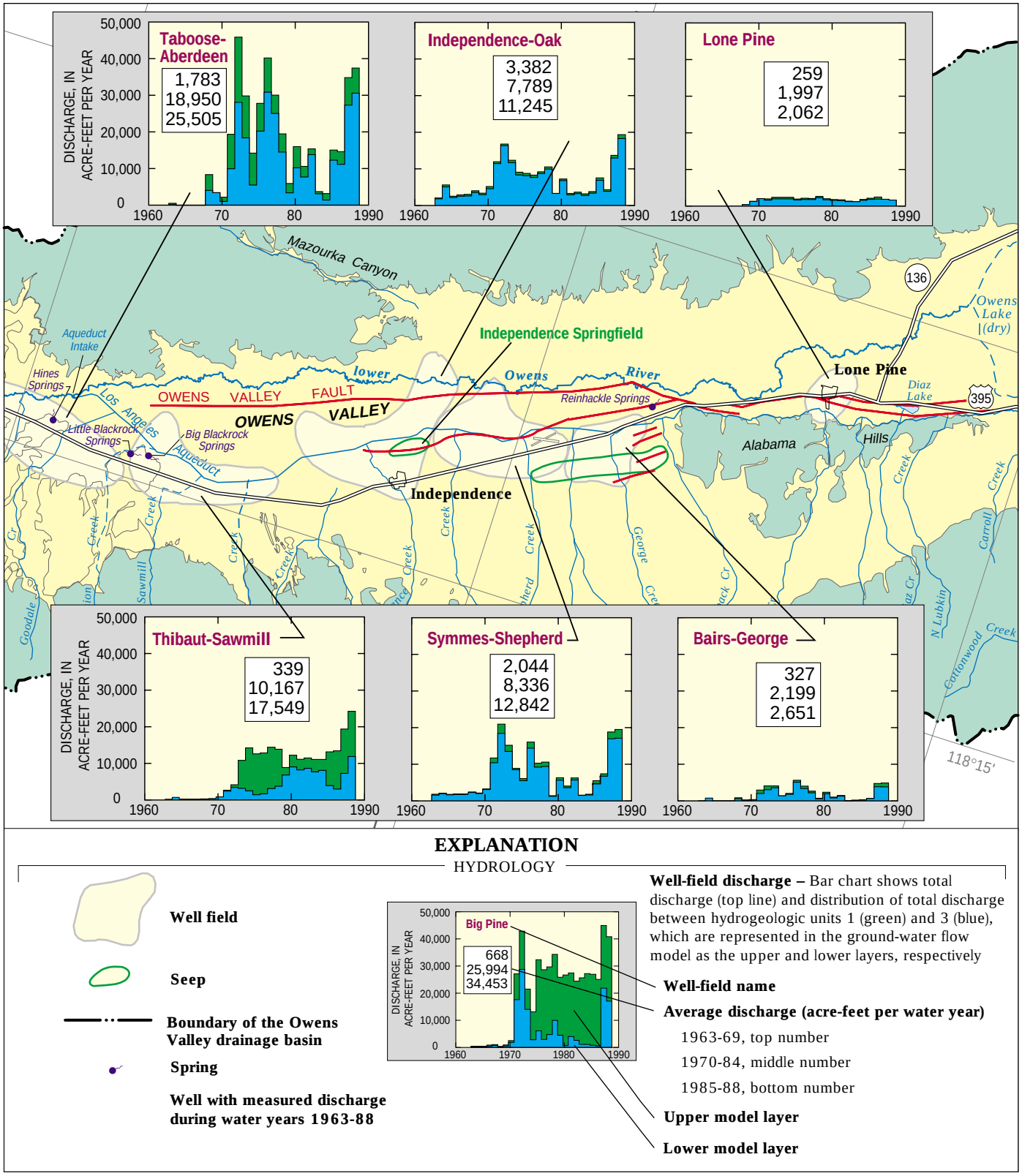


Figure 17. Continued.

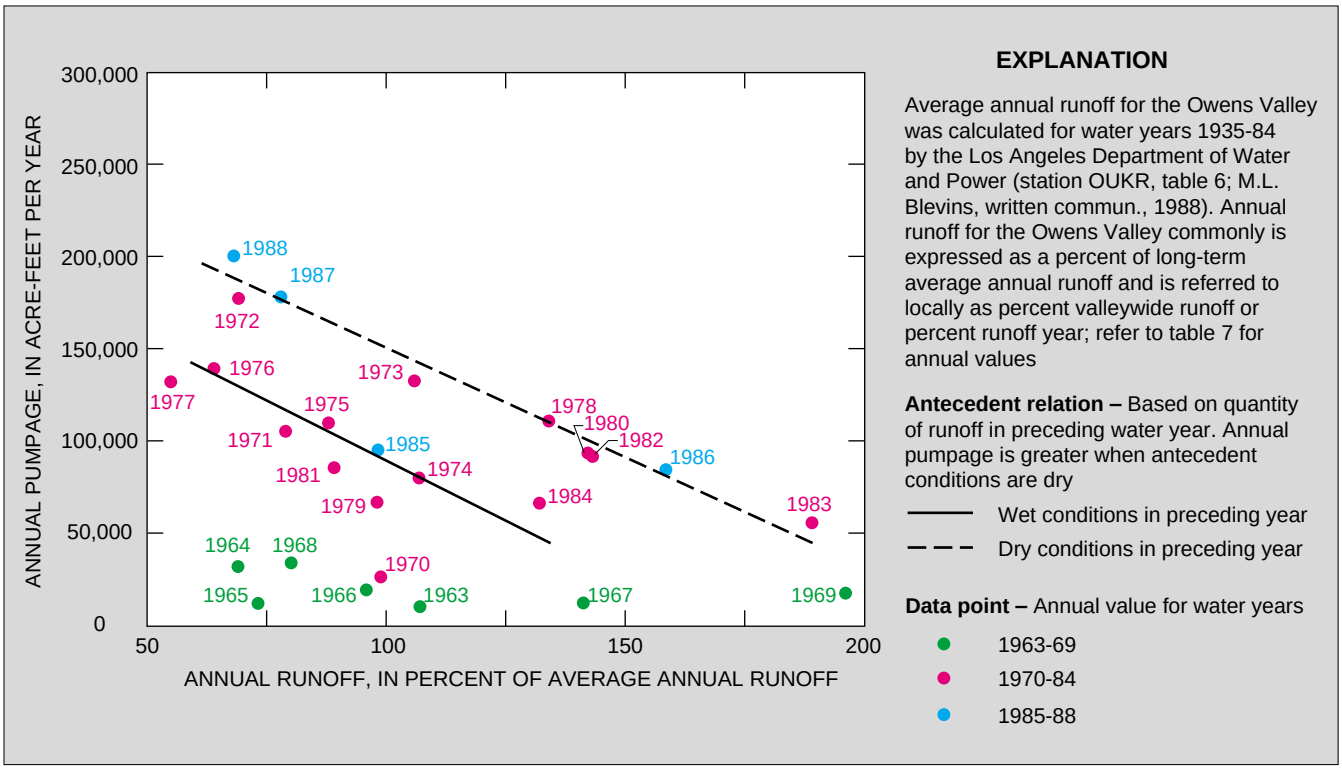


Figure 18. Relation between annual pumpage and annual runoff for the Owens Valley, California.

aquifer recharge and discharge resulted in greater fluctuations in ground-water levels.

Underflow

Underflow into and out of the aquifer system occurs at several locations shown in figure 14. Underflow from three drainages (Bishop and Big Pine Creeks and Waucoba Canyon) originates as recharge from tributary streams outside the aquifer system. For that reason, the quantity of underflow from those areas, totaling about 500 acre-ft/yr, is included for water-budget purposes as part of tributary stream recharge (table 10).

The quantity of underflow from Round Valley, the Volcanic Tableland, and Chalfant Valley is much greater and was estimated to average about 4,000 acre-ft/yr (table 10). Prior estimates of underflow from these areas were significantly higher, totaling as much as 25,000 acre-ft/yr. These estimates were based on Darcy's law (Los Angeles Department of Water and Power 1972, 1976, 1978, 1979) and on steady-state ground-water-model simulations (Danskin, 1988). As shown in table 10, the quantity of underflow into the aquifer system is not known with certainty. However, the present estimates, which are consistent with results from several different ground-water flow models developed during the cooperative USGS studies, probably are more accurate than previous estimates. The models also are based on Darcy's law, but they have additional advantages; these include incorporating nearby ground-water recharge and discharge, accounting for changes in ground-water storage, and matching various historical conditions (calibration).

Underflow out of the aquifer system occurs only across an arbitrary east–west line south of Lone Pine. In the area east of the Alabama Hills, most ground water flows out of the aquifer system through hydrogeologic unit 3, which is thicker and more transmissive than hydrogeologic unit 1. In the area west of the Alabama Hills, hydrogeologic units 1 and 3 act together, and there is no clear distinction between the two units, or indication of the relative quantity of underflow from each. Total underflow from both areas was estimated to be about 10,000 acre-ft/yr. This estimate is based on calibration of the valleywide ground-water flow model and on a water-budget analysis of the Owens Lake area by Lopes (1988). No difference in the quantity of underflow before and after 1970 was detected (table 10).

Irrigation and Watering of Livestock

Irrigation of agricultural and pasture land is still (1988) prevalent in the Owens Valley (fig. 3), although the total acreage of irrigated lands and the quantity of water applied to irrigated lands is much less than in previous years (D.E. Babb and R.H. Rawson, Los Angeles Department of Water and Power, written commun., 1988). The most recent change in water-management practices in the Owens Valley occurred in about 1968 in anticipation of providing sufficient water to fill the second aqueduct (table 4). Some land was taken out of production. Historical agricultural practices that resulted in an excessive application of water, such as using flood irrigation, were discouraged. Fields were leveled and irrigation sprinklers were installed. Water supplied by the Los Angeles Department of Water and Power to lessees was reduced from about 6 acre-ft/acre to about 5 acre-ft/acre. Watering of livestock, which typically involves diverting surface water from a canal or ditch and flooding a small area of the land surface, continued, but to a lesser degree. As a result, the total recharge from both irrigation and stock watering decreased, and the salvaged water was available for export.

Recharge to the aquifer system from irrigation and watering of livestock was estimated from maps of land use compiled by the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1988). Digitized map information was combined with assumptions about the quantity of water supplied and used per acre and the likely recharge rates on different types of soils. For years prior to 1970, water applied on volcanic materials was assumed to recharge at a rate of 24 in/yr, and water applied on other permeable materials, at a rate of 12 in/yr. For 1970–84, these rates were reduced to 12 in/yr and 6 in/yr, respectively. On the basis of these assumptions, the average recharge from irrigation and watering of livestock within the aquifer system (fig. 14) was estimated to be about 18,000 acre-ft/yr in water years 1963–69 and about 10,000 acre-ft/yr in water years 1970–84 (table 10).

Ground-Water Quality

Ground water in most parts of the Owens Valley has a preponderance of calcium and bicarbonate ions, and the range of concentrations for dissolved constituents is small (Hollett and others, 1991, fig. 21). Concentrations of dissolved solids are generally less than 300 mg/L. However, at the extreme southern end of the basin near the Owens Lake, ground-water quality is much different. A well named “Dirty Socks” (Hollett and others, 1991, fig. 18) was found to have markedly different water quality—mostly sodium, chloride, and bicarbonate ions and a concentration of dissolved solids greater than 5,000 mg/L.

In 1973–74, the Los Angeles Department of Water and Power (1974a) conducted an areally extensive study of ground-water quality that included samples from selected wells in each well field (fig. 17). Although the study focused primarily on drinking-water standards (California Department of Health Services, 1983; U.S. Environmental Protection Agency, 1977a, b, 1986), results did not reflect any major differences in ground-water quality throughout most of the valley. It was also concluded in the study that no significant changes have occurred in ground-water quality in the valley during the past 10 to 35 years.

One area of exception was noted, however. On the basis of earlier data, ground-water quality just south of the Tinemaha Reservoir seemed to be different and possibly changing from 1972 to 1973 (Roland Triay, Jr., Los Angeles Department of Water and Power, written commun., 1973). Alkalinity for wells near the Taboose–Aberdeen well field (table 9, wells 118, 349, and 116) increased between June 1972 and April 1973 by as much as 90 percent. One possible explanation is that the extensive pumping from 1970 to 1973 (fig. 17) induced movement of water from the east side of the valley toward the Taboose–Aberdeen well field. Ground water in contact with sedimentary and metamorphic rocks along the east side of the valley likely has a higher concentration of dissolved solids and a higher alkalinity than does ground water in contact with granitic rocks and near the dominant recharge areas on the west side of the valley. The significant drawdown observed at nearby wells (pl. 1, wells 362 and 347), a steep hydraulic gradient from east to west, and a pattern of increasing dissolved-solids concentration from west to east lend credibility to this explanation.

Another possible explanation is that dissolution and mobilization of soluble minerals in nearby fine-grained deposits caused the observed changes in ground-water quality (Roland Triay, Jr., Los Angeles Department of Water and Power, written commun., 1973). Also, the increased hydraulic gradient may have induced vertical movement of ground water of different quality from an adjacent part of the aquifer. Additional localized water-quality studies would help in identifying the specific flow paths of ground-water

movement, particularly as influenced by pumping and artificial recharge.

More generally, a complete inventory of ground-water quality in the Owens Valley is needed to confirm ground-water concepts presented in this report and by Hollett and others (1991). Many of the older wells are open to a combination of hydrogeologic units 1, 2, and 3. Water-quality data from these wells are ambiguous and difficult to interpret. Recently installed production and observation wells that are open only to specific strata offer the opportunity to sample ground-water quality for specific hydrogeologic units of the aquifer system. Also, some of the new wells are located near and some far from areas of recharge and discharge. Water-quality information from these new wells could aid considerably in confirming the areal and vertical ground-water flow paths (fig. 14), and in identifying likely changes in flow paths. The water-quality characteristics of interest are major and minor ions; trace metals; nitrate and nitrite; hydrogen, oxygen, and carbon isotopes to date the water and identify different sources of recharge; and possibly pesticides or organic contaminants to document issues of public health.

Studies of oxygen- and hydrogen-isotope concentrations across much of southern California by Gleason and others (1994) revealed strong regional differences. Ground water from eight wells in the Owens Valley had less deuterium (that is, was much "lighter" in hydrogen isotopes) than did ground water in basins to the east and south. This trend implies that the dominant recharge to the Owens Valley ground-water basin comes from precipitation from storms that are moving westward. No trend within the Owens Valley could be detected from the scant number of samples. Although storm cells originating to the south may be important in providing water for native vegetation, the quantity of recharge to the ground-water system from such storms is much less than the quantity of recharge resulting from runoff from the Sierra Nevada.

Ground-Water Flow Model

A valleywide ground-water flow model was developed to integrate and test the concepts about the structure and physical properties of the aquifer system, the quantity of recharge and discharge, and the likely effects of water-management decisions. A numerical ground-water flow model, such as the valleywide model, is a group of mathematical equations that describe the flow of water through an aquifer. Variables (parameters) in the equations include hydraulic heads, transmissive characteristics, storage characteristics, and the rates of inflow and outflow. Different values for each variable, such as transmissivity or pumpage, can be distributed throughout the area being modeled in order to simulate observed spatial and temporal variations. This general technique is referred to as a distributed-parameter approach in contrast to a lumped approach, which uses a single value for each type of parameter.

Even when using a distributed-parameter approach, however, not all characteristics of the actual aquifer system can be included in the ground-water flow model. Simplifying assumptions are required to make the modeling effort manageable. Many of the assumptions used in developing the Owens Valley ground-water flow model are characteristic of most numerical ground-water flow models. Explanations of these assumptions are given by Remson and others (1971), Durbin (1978), Freeze and Cherry (1979), Wang and Anderson (1982), and Franke and others (1987). Assumptions underlying the particular computer program used in this study are described by McDonald and Harbaugh (1988). Additional assumptions made in the application of the computer program to the Owens Valley aquifer system are discussed in the next sections of this report.

For purposes of clarity in this report, hydraulic head (head) is used when referring to simulated hydraulic potential, which is well defined and has a precise x–y–z location. Ground-water level (level) is used when referring to general concepts of ground-water flow and to measured data, which are less well defined vertically and often represent a composite hydraulic potential.

Although a simulation model is only an approximation of the real world, it can be extremely useful in gaining an improved understanding of a complex system—in this case, a ground-water system interacting with many surface-water features. A ground-water flow model assures that estimates of local aquifer characteristics, the water budget, and hydraulic heads all are compatible. It is this attribute that gives additional confidence in the concepts and quantities presented in this report and in those described by Hollett and others (1991). In areas where data are sparse or uncertain, the ground-water flow model can be used to test the reasonableness of assumed values. Finally, a calibrated model—one for which all the parameter values are acceptable—can be

used to compare the likely effects of different water-management alternatives.

General Characteristics

The computer program developed by McDonald and Harbaugh (1988) uses standard finite-difference techniques to approximate the partial differential equations that describe saturated ground-water flow. General characteristics of the numerical code include division of a ground-water system into finite-difference cells, each with uniform hydraulic properties. Multiple layers can be identified and linked with Darcy's law. A variety of different types of recharge and discharge can be simulated with constant-head, head-dependent, or specified-flux terms. Transmissivity can be constant or calculated as the product of hydraulic conductivity and saturated thickness. Both steady-state and transient conditions can be simulated, each with its own formulation. Several solvers are available, including those provided by Hill (1990a,b) and Kuiper (1987a,b) that constrain convergence of the solution using both head and mass-balance terms. The computer code is stable and flexible, and it is widely used in the public and private sectors.

Application of the numerical code to the aquifer system of the Owens Valley involved the use of two model layers. Flow between the layers was approximated by a relation that uses calculated head in vertically adjacent cells and an estimate of “vertical conductance” between the cells. Vertical conductance is calculated from vertical hydraulic conductivity, thickness between the layers, and horizontal area of the cell (McDonald and Harbaugh, 1988, p. 5–11). Transmissivity was varied between groups of model cells (model zones), but was assumed to remain constant over time. Specified flux terms were used to approximate discharge from wells and recharge from precipitation, tributary streams, canals, and ditches. Head-dependent relations were used to simulate springs, evapotranspiration, and interaction of the aquifer system with the river–aqueduct system and the lower Owens River. A 26-year simulation period included water years 1963–88 and used annual approximations of recharge and discharge.

A geographic information system (GIS) was developed to ensure an accurate spatial control of physical features and the finite-difference model grid. This accuracy was critical in linking map information, such as the vegetative mapping by the Los Angeles Department of Water and Power (fig. 9), the valleywide ground-water flow model, and the several more detailed ground-water flow models developed by Inyo County and the Los Angeles Department of Water and Power (table 2). The original digitizing of geologic and hydrologic information was done in latitude and longitude coordinates, using the North American Datum 1929, from maps with scales of 1:24,000 and 1:62,500. Replotting was done using a Universal Transverse Mercator (UTM) projection (Newton, 1985). This GIS methodology was used for all map illustrations in this report and in Hollett and others (1991). Because of the accuracy of the GIS method, subsequent computer scanning of the map illustrations should produce an accuracy of approximately 0.01 in. and permit registration with other maps drawn from a UTM projection. Detailed information on GIS and UTM mapping systems is given by J.P. Snyder (1982, 1985, 1987) and Newton (1985).

As part of the GIS system, the finite-difference model grid was linked mathematically to latitude and longitude and the UTM coordinate system. Coordinates of the finite-difference model grid are given in table 12. Projection and translation of coordinate systems (latitude-longitude, UTM, model) were done using computer programs based on those developed by Newton (1985). Use of the coordinates in table 12 and similar computer projection programs will enable future investigators to reproduce the model locations precisely. Use of this technique reduces any differences caused solely by spatial discretization and aids in duplicating specific results presented in this report.

Representation of the Aquifer System

Boundaries of the ground-water flow model conform to the physical boundaries of the Owens Valley aquifer system as shown in figure 14 and as described by Hollett and others (1991). Lateral underflow boundaries are present in eight locations: Chalfant Valley, the edge of the Volcanic Tableland, Round Valley, Bishop Creek, Big Pine Creek, Waucoba Canyon, and east and west of the Alabama Hills. All other boundaries of the aquifer system were assumed to be impermeable and were simulated with no-flow boundary conditions. The top of the aquifer system is the water table, and the bottom is either bedrock, the top of a partly consolidated unit, or an arbitrary depth based on the depth of production wells. Hydrogeologic unit 4 (fig. 5) lies below the aquifer system in the center of the valley and is a poorly transmissive part of the ground-water system. Simulation studies by Danskin (1988) concluded that this unit could be eliminated

Table 12. Map coordinates for the ground-water flow model of the aquifer system of the Owens Valley, California

[Coordinates are calculated at the outside edge of the finite-difference model grid]

Corner of model grid	Map coordinates				
	Model grid (row, column)	Latitude (north) (decimal value in parentheses)	Longitude (west) (decimal value in parentheses)	Universal Transverse Mercator (UTM) coordinates, zone 11, in meters	
Northwest	(0.0, 0.0)	37° 26' 14" (37.4371)	118° 34' 12" (118.5700)	361,101	4,144,319
Northeast	(0.0, 40.0)	37° 30' 16" (37.5044)	118° 18' 27" (118.3076)	384,423	4,151,436
Southwest	(180.0, 0.0)	36° 29' 44" (36.4955)	118° 11' 36" (118.1933)	393,126	4,039,368
Southeast	(180.0, 40.0)	36° 33' 43" (36.5619)	117° 56' 01" (117.9337)	416,449	4,046,485

from future ground-water flow models with little loss of accuracy in the upper 1,000 ft of more transmissive materials. Round Valley and the Owens Lake area also were excluded as suggested by Danskin (1988), primarily for computational reasons and because the areas were peripheral to the specific objectives of this study. Future simulation studies with more powerful computer capabilities may find that including both areas is an advantage in analyzing some water-management questions as well as in eliminating the use of specified-flux boundary conditions.

Division of the aquifer system into hydrogeologic units and model layers is more complex and somewhat more arbitrary than the selection of boundary conditions. For this study, the aquifer system was simulated using two model layers. The upper model layer (layer 1) represents hydrogeologic unit 1, the unconfined part of the aquifer system. The lower model layer (layer 2) represents hydrogeologic unit 3, the confined part of the aquifer system. Each model layer is composed of 7,200 cells created by 180 rows and 40 columns (pl. 2, in pocket). The active area of ground-water flow (active model cells) is the same in both model layers.

This division of the aquifer system permits simulation of the measured ground-water levels, which generally are either for shallow wells that monitor unconfined conditions or for deeper wells that monitor a composite confined zone. The use of two layers is consistent with the assumption that both unconfined and confined storage conditions are present in some parts of the valley (fig. 14).

To test the value of additional model layers, a smaller, more detailed ground-water flow model was developed to simulate conditions in the Big Pine area (P.D. Rogalsky, Los Angeles Department of Water and Power, written commun., 1988). Although three layers were used in the model in order to more closely approximate the complex layering of volcanic and fluvial deposits described by Hollett and others (1991), results from the more detailed model were not significantly different from results obtained using the valleywide model.

Hydrogeologic unit 2, as defined by Hollett and others (1991), usually represents either a massive clay bed, such as the blue-green clay near Big Pine (fig. 5, section *B–B'*), or overlapping lenses or beds, which are more typical of the valley fill. The Darcian relation that simulates vertical flow between the model layers was used to approximate the vertically transmissive properties of hydrogeologic unit 2. Storage characteristics of hydrogeologic unit 2 were included in the storage coefficients of the surrounding model layers. This formulation is typical of most models used to simulate ground-water movement in unconsolidated, poorly stratified deposits, such as those in the Owens Valley (Hanson and others, 1990; Berenbrock and Martin, 1991; and Londquist and Martin, 1991).

Along the edge of the basin, the clay beds thin, and hydrogeologic unit 2 virtually disappears (fig. 5, section *C–C'*). In these areas, a high value of vertical conductance was used, allowing water to move between the model layers with minimal resistance. The spatial distribution of vertical conductance and its relation to hydrogeologic model zones are shown on plate 2.

In some parts of the valley, hydrogeologic unit 2 represents volcanic deposits, such as those near Big Pine (section *B–B'* in fig. 5). The volcanic deposits have a high transmissivity but can restrict the vertical movement of water as a result of the depositional layering of individual volcanic flows. Where faulted or highly

brecciated, the volcanic deposits of hydrogeologic unit 2 were represented by a high value of vertical conductance. As with other deposits represented by hydrogeologic unit 2, the transmissivity of the volcanic deposits was included in the model layer that best approximates the storage properties of the deposit—usually the upper model layer, which represents unconfined conditions.

To facilitate modeling, the aquifer system was divided into model zones, each representing part of a hydrogeologic unit or subunit (Hollett and others, 1991, pl. 2). This technique was shown to be effective in preliminary model evaluations (Danskin, 1988), although the use of additional model zones was suggested in order to simulate key areas of the basin, such as along the toes of alluvial fans. Therefore, development of the valleywide model included additional model zones—specifically, zones to represent the transition-zone deposits. Each model zone represents similar geologic materials that have fairly uniform hydraulic properties. In the volcanic areas of the basin, maintaining this uniformity was not possible. Instead, a single model zone included highly transmissive volcanic deposits along with other much less transmissive fluvial deposits (fig. 5). For these zones, the presence of volcanic deposits dominated the hydraulic properties. Outcrops of volcanic flows and cinder cones on the land surface identified likely locations of volcanic deposits in the subsurface. The actual presence of volcanic deposits was confirmed using lithologic information from well logs wherever possible. Calibration of the model was necessary to refine the locations and hydraulic properties of the volcanic zones.

A likely range of transmissivity for each model zone was determined by using the values given in table 9 and the distribution shown in figure 15. In some areas of the basin, however, little or no data were available. In these areas, the depositional models described by Hollett and others (1991, fig. 14) were used to extrapolate data and concepts. This technique based on general depositional models with specific data points throughout the aquifer system worked surprisingly well. Values of average horizontal hydraulic conductivity (fig. 16) times estimated saturated thickness were compared with estimated transmissivity values in each zone in order to ensure consistency of hydraulic conductivity, saturated thickness, and transmissivity. Other methods of interpolating transmissivity, such as kriging (Journel and Huijbregts, 1978; Sampson, 1978, 1988; Yeh, 1986), were evaluated and found to be of little use in the faulted, complex structure of the Owens Valley (figs. 4 and 5).

The transmissivity of volcanic areas was determined by means of arithmetic weighting of the estimated hydraulic conductivity and thickness of volcanic deposits with that of the surrounding sand, gravel, and silt deposits. Not surprisingly, the exceptionally transmissive volcanic deposits dominated the value of all zones where they were present (pl. 2). Only a few electric logs were available, but lithologic well logs were of great value in identifying the general type of depositional material and its appropriate zone.

Transmissivity in all areas of the model was assumed to remain constant over time (pl. 2). This assumption implies that saturated thickness of the model layer—particularly the upper, water-table layer—does not change significantly during model simulations. Changes in saturated thickness may result in differences in computed heads as a result of a mathematical nonlinearity in the ground-water-flow equations (Bear, 1979, p. 308). Because of the relative thinness of hydrogeologic unit 1, a 20-foot change in saturated thickness of unit 1 produces a 10-percent greater fluctuation in nearby water-table altitude than that predicted by the model. The modeling option to vary transmissivity over time (McDonald and Harbaugh, 1988, p. 5–10), however, creates its own set of problems. These problems include the need for significantly more detailed data for model construction and the conversion from active to inactive model cells when dewatered conditions are simulated. For the Owens valleywide model, these problems outweighed the benefits gained by varying transmissivity over time.

Vertical conductance between the two model layers was estimated from aquifer tests, development of preliminary dewatering and cross-sectional models (fig. 2), and calibration of the final valleywide model. A high correlation was found between the value of vertical conductance and the type of material in the lower model layer. In most instances, the thicker lower model layer contributed most of the impediment to vertical ground-water flow. As a result, the values of vertical conductance were keyed to the model zones representing the lower model layer (pl. 2).

Faults that restrict ground-water movement (fig. 14) were represented by lower values of transmissivity in model cells. The ratio of reduced transmissivity caused by the fault to transmissivity of adjacent aquifer materials is noted on plate 2. For example, a section of the Owens Valley Fault (F20) was

determined to reduce transmissivity of the aquifer materials for that zone by a factor of 20—from 80,000 to 4,000 (gal/d)/ft.

Approximation of Recharge and Discharge

The physical characteristics of recharge to and discharge from the aquifer system are described in detail in earlier sections of this report, specifically in the sections entitled "Surface-Water System" and "Ground-Water Budget." The following discussion describes only the approximations of ground-water recharge and discharge that were made in order to simulate these processes in the ground-water flow model. The type of boundary condition and method of approximation for each recharge and discharge component are given in table 13. Annual values for each component for water years 1963–88 are given in table 11, along with the derivation of the value (measured, estimated, or calculated by the model). The areal distribution of each recharge or discharge component in the model and the average values for each model cell for water years 1970–84 are shown on plate 3 (in pocket).

Well package.—Most of the recharge and discharge components were simulated using the well package of McDonald and Harbaugh (1988, p. 8–1). This package simulates extraction of a defined quantity of water from a specific cell in the ground-water flow model. Annual estimates for several recharge and discharge components (table 13) were combined in a pre-processing program, and the net result was used as input for the well package. In most areas of the model, only a few values in the well package represent actual discharge from wells (pl. 3*F*). Estimated flux for individual items, such as for a stream or an area of ground-water recharge, was distributed uniformly to all model cells related to that item. For example, recharge for a specific stream was the same for each model cell along its length. The individual items are listed in table 11. A few components (precipitation, spillways, and underflow) were assumed to have a virtually constant recharge or discharge rate from one year to another, and were simulated with a constant value for water

Table 13. Recharge and discharge approximations for the ground-water flow model of the aquifer system of the Owens Valley, California

[Type of boundary condition: Franke and others (1987). Ground-water flow model approximation: McDonald and Harbaugh (1988). Recharge and discharge components defined in text. Temporal variation in stress: A, annually varying rate; C, constant rate; $\overline{C}$, constant rate for several years]

Type of boundary condition	Ground-water flow model approximation	Recharge (R) or discharge (D) component	Temporal variation in stress
Specified flux	Well package	Precipitation (R)	C
		Spillgate releases (R)	C
		Underflow (R,D)	C
		Canals and ditches (R)	$\overline{C}$
		Irrigation (R)	$\overline{C}$
		Watering of livestock (R)	$\overline{C}$
		Tributary streams (R)	A
		Miscellaneous water use (R)	A
		Mountain-front runoff (R)	A
		Pumpage (D)	A
		Runoff from bedrock within the valley (R)	A
Head-dependent flux	River package	Lakes (R,D)	A
		Lower Owens River (R,D)	A
		River–aqueduct system (R,D)	A
		Sewage ponds (R,D)	A
		Tinemaha Reservoir (R,D)	A
Head-dependent flux	Evapotranspiration package	Evapotranspiration (D)	A
Head-dependent flux	Drain package	Springs and seeps (D)	A

years 1963–88. Recharge from irrigation and watering of livestock was simulated as having a constant rate for each of two periods, water years 1963–69 and 1970–88. All other components were simulated as having different annual values. Any major changes that were made to initial estimates of recharge and discharge components simulated by the well package are described below.

Some canals, ditches, and ponds probably gain water from the aquifer system, at times, instead of acting as recharge components (table 13). To attempt to account for this dual character, a head-dependent relation (in particular, the river package described below) was used to approximate some of the larger canals during development of the detailed ground-water flow model of the Bishop area (Hutchison, 1988). This technique, however, was found to dampen fluctuations in ground-water levels too severely, and it was abandoned.

Estimates of recharge from ponds were not changed, except for an initial estimate of a 90-percent percolation rate for purposeful ground-water recharge in the Laws area. This rate produced poor model results, and it was reduced during calibration to 75 percent.

Pumpage for each well was assigned to individual model cells using the map-projection and translation programs described in the previous "General Characteristics" section of this report and the well-location information given in table 9. Distribution of average measured pumpage from both model layers is shown on plate 3*F*.

Underflow was approximated, at first, using Darcy's law. The calculated quantities of underflow were distributed along the flow boundary on the basis of estimated transmissivities. These initial estimates of underflow had a high degree of uncertainty associated with them, and they did not work well in the model; subsequently, they were reduced significantly during calibration (pl. 3*G*).

River package.—Permanent surface-water bodies exchange water with the aquifer system—gaining water if nearby ground-water levels are higher than the surface-water stage, and loosing water if nearby levels are lower. A head-dependent relation, referred to as "the river package" by McDonald and Harbaugh (1988, p. 6–1), permits simulation of this type of interaction. The quantity of water exchanged is calculated by the model from the average stage of the stream, altitude of the bottom of the streambed, transmissive properties of the streambed, and model-calculated head for the upper model layer.

In order to simulate different surface-water features (table 13), the average stage and altitude of the bottom of the streambed (or equivalent riverbed or lakebed) were estimated for each model cell from values of land-surface datum obtained from 1:62,500-scale USGS topographic maps. For the Owens River, the Los Angeles Aqueduct, and the lower Owens River, the slope of the river stage from upstream to downstream model cells was checked to ensure that the slope was relatively smooth and uniformly downhill. The concrete-lined, nearly impermeable section of the Los Angeles Aqueduct near the Alabama Hills was not included in the model.

A "conductance" term is used in the river package to incorporate both the transmissive properties of the streambed and the wetted area of the surface-water feature. The transmissive properties of the streambed (bottom sediment) for each feature were estimated from typical values for valley-fill deposits (table 9; Hollett and others, 1991, table 1) and later were modified during calibration. For example, values of conductance for the lower Owens River were decreased somewhat from values for the Owens River in the Bishop Basin because deposits near the river in the Owens Lake Basin are characteristically finer and less transmissive. The wetted area of each feature was estimated from topographic maps, photographs, and field reconnaissance.

The Pleasant Valley Reservoir was not simulated explicitly in the model, although recharge from the reservoir was considered in selecting values of underflow and in evaluating the simulated gain of water by the Owens River immediately downstream from the reservoir. Use of the river package to simulate sewage ponds near the four major towns was physically realistic, but the parameters and results are highly uncertain.

Evapotranspiration package.—Evapotranspiration was calculated in the model from a piecewise-linear relation, a series of connected straight-line segments, that is based on depth of the water table below land surface (McDonald and Harbaugh, 1988, p. 10–3). An assumption was made that evapotranspiration ceases when the water table is more than 15 ft below land surface (Groeneveld and others, 1986a; Sorenson and others, 1991). When the water table is at land surface, a maximum evapotranspiration rate is reached. At intermediate depths, the evapotranspiration rate linearly decreases from the maximum rate to zero.

The average maximum evapotranspiration rate for vegetation on the valley floor was estimated to be 24 in/yr for the period prior to 1978. This estimate is based on measured evapotranspiration (table 5), results from previous modeling (Danskin, 1988), and measurements of transpiration by Groeneveld and others (1986a, p. 120). The dramatic increase in average pumping after 1970 and the drought of 1976–77 were assumed to permanently decrease the maximum vegetative cover on the valley floor. As a result, the maximum evapotranspiration rate was reduced by 25 percent from 24 in/yr to 18 in/yr for the period after 1977. This reduction was based on the reduced quantity of water available for evapotranspiration (table 10), on the variability of maximum evapotranspiration rates (table 5), and on the observed response to decreased water availability (Sorenson and others, 1991).

The maximum evapotranspiration rates used in the ground-water flow model (28 or 24 in/yr) were chosen to represent the broad areas of native vegetation covering most of the valley floor. These rates tend to underestimate evapotranspiration from riparian vegetation, for which evapotranspiration exceeds 40 to 60 in/yr (D.P. Groeneveld, Inyo County Water Department, written commun., 1984; Duell, 1990). In particular, along the lower Owens River, evapotranspiration is influenced greatly by an abundance of high-water-use cattails (fig. 10*C*). As a result, evapotranspiration calculated by the model underestimates the actual evapotranspiration near the lower Owens River, possibly by as much as 2,000 acre-ft/yr. Most of this extra discharge, however, is simulated by the river package as a gain to the lower Owens River. The net effect on the aquifer system is the same although the accounting is different. This artifact of the model is recognized as potentially confusing, but it does not alter any of the basic conclusions presented in this report.

Drain package.—Springs and seeps were simulated with the head-dependent relation referred to as “the drain package” by McDonald and Harbaugh (1988, p. 9–1). This relation uses a value of the transmissive properties (conductance) of the spring and the simulated model head to compute a discharge—if the model head is higher than a specified drain altitude. If the model head is lower, discharge is zero. The drain altitudes were chosen on the basis of a leveling survey of each spring (R.H. Rawson, Los Angeles Department of Water and Power, written commun., 1988), or on a value of land surface obtained from 1:62,500-scale USGS topographic maps.

Simulation Periods

Simulation periods were chosen to calibrate and verify the ground-water flow model, to evaluate past water-management practices, and to predict the likely condition of the aquifer system after 1988. Historical periods of similar water use, as summarized in table 4, were used as an aid in selecting simulation periods that capture the main elements of water management in the Owens Valley and rigorously test the model.

Water year 1963 was chosen to calibrate the ground-water flow model under equilibrium or steady-state conditions. This particular period was chosen for three reasons. First, ground-water levels did not seem to change significantly during water year 1963, a prerequisite for a steady-state analysis. Second, the percent of valleywide runoff for water year 1963 was about average (107 percent of normal). Third, although water year 1963 was preceded by a short-term increase in ground-water pumpage, the year was sufficiently isolated from major runoff or pumping effects that the aquifer system was assumed to be in a quasi-steady-state condition—that is, sufficiently stable to begin a transient simulation.

Water years 1963–84 were chosen to calibrate the ground-water flow model under nonequilibrium or transient conditions. Stable initial conditions were ensured by beginning the transient simulation with results from the steady-state simulation of water year 1963. The first part of this period, water years 1963–69, represents conditions in the valley prior to completion of the second aqueduct (table 4). The Los Angeles Department of Water and Power (1972) showed that the valleywide system was in approximate equilibrium for water years 1935–69 and, except for brief periods of heavy pumping during the 1930's and early 1960's, probably in near-equilibrium for most of the period between the completion of the first aqueduct in 1913 and the second in 1970. Therefore, the first part of the calibration period, water years 1963–69, was assumed to be fairly analogous to the entire period prior to operation of the second aqueduct.

The second part of the calibration period, water years 1970–84, represents the significantly different conditions in the valley after completion of the second aqueduct and the related changes in water use (table 4). This second period was a time of significantly increased pumpage, a decrease in water supplied for

agricultural and ranching operations, a severe drought (1976–77), and extremely wet conditions following the drought. The ability of the model to simulate such diversity of conditions within the same calibration period reflects on its appropriate design and helps to confirm that the model is a fairly complete representation of the actual aquifer system.

Water years 1985–88 were chosen to verify that the ground-water flow model was not uniquely tuned to the calibration period and could be used to evaluate non-calibration periods. The verification period, although short, is a good test of the calibrated ground-water flow model because there are significant fluctuations in runoff and pumpage. Also, new high-production "enhancement and mitigation" wells were put into service. The verification period was simulated after calibration of the model was complete. Recharge and discharge components required for the verification period were calculated in the same way as for the calibration period. No changes were made to recharge, discharge, or other parameters in the ground-water flow model. In fact, as it turned out, all model simulations for the verification period were completed prior to obtaining and reviewing measured ground-water-level data for the period—a rather unnerving, if somewhat fortuitous sequence for verification.

A final simulation period was defined to represent "1988 steady-state conditions"—that is, the equilibrium that the aquifer system would reach if operations as of 1988 were continued well into the future. Preliminary evaluation at the beginning of the cooperative studies identified water year 1984 as a likely period that could be used to simulate average present conditions. Subsequent analysis, however, determined that the Owens Valley was in the midst of significant vegetation and hydrologic changes and that stable quasi-steady-state conditions did not exist in 1984. Therefore, a more generalized steady-state simulation was designed, taking into account long-term average runoff and new enhancement and mitigation wells that were installed after 1984. This simulation and the related assumptions and approximations are described later in this report in a section entitled "Alternative 1: Continue 1988 Operations."

Calibration

Calibration of the ground-water flow model involved a trial-and-error adjustment of model parameters representing aquifer characteristics and certain recharge and discharge components in order to obtain an acceptable match between measured ground-water levels and computed heads and between estimated and computed recharge and discharge. For example, more than 200 hydrographs displaying levels and heads were reviewed throughout the calibration process; 67 of these hydrographs for 56 model cells are shown on plate 1. Also, simulated recharge and discharge were reviewed extensively on a "cell-by-cell" basis (McDonald and Harbaugh, 1988, p. 4–15) to ensure that the magnitude and distribution of computed ground-water flows (fluxes) were appropriate. The calibration process was continued until further changes in the ground-water flow model did not significantly improve the results and until the model parameters, inflows and outflows, and heads were within the uncertainty of historical data.

The philosophy of model development and calibration was to use general relations for as many components of the model as possible. These relations, or conceptual themes, permit an improved understanding of the overall model and its more than 100,000 parameters. For example, the hydraulic characteristics of the model were based on hydrogeologic subunits (model zones), each with uniform hydraulic properties. Reductions in transmissivity caused by faults were calculated as a percentage of the transmissivity of the faulted material (pl. 2). Recharge and discharge commonly were related to a more general concept, such as the percent of average valleywide runoff. Detailed, site-specific adjustment of parameters or relations was done rarely, if at all. Because of the way it was calibrated, the model is most useful for evaluating valleywide conditions, not for predicting small-scale effects covering a few model cells. Site-specific ground-water flow models or multivariate regression models, such as developed by P.B. Williams (1978) and Hutchison (1991), can give more accurate predictions at selected sites. However, these models in turn are less useful for evaluating valleywide hydrogeologic concepts or predicting valleywide results of water-management decisions.

The calibration procedure first involved estimating initial values of inflow and outflow to the aquifer system for the steady-state period, water year 1963. Many of the estimates were obtained from preliminary work by Danskin (1988). Adjustments were made in some of the initial estimates in order to ensure a balance of inflow and outflow as well as to match the distribution of measured ground-water levels. An assumption in the calibration of steady-state conditions was that ground-water levels in 1963 were similar to

those in 1984 for most parts of the basin (fig. 14). This assumption was necessary because of the absence of virtually any ground-water-level data prior to 1974 for hydrogeologic unit 1.

The bulk of the calibration involved making adjustments to the model that are based on the transient behavior of the aquifer system during the 22-year period, water years 1963–84. To ensure stable initial conditions, the steady-state period was resimulated each time changes were made to the model. Also, the distribution of head and the pattern of ground-water flow were reevaluated for each steady-state simulation to ensure that they remained conceptually valid and similar to those shown in figure 14.

Transmissivity values were adjusted within the general range indicated by aquifer tests (fig. 15 and table 9) and related studies (Hollett and others, 1991; Berenbrock and Martin, 1991). Calibrated values of transmissivity were slightly higher than initial estimates for highly transmissive volcanic deposits, especially in the area of Crater Mountain near Fish Springs (fig. 15 and pl. 2).

Values of vertical conductance were constrained to approximately the same values derived from the preliminary models (fig. 2) and from aquifer tests described by Hollett and others (1991). Values were adjusted until simulated heads in the upper and lower model layers matched measured ground-water levels indicated on contour maps (fig. 14) and on hydrographs (pl. 1). For most of the area covered by alluvial fan deposits, measured levels were not available. In these areas, values of vertical conductance were adjusted so that simulated heads in the two layers differed by less than 1 ft.

Storage coefficients were held constant at 0.1 and 0.001 for the upper and lower model layers, respectively. For the upper model layer, the storage coefficient is virtually equivalent to specific yield. Values determined from aquifer tests (table 9), as expected, were lower than model values. Aquifer tests, even those extending several days, are affected most by the compressive response of the aquifer and expansion of ground water and are affected very little by actual drainage of the aquifer materials. This drainage, which accounts for nearly all of the specific-yield value, is delayed and occurs slowly over a period of weeks, months, or years. As a result, storage coefficients obtained from model calibration of long-term conditions usually are much more indicative of actual values than are those calculated from aquifer tests. Attempts at specifying unique storage coefficients for each hydrogeologic unit proved to be tediously unproductive.

All recharge and discharge components had conceptual or semi-quantitative bounds associated with them. These bounds (which are discussed in greater detail in other sections of this report, including "Surface-Water System" and "Ground-Water Budget") restricted model calibration in much the same way as did measured ground-water levels (pl. 1). Some recharge and discharge components (recharge from precipitation, recharge from spillgates, and underflow) were assigned constant rates on the basis of their uniform characteristics from one year to another (tables 11 and 13). All other components were varied annually on the basis of a general concept such as percent annual runoff.

Most recharge and discharge components did require some degree of adjustment, often minor, during calibration. This adjustment was needed not only to match measured conditions, but also to ensure that a consistency between different recharge and discharge components was maintained. For example, changing recharge from a narrow canal on the valley floor required re-evaluating the quantity of recharge from narrow tributary streams on alluvial fans and from broad river channels on the valley floor. The philosophy of calibration did not permit adjusting values in individual model cells in order to match historical conditions.

The location and type of model boundaries were assumed to be known and were not varied. The quantity of underflow, however, was reduced considerably from previous estimates by Danskin (1988) and the Los Angeles Department of Water and Power (1976). Recharge from canals was slightly less than original estimates. Recharge from purposeful water-spreading operations was about two-thirds of the initial estimate. Conductance of both the river–aqueduct and the lower Owens River were increased during calibration, thereby increasing ground-water recharge to or discharge from them. The quantity of evapotranspiration was less than original estimates. Pumpage was assumed to be known and was not changed.

Land-surface datum was used in many parts of the model, particularly in defining head-dependent relations and estimating precipitation (fig. 7*B*). Attempts at computing land-surface values from 1:250,000-scale AMS (Army Mapping Service) point data sets obtained from R.J. Blakely (U.S. Geological Survey, written commun., 1986) required fitting a

surface to the point data; results were not satisfactory, especially in areas of abrupt change in slope of the land surface, such as near the Tinemaha Reservoir. Therefore, the values were interpolated by hand from 1:62,500-scale USGS topographic maps and held constant during calibration.

Results of the model calibration are displayed in figures 19 and 20, which show comparisons of measured ground-water levels and simulated heads during spring 1984 for the upper and lower model layers, respectively. This was a time when levels were higher than they had been for several years, dormant springs had resumed some discharge, and the basin was assumed to be in a nearly full condition (Hollett and others, 1991). The match between measured levels and simulated heads for both the upper and the lower model layers seems to be quite good for most parts of the basin. A notable exception is the area west of Bishop near the Tungsten Hills.

Measured water levels and simulated heads for individual wells are compared on plate 1. Although more than 200 wells were used extensively in the calibration process, only 67 wells are included on plate 1. The 67 wells were selected to represent different well fields, different model layers, and different hydrogeologic subunits (model zones). Some wells were included on plate 1 to illustrate those parts of the valley where the ability of the model to simulate actual conditions is not as good as in other locations—for example, well 278 near Bishop and well 172 near Lone Pine (pl. 1).

Precise tracking of the measured and simulated hydrographs (pl. 1) was not deemed necessary, and might not be desirable or correct depending on the characteristics of the well, the surrounding aquifer material, and the model cell approximating the well. Of primary importance was that the measured and simulated hydrographs be of the same general shape and trend. Shape of a hydrograph is influenced by aquifer characteristics, recharge, and discharge; trend is influenced most by change in aquifer storage. The magnitude of vertical deflection likely will be different for measured and simulated hydrographs because of spatial discretization required for the model. The ratio of vertical deflections between the two hydrographs, however, should remain similar over time. Vertical offsets might or might not be important depending on the specific well. For example, an acceptable vertical offset can result when a well is located away from the center of a model cell; this type of offset is particularly noticeable in areas of steep hydraulic gradients, such as on the alluvial fans.

During calibration of the valleywide model, the comparison between estimated and simulated recharge and discharge was as important as the comparison between measured ground-water levels and simulated heads. Recharge and discharge components that act as hydraulic buffers respond to changes in other model parameters and reflect the dynamics of the aquifer system—sometimes much better than do changes in head. The simulated recharge and discharge for the dominant fluxes in the model after calibration are shown in figure 21.

As an aid in using and extending the work presented in this report, simulated values for each component of recharge and discharge in the ground-water flow model are given in table 11. The individual values are important aids in compiling water budgets for specific parts of the valley; developing linked water budgets for the surface-water and ground-water systems; defining the relative degree of confidence to be placed in model results in different parts of the valley; identifying how to revise and improve the model; and making local water-management decisions.

In places where concepts or data were uncertain, the ground-water flow model was not calibrated forcibly to produce a match between simulated heads and measured levels. For example, in the area north of Laws, something is missing in the ground-water flow model. Simulated heads in layer 1 do not recover after 1974 as fully as do the measured levels (well 107T, pl. 1). The actual recovery could be caused by any of several processes—increased underflow during the drawdown period, induced flow of water from Fish Slough or the Bishop Tuff, increased percolation of operational spreading of surface water, or changes in the operation of nearby canals. Without a valid reason to pick one process rather than another, none was altered during calibration—thus highlighting an area of uncertainty and an area where further work is necessary. This approach was a major philosophy of the modeling study and the rationale for including some of the hydrographs shown on plate 1.

Verification

Water years 1985–88 were used to verify that the calibrated ground-water flow model will duplicate measured data for a non-calibration period. The 4-year verification period included significant stress on the aquifer system because of unusually wet and dry

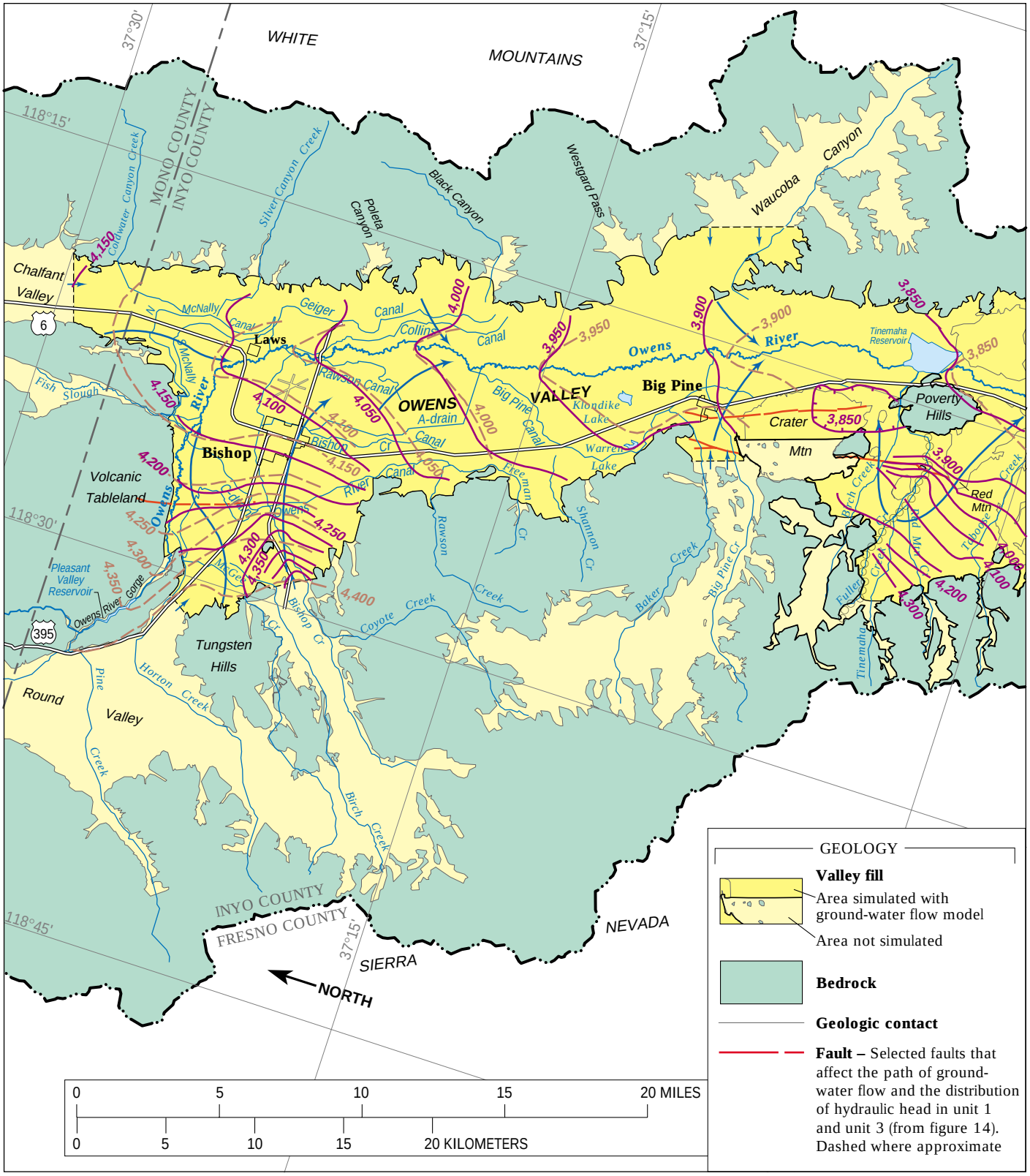
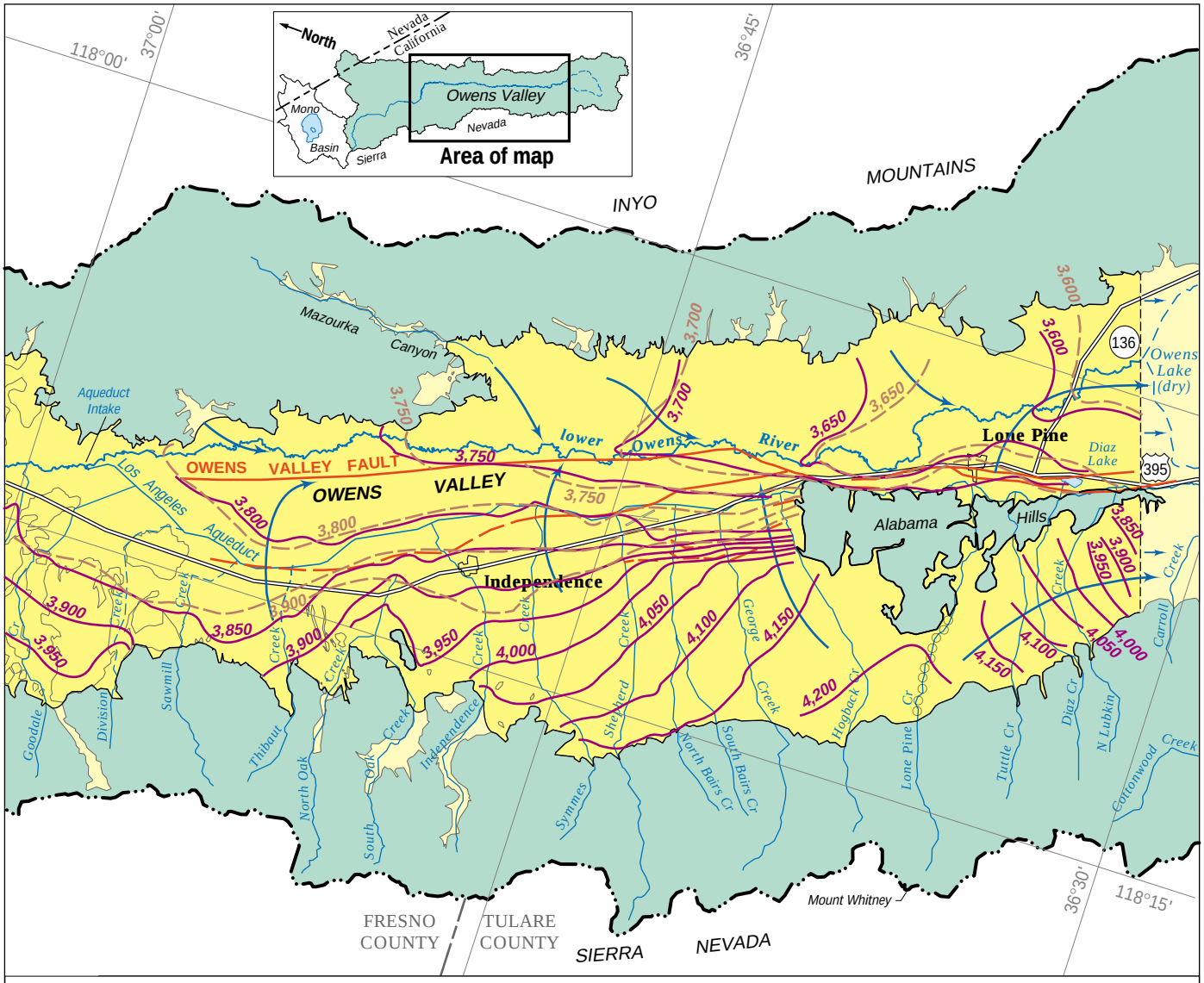


Figure 19. Measured and simulated potentiometric surfaces for hydrogeologic unit 1 (upper model layer) in the Owens Valley, California, spring 1984.



EXPLANATION

HYDROLOGY

Measured — 3,900 — **Potentiometric contour –** Shows approximate altitude of the water table in hydrogeologic unit 1, represented by the upper layer of the ground-water flow model, spring 1984. Contour interval 50 feet. Datum is sea level

Simulated — 3,900 —

Generalized direction of ground-water flow – Combined direction of ground-water flow in hydrogeologic units 1 and 3

Ground-water divide – Approximately located

Boundary of the aquifer system – As defined in this report. Arrows indicate the direction of ground-water flow to or from adjacent permeable materials

Boundary of the Owens Valley drainage basin

Figure 19. Continued.

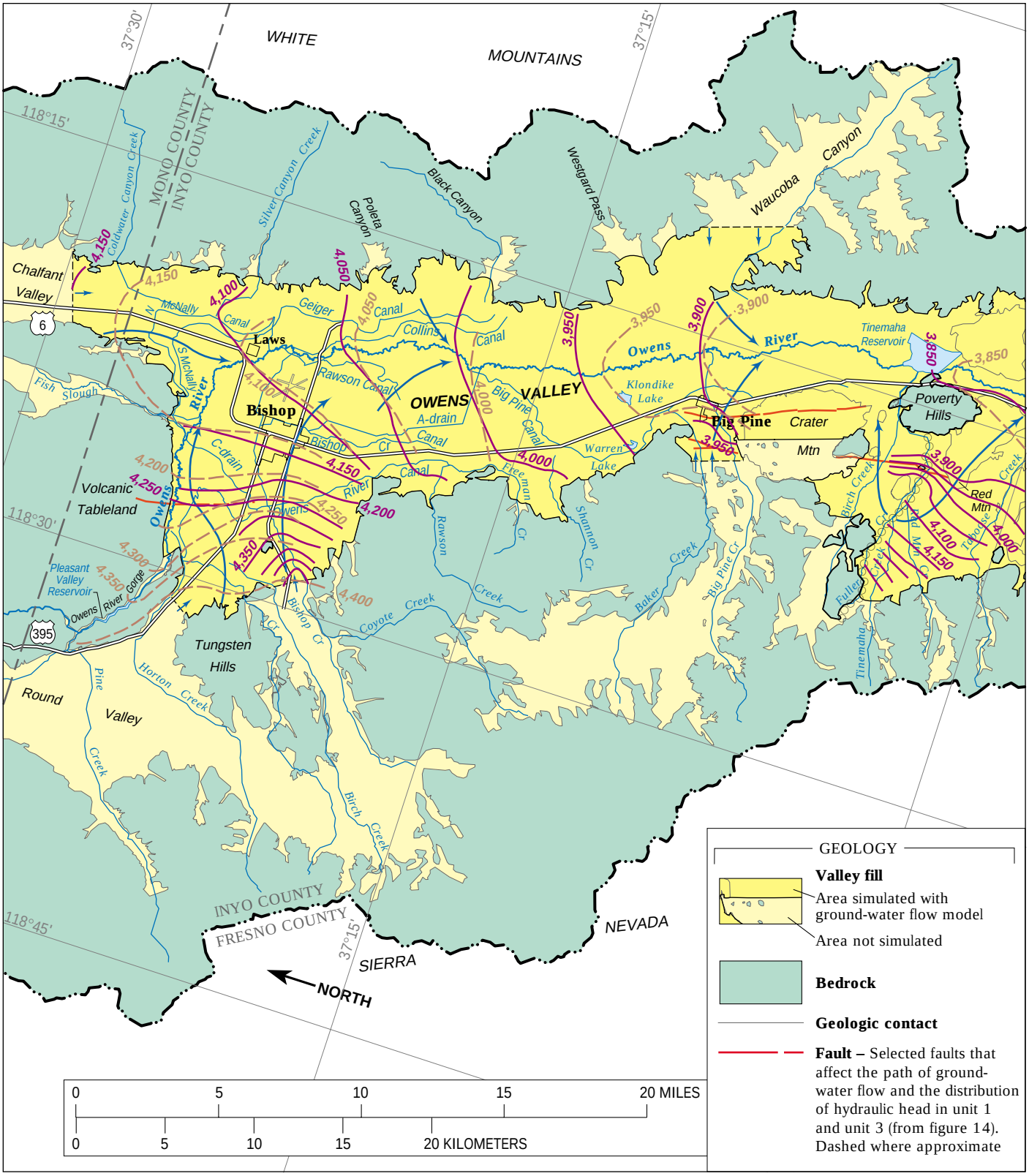
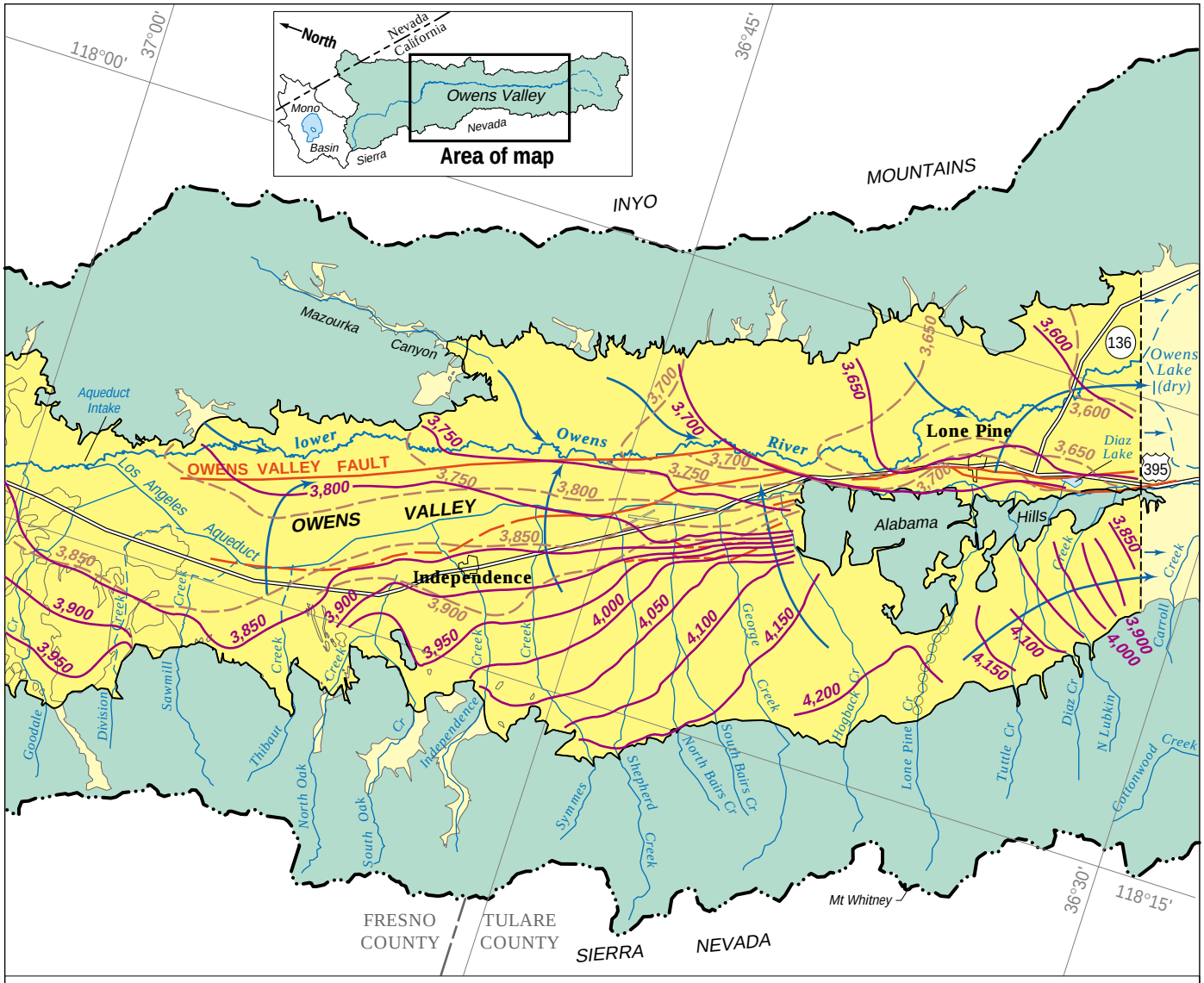


Figure 20. Measured and simulated potentiometric surfaces for hydrogeologic unit 3 (lower model layer) in the Owens Valley, California, spring 1984.



EXPLANATION

HYDROLOGY

—3,900— Measured

—3,900— Simulated

Potentiometric contour – Shows approximate altitude of the hydraulic head in hydrogeologic unit 3, represented by the lower layer of the ground-water flow model, spring 1984. Contour interval 50 feet. Datum is sea level

Generalized direction of ground-water flow – Combined direction of ground-water flow in hydrogeologic units 1 and 3

Ground-water divide – Approximately located

Boundary of the aquifer system – As defined in this report. Arrows indicate the direction of ground-water flow to or from adjacent permeable materials

Boundary of the Owens Valley drainage basin

Figure 20. Continued.

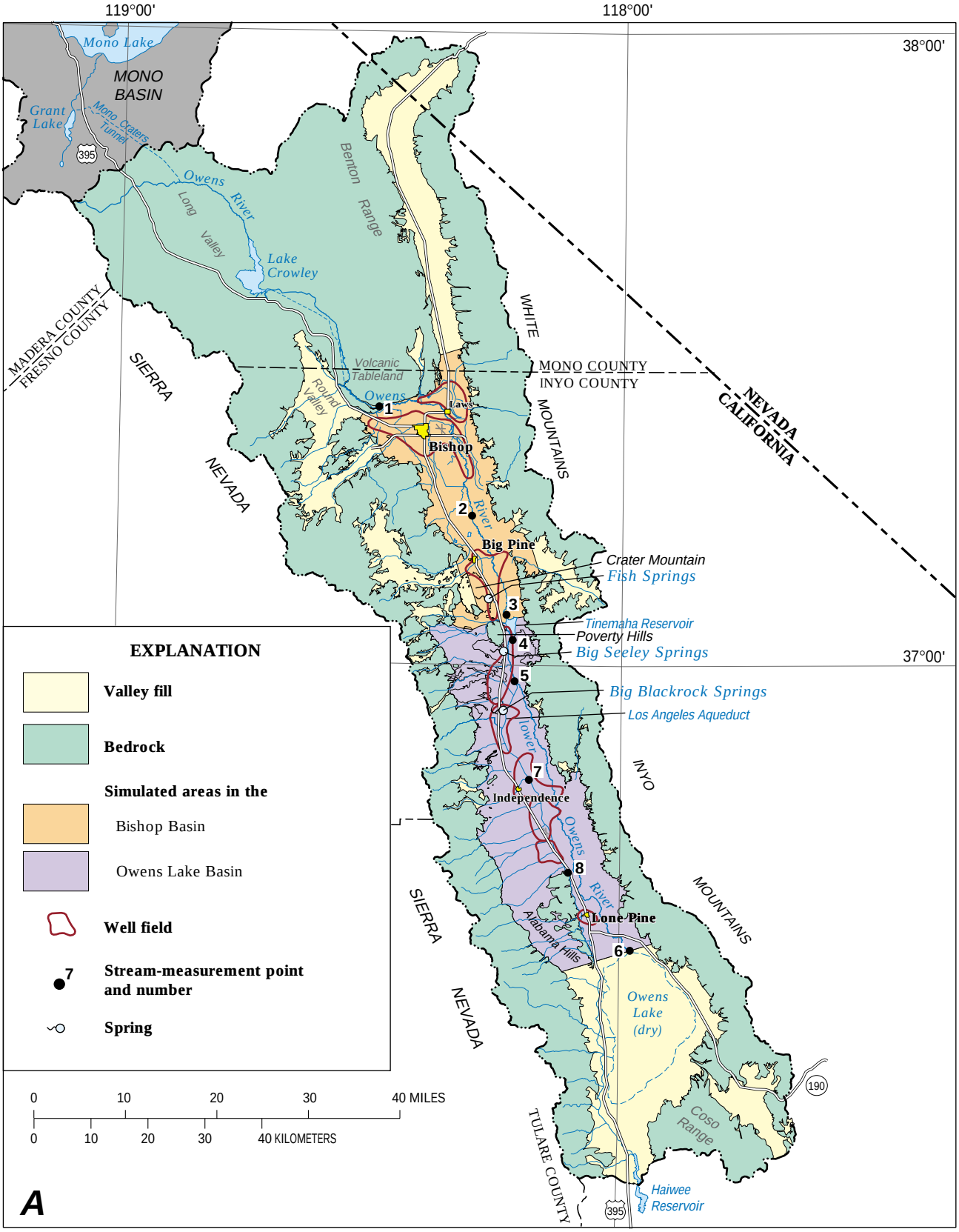
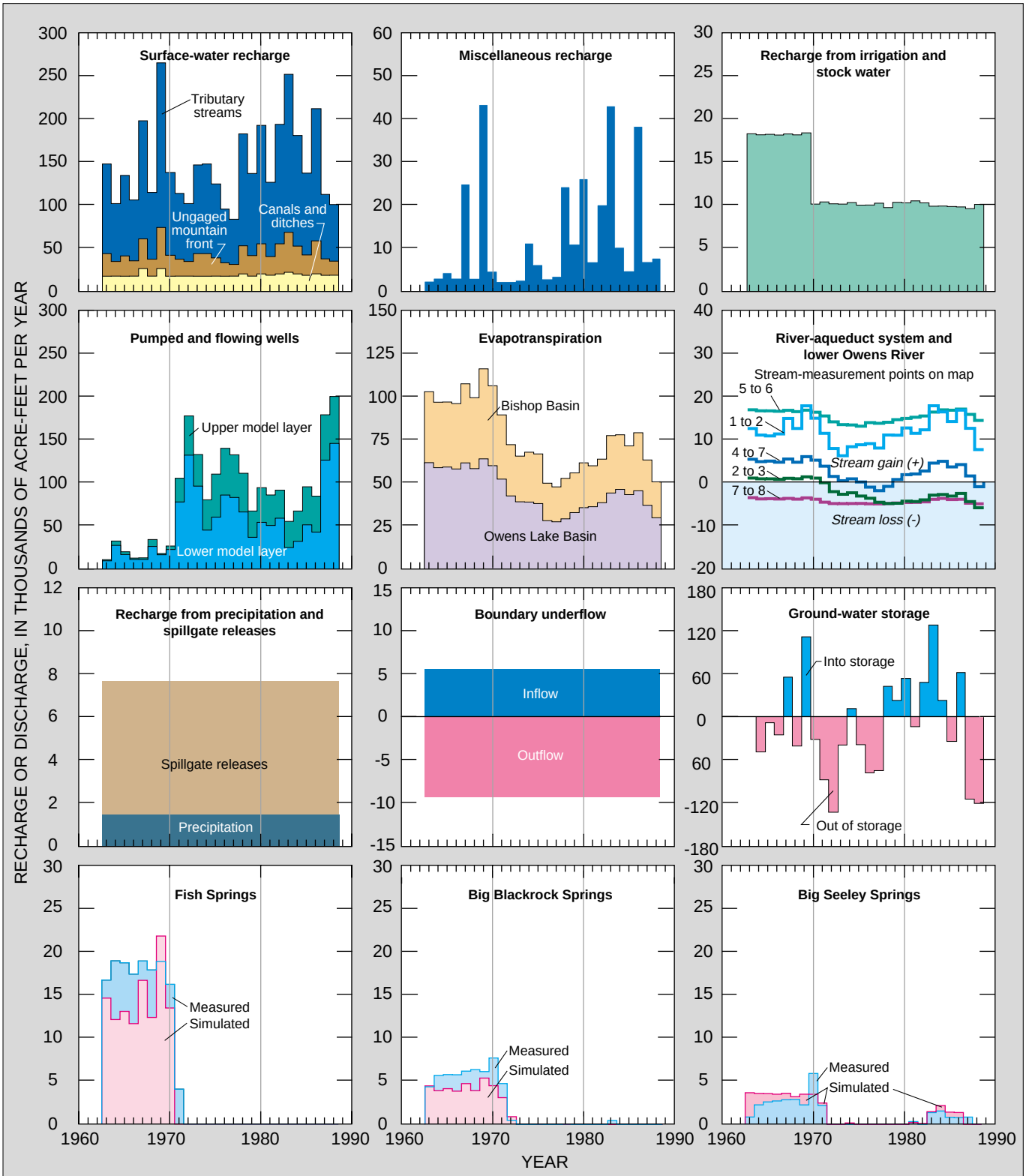


Figure 21. Simulated ground-water recharge and discharge during water years 1963–88 in the Owens Valley, California. Values for each water-budget component are given in table 11.



Multiple items in bar graphs are shown as added (stacked) values, except for stream gain, stream loss, and spring discharge. Miscellaneous recharge includes "operations, ground-water recharge, recreation/wildlife, enhancement and mitigation" as defined by the Los Angeles Department of Water and Power (R.H. Rawson, written commun., 1987-89). Refer to table 11

B

Figure 21. Continued.

conditions. Valleywide runoff varied from 158 to 68 percent of normal (table 7). In addition, new enhancement and mitigation wells were put into production in various locations throughout the valley (tables 9 and 11). Initial conditions for the verification were simulated heads for water year 1984 at the end of the calibration period. Recharge and discharge data were developed for the ground-water flow model in exactly the same way and using the same relations as had been done for the calibration.

A comparison of measured ground-water levels and simulated heads during the verification period is shown on plate 1. In general, the match is very good, particularly in the Laws area where the aquifer was highly stressed. The model also simulates the return of spring discharge during the period (fig. 21). The close agreement between measured ground-water levels and simulated heads and between measured and simulated spring-discharge rates was achieved without any adjustment of model parameters. This ability to reasonably match data from another time period suggests that the ground-water flow model can be used to predict results from stresses that are similar in type and magnitude, but not exactly the same as those used during calibration—a prerequisite for a predictive model.

Sensitivity Analysis

Sensitivity analysis is a procedure to determine how sensitive the model solution is to a change in each model parameter, including transmissivity, vertical conductance, storage coefficients, and inflow and outflow rates. As is always the case with numerical models, not all parameters of the model were known completely. Because some uncertainty is present in each parameter, there is some uncertainty in the model solution. This uncertainty is reflected in heads and inflow and outflow rates that are somewhat in error. A sensitivity analysis identifies which parameters exert the most control over the model solution and, therefore, have the potential to generate the largest errors. An improved understanding of those parts of the aquifer system represented by the most sensitive parameters yields the greatest improvement in the ground-water flow model.

One of the sensitivity tests that was most illuminating is presented in figure 22. For the test, water years 1963–88 were resimulated with slight modifications in recharge and discharge. For the first part of the test (fig. 22*A*), recharge from tributary streams, recharge from ungaged areas between tributary streams, and recharge from runoff from bedrock outcrops within the valley fill were held constant at 100 percent of long-term average conditions (100-percent runoff year). In the second part of the test (fig. 22*B*), calibration values were used for everything except ground-water pumpage, which was held constant at the values for water year 1963. Effects from each test were observed at wells in recharge areas, near well fields, and away from both recharge areas and well fields. As expected, the effects in recharge areas are most dependent on recharge, and the effects near well fields are most dependent on pumpage. Away from either area, heads are relatively unaffected by changes in either recharge or pumpage, probably as a result of the many hydraulic buffers in the aquifer system. What is somewhat surprising is the degree to which both recharge areas and well fields are affected by pumpage. Clearly, pumpage plays the dominant role in affecting heads (ground-water levels) in the valley.

For the rest of the sensitivity analysis, each of the model parameters was altered by a certain amount from the calibrated values. The amount of the alteration was determined by estimates of the likely range of the data (Hollett and others, 1991, table 1) (figs. 15 and 16; tables 9, 10, and 11). To simplify the analysis, similar variables, such as transmissivity on the alluvial fans, were altered together. The variables associated with the most change in the model solution were identified as the most sensitive. Similar sensitivity analyses were done using a ground-water flow model of the Bishop Basin (Radell, 1989) and a model of the Owens Lake Basin (Yen, 1985). Those analyses are presented graphically for several of the model parameters and depict results similar to those discussed here for the valleywide model.

Although useful, this method of testing sensitivity is subject to a potentially significant flaw. Because each variable in the model is tested separately, the additive effects of changes in more than one variable are not considered. For example, the simultaneous overestimation of both recharge and evapotranspiration in the model would tend to be self-correcting. However, overestimating recharge and underestimating evapotranspiration would produce a considerably different model solution. If neither recharge nor evapotranspiration by itself were a sensitive part of the model, the conclusion from a routine sensitivity analysis would be that additional refinement of these

rates is unnecessary. Nevertheless, the additive effects of errors in recharge and evapotranspiration might produce significantly erroneous results in some simulations of the aquifer system.

This type of error can be prevented by means of a more subjective analysis of sensitivity during development and calibration of the ground-water flow model. The modeling technique chosen for the valleywide model took advantage of this method. Those characteristics of the aquifer system believed to be most important were analyzed first using different-scale models (fig. 2). Then, the valleywide model was developed by adding sequentially greater complexity to the model—one recharge or discharge component, or one additional model zone at a time. In this way, during model development and calibration, the sensitivity of each model parameter could be identified more easily. These observations, which are as valuable as a post-calibration sensitivity analysis, also are included in the following discussion of the sensitivity of each parameter.

Transmissivity.—The areal distribution of transmissivity in the valley is based on scattered data (fig. 15) and an assumption of uniformity within each model zone (pl. 2). Model errors can be associated with the values of transmissivity chosen for an individual zone and with the choice of zone boundaries. The sensitivity of the model to the locations of the zone boundaries is best evaluated by altering the locations, recalibrating the model, and observing the differences. Although this time-consuming process was not part of this investigation, the location of the transition zone was found, during model development, to be a sensitive parameter. Equally sensitive was the location and, in particular, the continuity of volcanic deposits near the Taboose–Aberdeen and the Thibaut–Sawmill well fields (fig. 17).

Variations in the value of transmissivity within a model zone produced less effect on heads and ground-water discharge than was hypothesized initially. An exception to this was the area of highly transmissive volcanic materials between Big Pine and Fish Springs (pl. 2). Lower values of transmissivity produced much lower discharge from Fish Springs and unrealistically steep gradients from north to south along the edge of Crater Mountain. From a valleywide perspective, the addition of the more transmissive model zones representing transition-zone and volcanic deposits produced a much greater effect on heads than did variations of transmissivity within individual zones.

Vertical conductance.—Calibrated values of vertical conductance (the model equivalent of vertical hydraulic conductivity) were based on sparse field data and model calibration. To test a wide range of possible values, vertical conductance in each hydrogeologic area was varied by two orders of magnitude. However, the effect on heads was not as pronounced as was expected. In fact, the model seemed to be rather insensitive to changes in vertical conductance (Radell, 1989, fig. 6.4). Part of the reason for this may be the relatively large size of the model cells and use of an annual approximation of recharge and discharge. Both of these model characteristics, which require averaging simulated recharge and discharge over space or time, result in less change in simulated ground-water levels for a given recharge or discharge than would occur in the actual aquifer system. A greater sensitivity in vertical conductance might be expected in an analysis using smaller distances and shorter timeframes, similar to those used to analyze an aquifer test. During calibration, the value of vertical conductance was noted as being closely tied to the rate of evapotranspiration, which tends to dampen changes in heads near the valley floor. Lower values of vertical conductance result in less flow from the lower model layer to the upper, which in turn results in less water available for evapotranspiration. This spatial correlation between vertical conductance and evapotranspiration can be seen by comparing the vertical difference in head (figs. 19 and 20) with evapotranspiration rates (pl. 3*A*)

Storage coefficient.—Storage coefficient was determined to be one of the least sensitive variables. This result corresponds to similar findings by Yen (1985, p.150). Sensitivity analysis showed that storage coefficients higher than the calibrated values did not change heads significantly, but values less than about 0.0001 for the lower model layer (hydrogeologic unit 3) produced unrealistic variations in heads at many locations in the basin.

Precipitation.—Precipitation records for the Owens Valley, in general, are very good, except for an absence of precipitation stations on the east side of the valley (fig. 7*A*). Nearly all precipitation falling on the valley floor is assumed to be used by native vegetation, and recent monitoring of the unsaturated zone tends to confirm this assumption (Groeneveld and others, 1986a; Sorenson and others, 1991). Therefore, the effect of recharge from precipitation falling on the valley floor was not tested in the sensitivity analysis.

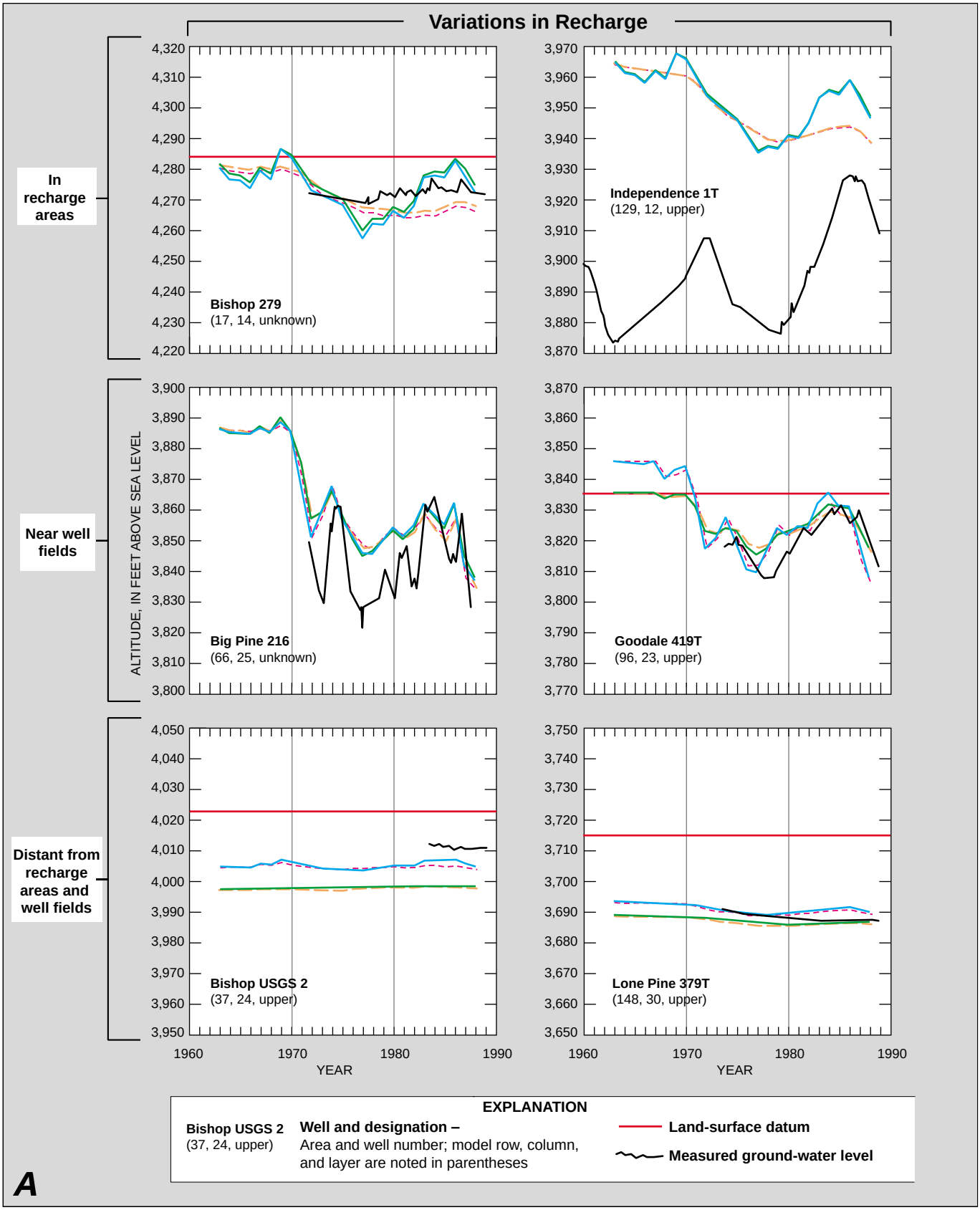


Figure 22. Sensitivity of simulated hydraulic heads in the Owens Valley, California, to variations in recharge ***(A)*** and pumpage ***(B)*** at wells in recharge areas, near well fields, and distant from both. Method of variation is described in text. Well locations are shown on plate 1.

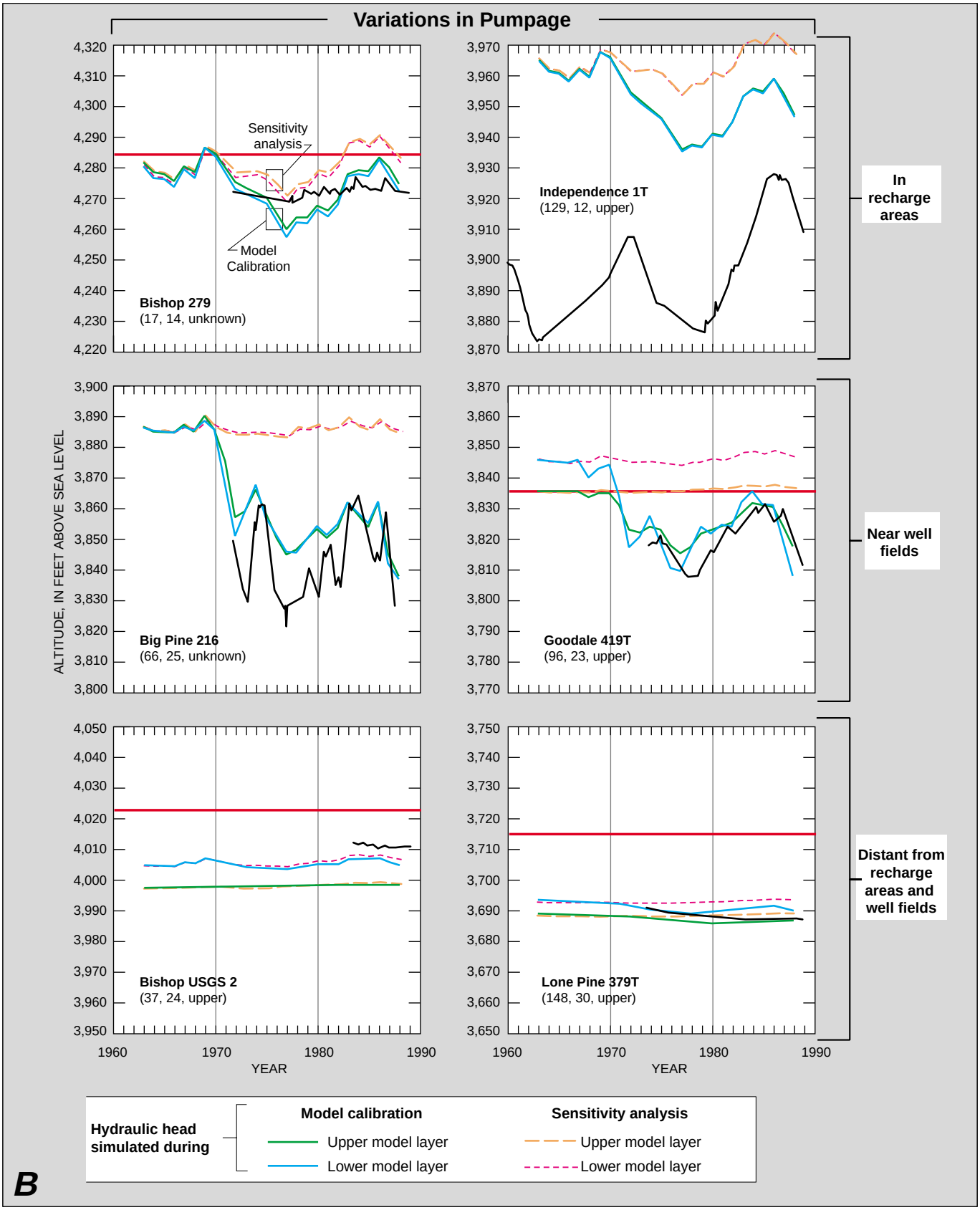


Figure 22. Continued.

In contrast, recharge from precipitation is assumed to occur along the mountain fronts, but the quantity is completely unknown. The present assumption is that about 95 percent of precipitation is evapotranspired, and 5 percent, or about 2,000 acre-ft/yr, is recharged (table 10). Variations of 3 to 4 times this value produced minor effects on model simulations, primarily increasing evapotranspiration from the valley floor and gains of water by the river–aqueduct system. Similar results were found by Radell (1989, fig. 6.10). If the present assumption is largely incorrect, then recharge from precipitation could be a sensitive model parameter with respect to ground-water flow rates as found by Danskin (1988). However, a large increase in recharge from precipitation probably would require a similar decrease in mountain-front recharge between tributary streams (tables 10 and 11) in order to maintain a calibrated model.

Tributary stream recharge.—Measurements of tributary stream discharge are among the most complete and most accurate hydrologic measurements in the valley. Because most tributary streams are measured at both a base-of-mountains gage and a river–aqueduct gage (fig. 11), estimates of tributary stream recharge do not vary greatly. An increase of 10 to 20 percent in tributary stream recharge for streams in the Owens Lake Basin resulted in moderate to significant changes—generally, higher heads on the fans and a greater gain of water by the river–aqueduct system. Heads and evapotranspiration rates on the valley floor showed much less effect. In the Bishop Basin, particularly near Big Pine, accounting for each stream is more difficult, and the uncertainty in recharge estimates is greater than in the Owens Lake Basin. Variations of as much as 50 percent in tributary stream recharge near Big Pine and Taboose Creeks resulted in a minimal change in heads in this area of high transmissivities, but an important change in the discharge of nearby springs (fig. 17).

Mountain-front recharge.—Mountain-front recharge between tributary streams is a large, poorly quantified component of the ground-water budget (table 10). Sensitivity analysis of this item included variations of a 50-percent increase or decrease and resulted in significantly different heads and ground-water fluxes along the west side of the basin. Results are similar to a 15-percent error in recharge from all tributary streams. The lack of measured data suggests that errors in estimating mountain-front recharge are more likely than for most other components of the ground-water flow model. This large degree of uncertainty makes the high sensitivity of this component even more important. During calibration of the Bishop area, an inverse correlation was observed between the quantity of mountain-front recharge and the quantity of recharge from canals and ditches; an increase in recharge for one component probably requires a decrease in recharge for the other.

Evapotranspiration.—Evapotranspiration data are sparse, even in the most intensively studied parts of the valley (fig. 2). Correlations of selected evapotranspiration data with extensive mapping of vegetation has permitted a far more detailed examination of evapotranspiration than was possible a few years ago. Even so, valleywide evapotranspiration remains a largely unquantified, highly variable component of the ground-water flow model. Given this uncertainty, variations of as much as 25 percent were investigated during the sensitivity analysis. Not surprisingly, these variations produced the greatest overall variations in heads, inflows, and outflows of any parameter in the ground-water flow model. This effect results primarily from the large role that evapotranspiration plays in the ground-water budget and from its broad areal distribution. Changes in evapotranspiration rates were most evident in the simulated gain of water by the river– aqueduct system and the lower Owens River.

Variations in the maximum evapotranspiration rate for the head-dependent evapotranspiration relation (McDonald and Harbaugh, 1988, p. 10–1) produced most of the change in the model. Variations in the depth below land surface at which evapotranspiration was assumed to be zero did not significantly affect the model solution—except that the solution became numerically less stable for depths less than 10 ft.

Underflow.—The quantity of underflow is relatively small in comparison with that of other components of the ground-water budget, but unlike many components, underflow in the model is concentrated in areas of limited extent. Variations in the quantity of underflow from Round Valley (fig. 14) significantly affected heads in that part of the basin. Variations in the quantity of underflow from the Chalfant Valley resulted in slightly different quantities of evapotranspiration near Bishop and some gain or loss of water by the Owens River near Laws. Variations in the quantity of underflow along the Volcanic Tableland made little difference in either nearby heads or gains by the Owens River.

Variations in the quantity of underflow south to the Owens Lake area produced a significant change in heads west of the Alabama Hills and relatively little change in heads east of the Alabama Hills. Much of the potential change in heads east of the Alabama Hills was dampened by changes in gains to the lower Owens River. Values of underflow near Bishop and Big Pine Creeks and near the Waucoba Canyon were locally less important and were not varied as part of the sensitivity analysis.

As was typical of much of the sensitivity analysis, changes in the quantity of underflow were not as evident in heads as in the distribution and quantity of other inflow and outflow components. The hydraulic buffering of heads by evapotranspiration, springs, and surface-water features was repeatedly demonstrated in the sensitivity testing. An analysis of sensitivity of the valleywide model, or similar models (Yen, 1985; Hutchison, 1988; Los Angeles Department of Water and Power, 1988; Radell, 1989), with respect only to changes in head would miss much of the response of the model.

Pumped and flowing wells.—Discharge from pumped and flowing wells was assumed to be known and was not varied as a part of the sensitivity analysis. The effect of withdrawing water from different model layers, however, was investigated. Initially during model development, all water was withdrawn from the lower model layer, and the model matched measured ground-water levels surprisingly well. Subsequently, discharge for each well was split between the upper and lower model layers on the basis of the length of perforations and the estimated hydraulic conductivity of adjacent aquifer materials. The match with measured data did not improve significantly. This is a curious result for a topic that has been thought to be critical in isolating the water table and native vegetation from the effects of pumping. The case of withdrawing all pumpage from the upper model layer was deemed physically impossible and was not simulated.

The causes of the lack of model sensitivity to the vertical distribution of pumpage may be the same as those suggested for the lack of sensitivity to changes in vertical conductance—that is, model cells are large in comparison with individual wells and the simulation period is long. A preliminary simulation model of the Independence fast-drawdown site (fig. 2; tables 1 and 2) used model cells as small as 10 ft on a side and simulated a time period of a few weeks. Results indicated that the smaller model was highly sensitive to changes in the pumpage distribution between layers. Similar results have been suggested by the Inyo County Water Department (W.R. Hutchison, oral commun., 1989).

The lack of sensitivity also may result from the proximity of many production wells to the edge of the confining unit (compare figs. 14 and 17). Over a longer timeframe, the pumping influence reaches the vertically transmissive alluvial fans and is transmitted vertically to both model layers. The confining clay layers are effectively short-circuited because of the geometry of the aquifer and the location of the production wells.

Surface water.—The head-dependent method of simulating the interaction of the aquifer system with the Owens River, the Los Angeles Aqueduct, and the Tinemaha Reservoir allows for adjustments in the prescribed stream stage, altitude of the bottom of the streambed, and conductance of the streambed. Stream stage and altitude of the bottom of the streambed were assumed to be known and were not varied. Variations in streambed conductance identified this parameter as important and narrowly defined. Increasing or decreasing streambed conductance resulted in significantly different gains to or losses from the aquifer system. This response implies that the head-dependent surface-water features exert a strong control on the simulated aquifer system, but do not act as constant heads (McDonald and Harbaugh, 1988, p. 3–16; Franke and others, 1987; S.A. Leake, U.S. Geological Survey, oral commun., 1989).

Springs.—Springs are simulated in the model using the drain package (table 13). Spring discharge is controlled mostly by a conductance term representing the transmissive properties of the spring conduit, such as fractured lava or lava tubes, and by nearby recharge or discharge. A decrease in the conductance of individual springs produced remarkable, although somewhat localized, results. Much of this sensitivity results from the high natural discharges for several springs (fig. 21). In contrast, increases in the conductance of individual springs produced much less effect. These results indicate that the transmissive properties of the spring conduits are much greater than those of the surrounding aquifer materials.

Use, Limitations, and Future Revisions

The valleywide ground-water flow model is best used to help answer questions of regional water use, ground-water flow, and surface-water/ground-water interaction. The conceptualization of the aquifer system described by Hollett and others (1991) provided the basis for a consistent, logical model for nearly the entire basin. This translation from qualitative concepts to quantitative testing was a major purpose for constructing the valleywide model and remains an important use of the model. Additional or alternative concepts of the aquifer system can be tested using the model as presently constructed or using the model as a skeleton for a somewhat different model. If changes to the present model are significant—for example, change in number of model zones, in transmissivities, or in areal extent—then recalibration will be required.

The philosophy and methodology of developing the valleywide model indicate its strengths and possible uses. The modeling technique used in this study was the development of successively more complex models to simulate the aquifer system. The initial model resembled that documented by Danskin (1988). Subsequent site-specific models (fig. 2) were developed to investigate specific questions about the aquifer system (table 2), and information gained from these smaller models was incorporated in the design of the valleywide model. Final refinements in the valleywide model were critiqued in concert with ongoing modeling studies by Inyo County and the Los Angeles Department of Water and Power. In this way, important information was obtained at several different scales and from several different viewpoints. As a result, the valleywide model reflects this technical and numerical consensus. During the cooperative studies, the model played an important role as a neutral, technical arbitrator in answering complex and often volatile water-use questions. Future beneficial use of the model may be in a similar way.

Valuable information gained from design, development, calibration, and sensitivity analysis of the ground-water flow model is not complete. Additional information and insight certainly can be obtained without any new model simulations simply by additional review of model data and results presented in this report. Additional sensitivity analysis may be helpful in identifying which new data are most beneficial in answering water-management questions. Although regional by design, the valleywide model does include many small-scale features and site-specific data and concepts. Future analysis of these smaller-scale features or issues—such as a volcanic deposit, a facies change, or a question of local water use—might best be done by use of smaller-scale models or field studies, in combination with simulations from the valleywide model.

The most appropriate use of the valleywide model is best illustrated by the results presented in this report. The goal in designing both water-management alternatives and figures was to maintain the "regional" character of the model, focusing on larger issues, over longer periods of time. Results are presented precisely (table 11) in order that they can be duplicated and extended; however, use of model results needs to be more schematic—for example, more change occurs in this part of the basin, less in that part. The specific value of drawdown at a well (pl. 1) or for an area of the basin (fig. 23) is far less important than the relative value (more drawdown or less drawdown) in comparison with other areas of the basin. Use of the model in this way will maximize its utility and minimize the limitations.

The primary limitation of the valleywide ground-water flow model is that it is regional in nature. Interpreting results at a scale of less than about 1 mi^2 is inappropriate. The model also is "regional" with respect to the time scale that was chosen for calibration. Interpreting results at a scale of less than a single year is inappropriate. Many limitations of the valleywide model are common to all numerical models and are described by Remson and others (1971), Durbin (1978), Wang and Anderson (1982), Franke and others (1987), and McDonald and Harbaugh (1988). Despite these general limitations of modeling and the specific limitations of the valleywide model of the Owens Valley, as described below, no other methodology provides such a complete testing of ground-water concepts and data.

Interpretation of model results in selected areas of the basin requires special caution. In particular, the area west of Bishop and the area near Lone Pine are simulated poorly. The area west of Bishop has a combination of faults, buried Bishop Tuff, terrace gravel deposits, and abundant recharge. The measured levels and simulated heads (figs. 19 and 20; pl. 1) do not match well, indicating that the model does not

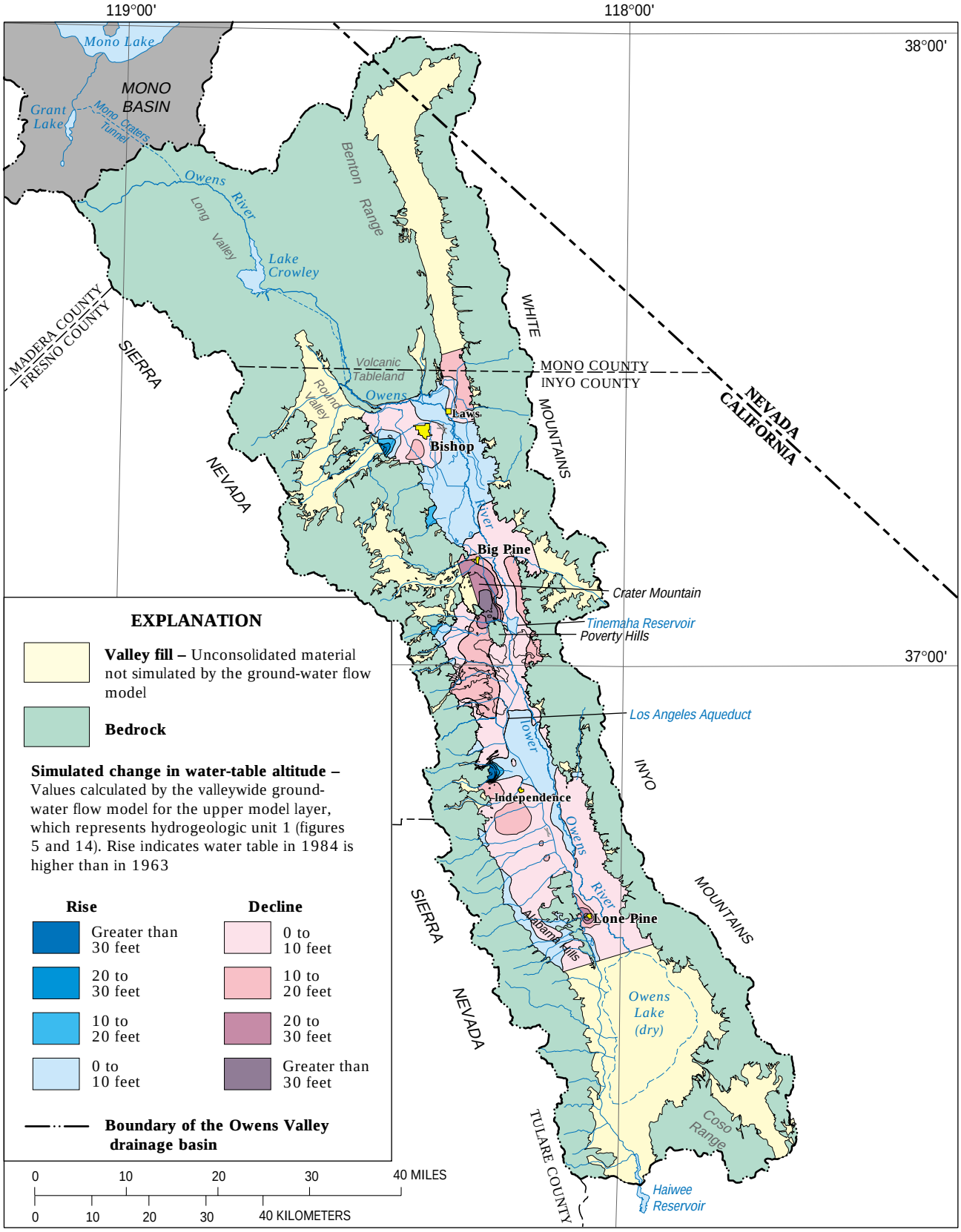


Figure 23. Simulated change in water-table altitude in the Owens Valley, California, between water years 1963 and 1984.

represent actual conditions. It is not clear at this point whether a more detailed simulation of the complex geometry of the Bishop Basin described by Hollett and others (1991) is needed, or if refinement of present hydrogeologic concepts is necessary.

The area around Lone Pine also is simulated poorly. Any number of changes in the model—in the location or hydraulic properties of nearby en echelon faults, in underflow rates, or in recharge from Lone Pine Creek—did little to improve the match for wells in the immediate area, such as well 172 (pl. 1). A basic problem may be that the wells are in small, isolated compartments created by the en echelon faulting. This same phenomenon probably is present north of the Alabama Hills near well 363T (pl. 1). These wells do not interact with the rest of the aquifer system in a way readily approximated by this model. The complex hydrogeology of the areas requires extensive data collection in order to provide the concepts, spatial definition, and parameters necessary to design and calibrate a more accurate numerical model. An alternative method for predicting local ground-water-level changes is to use a simple regression model that avoids many of the spatial and conceptual issues. However, as noted by Hodgson (1978), use of a regression model does not obviate the need for a more rigorous ground-water flow model, at least at a regional scale.

In some parts of the valley, critical hydrologic features are located within a few thousand feet of each other. In the Independence area, for example, the aqueduct, pumped wells, changes in transmissivity and vertical conductance, and changes in vegetation from dryland sagebrush to valley-floor phreatophytes (xerophytes) all are present within about 3,000 ft of each other. Abrupt changes, such as these, result in differences between measured ground-water levels and simulated heads (figs. 19 and 20). From a regional perspective, the differences are acceptable; however, an evaluation of specific local conditions may require a better match.

In the area north of Laws, measured ground-water levels in the immediate vicinity of the boundary of the aquifer system (wells 107T and 252, pl. 1) recover more rapidly than do heads predicted by the model. Although noted, this discrepancy does not affect model simulations or the related results significantly. Simulation of the western alluvial fans and the area east of the Owens River produced reasonable results that seem to validate the basic hydrogeologic concepts about each area; however, an absence of measured data in each area suggests that results in these areas should be interpreted cautiously.

Some of the chosen methods for approximating the aquifer system may produce undesirable effects in some parts of the basin under some conditions. The choice of simulating a constant saturated thickness for hydrogeologic unit 1 may lead to differences in drawdown near sites of significant recharge or pumpage when compared with simulated results that account for changes in saturated thickness. Simulation of canals and ditches only as sources of recharge underestimate their capacity to drain the aquifer system during extended periods of high runoff. The simulation of underflow as a specified, constant rate limits the accuracy of the model for predicting effects of recharge or discharge near a flow boundary, such as north of Laws.

The valleywide model, which simulates the saturated aquifer system, does not incorporate the complex process of vegetative growth and water use as explicit variables, nor does the model simulate the unsaturated soil-moisture zone. Vertical one-dimensional models with these capabilities were developed for selected areas of the valley (table 1 and fig. 2) as a related part of the comprehensive studies of the Owens Valley (Welch, 1988). Incorporating these features in a valleywide model would make it numerically far too large to be useful. The ground-water flow model, however, does simulate changes in the water table and extraction of water from hydrogeologic unit 1 by various processes, including evapotranspiration. With these capabilities, the model can be used to predict areas of the valley where hydrologic stress, such as a decline in the water table or a decrease in ground-water flow rates or discharge, probably will occur.

A key assumption in using the saturated ground-water flow model to evaluate likely effects on native vegetation is that areas of significant hydrologic stress correspond to areas of vegetative stress. In related studies, researchers found that a significant decline in the water table corresponded to a significant stress on native vegetation, particularly rubber rabbitbrush (*Chrysothamnus nauseosus*) (Dileanis and Groeneveld, 1989; Sorenson and others, 1991). Other factors, including alkalinity and salinity (table 3), are

acknowledged to play an important role in the health of native plant communities (fig. 6). Therefore, results from the ground-water flow model should be viewed in general terms as areas of the valley where stress on native vegetation is likely.

A simplification of how the ground-water flow model simulates water use by plants may contribute to an underestimation of water-table recovery during wet periods immediately following dry conditions. During a drought, plants drop leaves in order to limit transpiration and loss of water. During the year following a drought, use of water by plants is restricted (because number of leaves is fewer) until more leaves can be grown. If abundant precipitation falls during this time when the plants have fewer leaves, then the precipitation may satisfy the bulk of the water needs of the plants. Relatively little ground water will be transpired even though ground-water levels are rising because of increased recharge. The ground-water flow model assumes that higher ground-water levels always result in higher evapotranspiration from the ground-water system. This feature may overestimate evapotranspiration during some wet years, and may not allow the simulated water table to recover as rapidly as measured data indicate.

During development of the valleywide model, the simulation of evapotranspiration by native vegetation was studied extensively. Several different approaches were tested, including use of a piecewise-linear, head-dependent relation with a fixed maximum evapotranspiration rate, as described for the final calibrated model; the same relation with a spatially varying maximum evapotranspiration rate based on mapped native vegetation; an evapotranspiration rate based on a separate soil-moisture-box accounting; and an evapotranspiration rate related to precipitation. Each method had its own advantages and disadvantages but yielded surprisingly similar results. This unanticipated conclusion probably stems from the annual approximation of recharge and discharge, the long simulation period, and the regional character of the model. In order to better simulate some transient conditions, future revisions of the valleywide model may consider use of a more complex evapotranspiration package with spatially varying parameters to simulate direct precipitation on the valley floor, antecedent soil moisture, and vegetative growth and water use.

Spatial and temporal discretization of the valleywide model generally does not adversely affect the simulation of regional or subregional water-management issues. The two-layer approximation of the aquifer system produced good results in nearly all areas of the valley. However, a three- or four-layer approximation of the Big Pine and the Taboose–Aberdeen areas, paralleling the conceptualization documented by Hollett and others (1991), would yield a more physically based and possibly more reliable model. Addition of more layers to the model allows a better spatial representation of the complex geometry between pumped volcanic deposits and nearby fluvial and lacustrine deposits, and might result in a more accurate simulation of pumping effects on different parts of the aquifer system. The approximation of numerous individual clay layers by a single confining layer, such as for the fluvial and lacustrine deposits (figs. 4 and 5), yielded good results and does not need to be changed in future revisions of the valleywide model. The present approximation of the massive blue-green clay near Big Pine with a simple Darcian relation is likely to result in inaccurate results for some simulations that are sensitive to the transient propagation of hydraulic head through the thick clay and the concurrent release of ground water from storage in the clay.

The use of model zones to group areas with similar geologic materials (hydrogeologic subunits) was a simple technique that produced good results. Identifying transition-zone deposits as a unique hydrogeologic unit (fig. 5) and incorporating the unit as a separate model zone, as suggested by Danskin (1988), substantially improved simulation along the toes of the western alluvial fans. Additional drilling east of the Owens River would help to confirm the presence and configuration of hydrogeologic subunits and related model zones in that area (pl. 2). A more detailed definition of the hydrogeology of the area west of Bishop is needed and might prompt a redefinition of model zones in that area.

One method of solving some limitations of the valleywide model is to decrease the size of the model grid. A finer grid-spacing facilitates a more gradual change in hydraulic parameters, which produces a better simulation of the aquifer system. Microcomputer capabilities as of 1988 permit design of a valleywide model with three or possibly four layers using a uniform grid size of 1,000 ft on a side. Use of

finite-element techniques facilitates increased spatial resolution in key areas (Danskin, 1988). However, prior to redesigning the present model, certain questions about hydrogeologic concepts need to be answered or the increased numerical resolution will not be accompanied by a commensurate increase in reliability. These questions are itemized in a later section entitled "Need for Further Studies."

Another method of improving the predictive capability of the valleywide model in selected areas of the basin is to use smaller, more detailed models, such as those developed by Inyo County and the Los Angeles Department of Water and Power (table 2). An important caveat in the use of this type of model became apparent during the cooperative studies when a detailed model of the Thibaut–Sawmill area was developed by Inyo County (Hutchison and Radell, 1988a, b). Although the boundary conditions of the smaller model were chosen carefully, the model could not be successfully calibrated. Inspection of the valleywide model revealed that the boundaries of the smaller model, although reasonable under steady-state conditions, were too dynamic under transient conditions to be simulated using the standard modeling techniques described by McDonald and Harbaugh (1988). Only transient specified-flux boundary conditions obtained from the valleywide model were sufficient to achieve a reliable transient simulation. Thus, use of more detailed models may offer advantages, particularly near well fields or spatially complex areas, but the models need to incorporate boundary conditions from a valleywide model.

Both the spatial distribution and method of simulating stream recharge worked well. Although ground-water-level data are sparse for the upper slopes of alluvial fans, the general distribution of recharge along individual streams produced reasonably good results in areas of known levels (figs. 19 and 20; pl. 1). Because of the considerable distance between land surface on the alluvial fans and the underlying water table, a noticeable lag may occur between a measured loss of water in a stream and the resulting response of the aquifer system (well 1T, pl. 1). Although recognized, this lag did not affect simulation results significantly. Future revisions that use stress periods of 6 months or less may need to account for this time lag.

The addition of spring discharge to the model, in comparison with previous modeling efforts by Danskin (1988), produced major improvements in simulating areas along the toes of alluvial fans and edges of volcanic deposits. These areas also are characterized by a relative abundance of water and native vegetation (fig. 3), which might indicate that evapotranspiration rates are higher than in most other parts of the valley. Simulation of these areas might be improved further by locally increasing the maximum evapotranspiration rate.

Future modeling also might benefit from a more detailed simulation of the interaction between the major surface-water bodies and the aquifer system. A variety of physically based relations are available that incorporate the wetted surface area of the interface, the hydraulic conductivity of intervening materials, and temporal variability in the hydraulic head of the surface-water body (Durbin and others, 1978; Yates, 1985; Prudic, 1989). Use of an explicit surface-water model linked to the ground-water flow model would allow more detailed mass balancing of the surface-water system than was possible in this study and would facilitate the development of integrated surface-water/ground-water budgets as suggested by Danskin (1988).

Discussion of Simulated Results, Water Years 1963–88

Calibration and verification of the ground-water flow model for water years 1963–88 enabled both a critique of model performance and an analysis of a critical period of basin operation—in particular, the conditions before and after the second aqueduct was put into operation. Because measured ground-water levels for hydrogeologic unit 1 (upper model layer) were collected at only a few sites prior to 1974, a quantitative analysis of the period requires the use of simulated results.

The simulated change in water-table altitude between water years 1963 and 1984, both times of a relatively "full basin," is shown in figure 23. Simulated conditions for water year 1963 generally reflect average conditions prior to 1970 (table 4). In some parts of the valley, antecedent pumping seems to have affected measured ground-water levels (pl. 1). Because this antecedent pumpage is not included in the model, simulated heads for water year 1963 may be slightly higher than measured levels in those areas. Simulated conditions for water year 1984 also reflect a nearly full

basin, but one after the substantive changes in basin management that occurred in 1970.

Major changes in the simulated water table between water years 1963 and 1984 are obvious in the Laws and the Big Pine areas (fig. 23), and are visible in measured levels (pl. 1). Equally major changes also are suggested beneath western alluvial fans, particularly near the Taboose–Aberdeen well field (fig. 17). Because no measured levels are available in the fan areas, this simulated result is less certain. However, the result is consistent with the large increase in pumpage from the Taboose–Aberdeen and the Thibaut–Sawmill well fields (fig. 17), the decrease in discharge from nearby springs (fig. 21), and the reasonable simulation by the model of other conditions during water years 1963–88.

The relatively wet conditions in 1984 are reflected by the blue areas in figure 23, indicating a rise in the simulated water table. It is important to note that many areas of the valley floor had a rise in the simulated water table between water years 1963 and 1984—even though elsewhere in the valley, the simulated water table declined. This duality of response is typical of the complexity observed in the valleywide system.

One of the primary questions at the beginning of the study was, "What effect does pumping have on ground-water levels and native vegetation in the middle of the valley?" The ground-water flow model was used to investigate this question for the Independence area, an area of intensive monitoring and modeling during the USGS studies (fig. 2 and table 1). Shown in figure 24 are simulation results from the valleywide model for water years 1963–88 at the Independence fast-drawdown site (site K, fig. 2; table 1). Values of ground-water-flow vectors for two periods, water years 1963–69 and water years 1970–84, are shown in figure 24*A*.

The principal components of the vectors show that the dominant ground-water flow direction is horizontal and generally eastward, although there is a significant southward component in hydrogeologic unit 3. These results are comparable to those depicted in figures 14, 19, and 20. As is typical of a layered aquifer, vertical flow rates are significantly less than the total horizontal flow rate in either unit. The difference in flow rates between the two periods is most evident as a decrease in the vertical flow rate, decrease in the evapotranspiration rate, and increase in the southward flow rate in hydrogeologic unit 3.

It is important to note that the vertical flow rate, and the related decrease in vertical flow rate, is a larger percentage of flow in hydrogeologic unit 1 than it is in hydrogeologic unit 3. Pumping may produce relatively minor effects in hydrogeologic unit 3, and at the same time, have a much greater effect on flow rates into and evapotranspiration from hydrogeologic unit 1. Native vegetation depends on the continuous flow of water into hydrogeologic unit 1 and is affected by a change in flow rates. Shown in figure 24*B* is the simulated change in flow rates and evapotranspiration for water years 1963–88. The effect of pumping is clearly evident, beginning in 1970, in simulated flow rates and evapotranspiration at the Independence fast-drawdown site.

The importance of maintaining an adequate ground-water flow rate into hydrogeologic unit 1 also is illustrated in figure 25, which shows a schematic east– west section in the same general area of Independence shown in figure 24. Two conditions are shown in the section (fig. 25)—ground-water levels with and without ground-water pumping. With no pumping, ground-water levels are fairly static. Ground water recharges hydrogeologic units 1 and 3 from the western alluvial fans in proportion to the saturated thickness of each unit. With pumping, the saturated thickness of hydrogeologic unit 1 is decreased, which in turn decreases the quantity of ground water flowing into hydrogeologic unit 1.

Eventually, this decrease will reduce the rate of evapotranspiration from the middle of the valley (fig. 24). This aspect of a fluctuating saturated thickness (time-variant transmissivity) was not simulated by the ground-water flow model; as a result, changes in actual ground-water flow rates into hydrogeologic unit 1 may be somewhat greater than those shown in figure 24.

In summary, the aquifer system, particularly the discharge components, changed significantly with the increase in pumping and export of ground water after 1970. Although changes in water use and distribution of surface water also were made in 1970, most of the changes in the aquifer system resulted primarily from increased ground-water pumpage. The increased efforts at ground-water recharge after 1970 did not compensate for the increased pumpage (table 10).

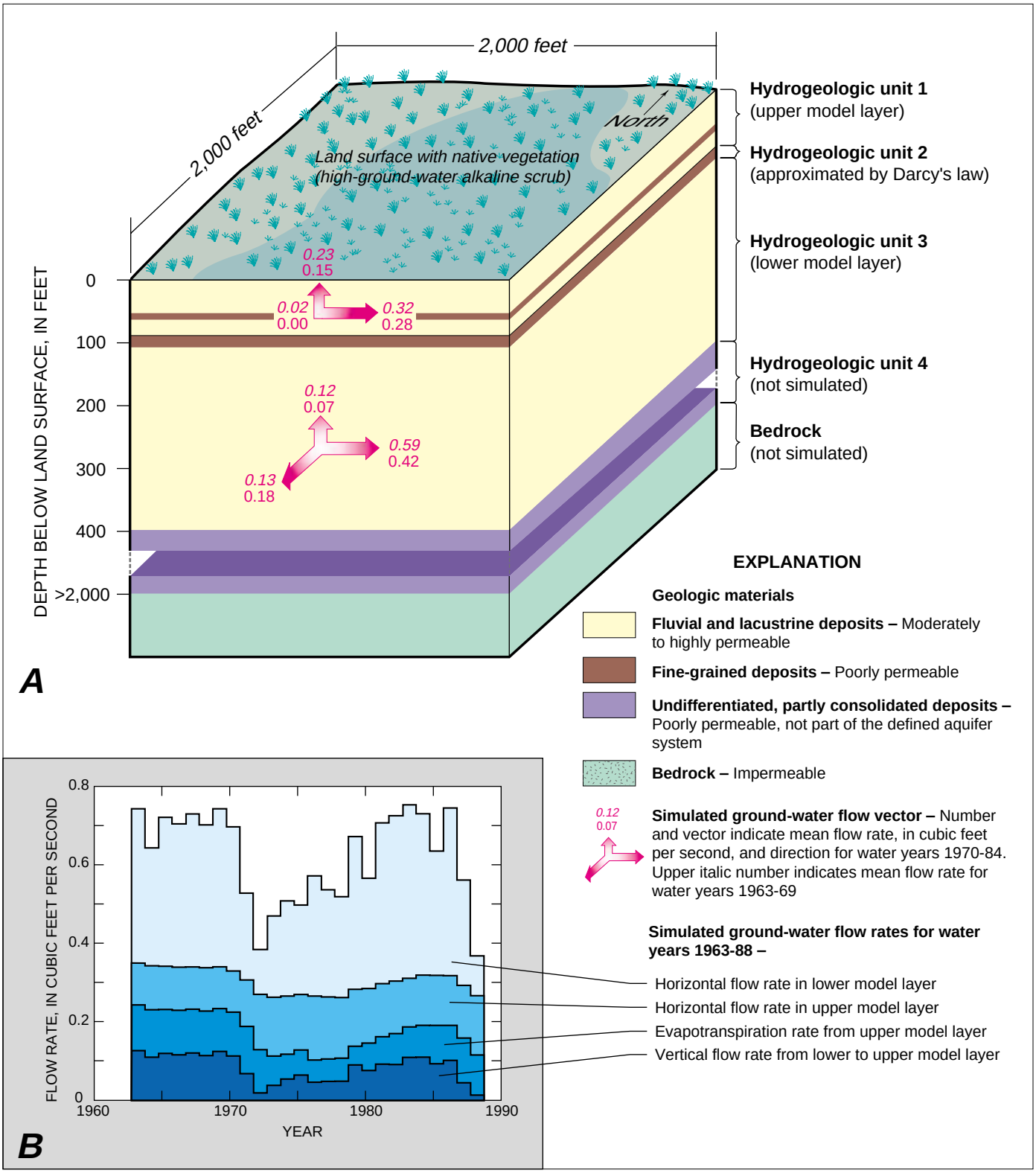


Figure 24. Simulated ground-water flow rates near the fast-drawdown site at Independence, California (figure 2, site K; table 1). ***A,*** average flow vectors for water years 1963–69 and 1970–84 for the ground-water model cell (row 128, column 23) that represents the area surrounding site K. Also refer to section ***C–C'*** (figure 5). ***B,*** annual flow rates for water years 1963–88.

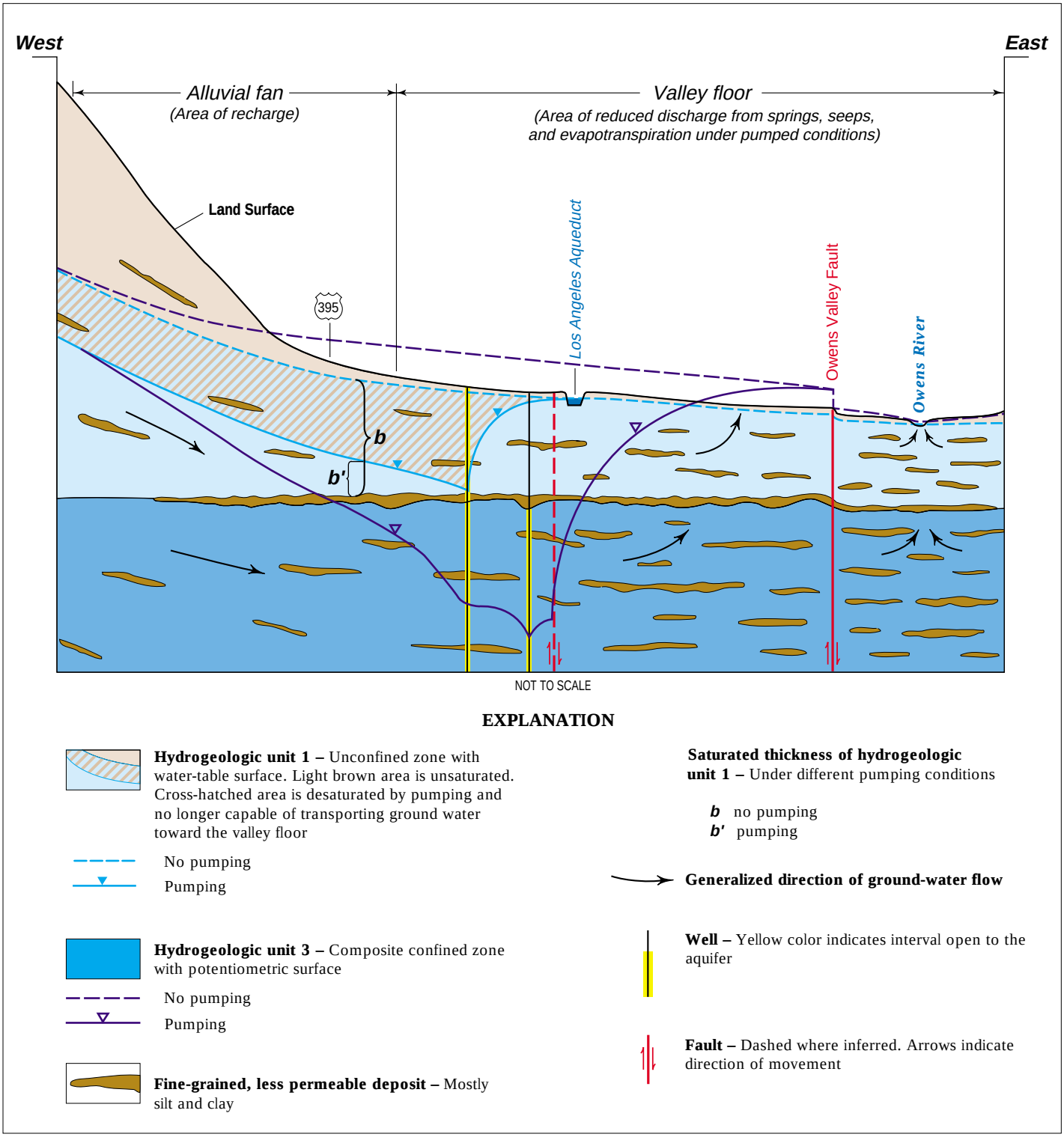


Figure 25. Schematic section across the Owens Valley near Independence, California, showing ground-water flow under different pumping conditions. Saturated thickness of hydrogeologic unit 1 beneath the alluvial fans may decrease markedly (from ***b*** to ***b'***) during pumping and result in significantly less ground-water flow toward the valley flow.