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; $\overline{\text{C}}$, 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)	$\overline{\text{C}}$
		Irrigation (R)	$\overline{\text{C}}$
		Watering of livestock (R)	$\overline{\text{C}}$
		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

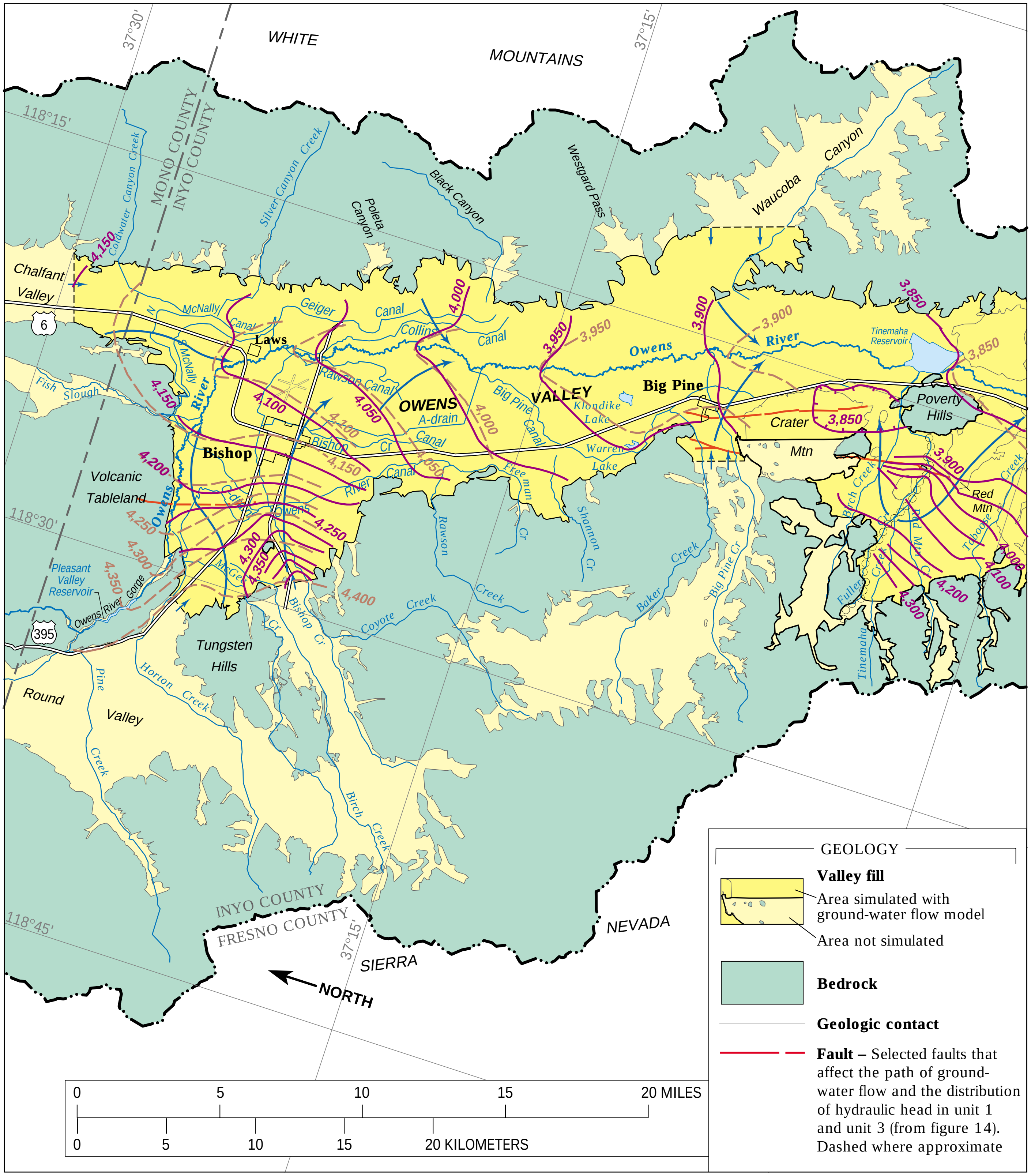
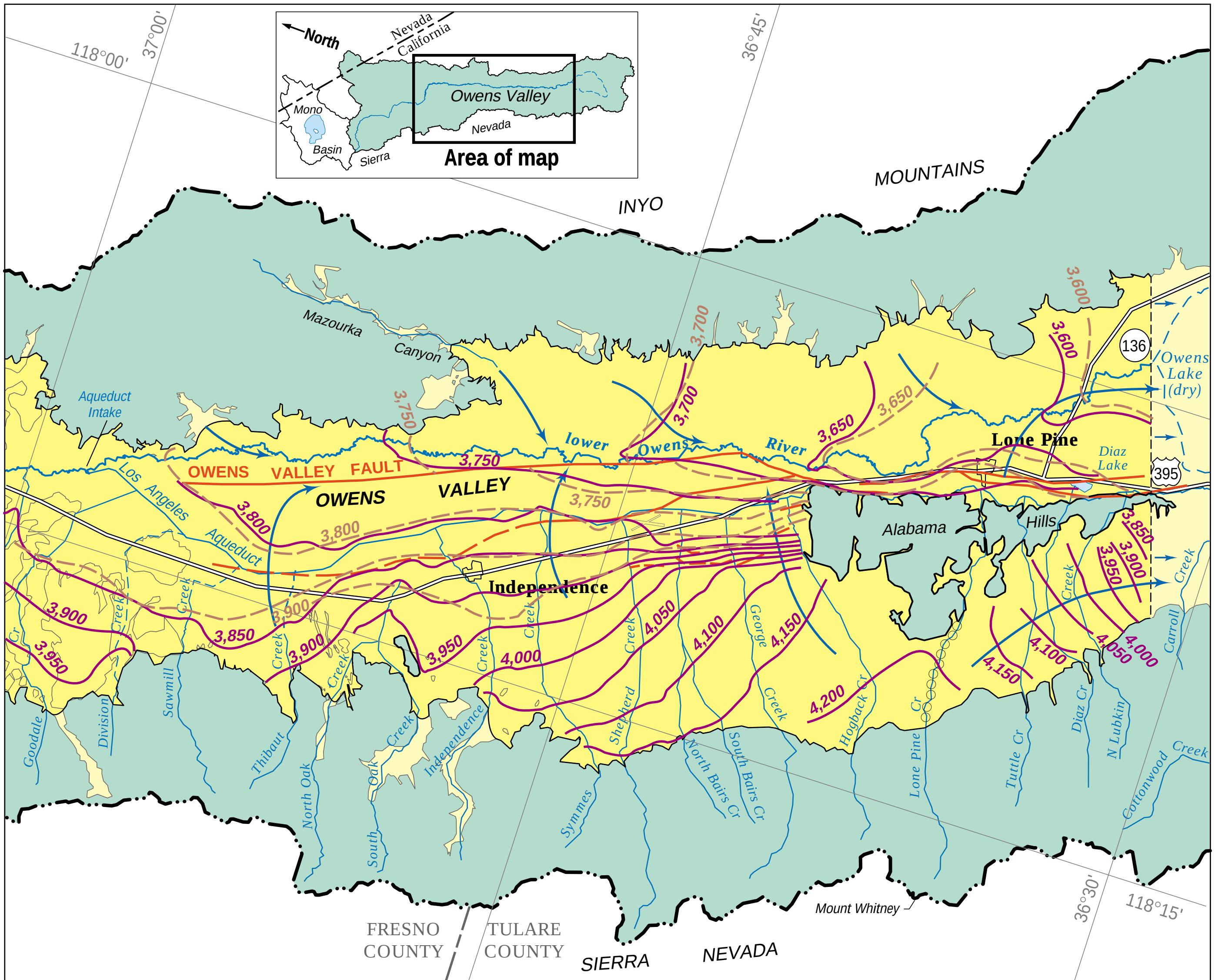


Figure 19. Measured and simulated potentiometric surfaces for hydrogeologic unit 1 (upper model layer) in the Owens Valley, California, spring 1984.



EXPLANATION

HYDROLOGY

3,900 **Measured**

3,900 **Simulated**

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

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.

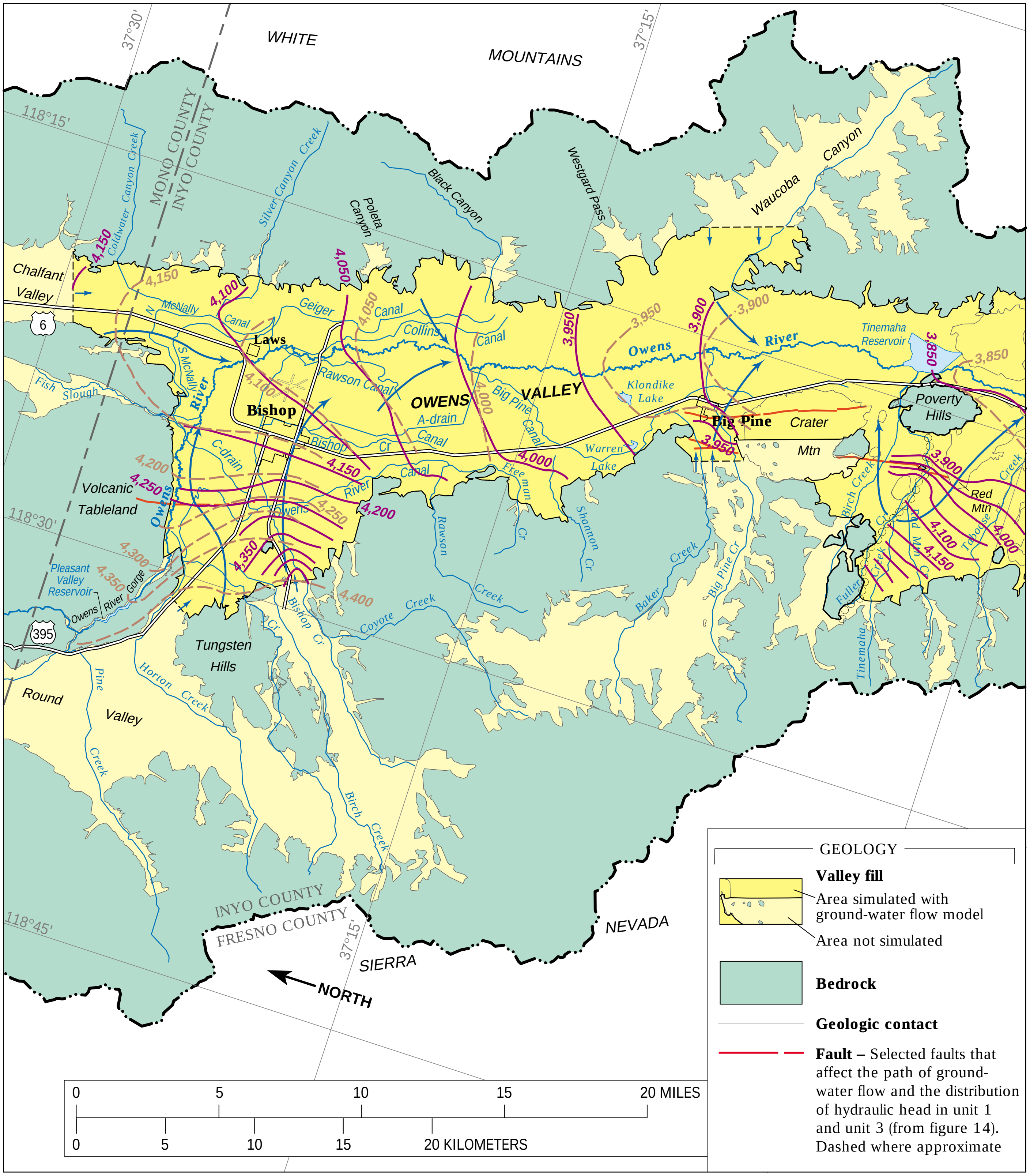
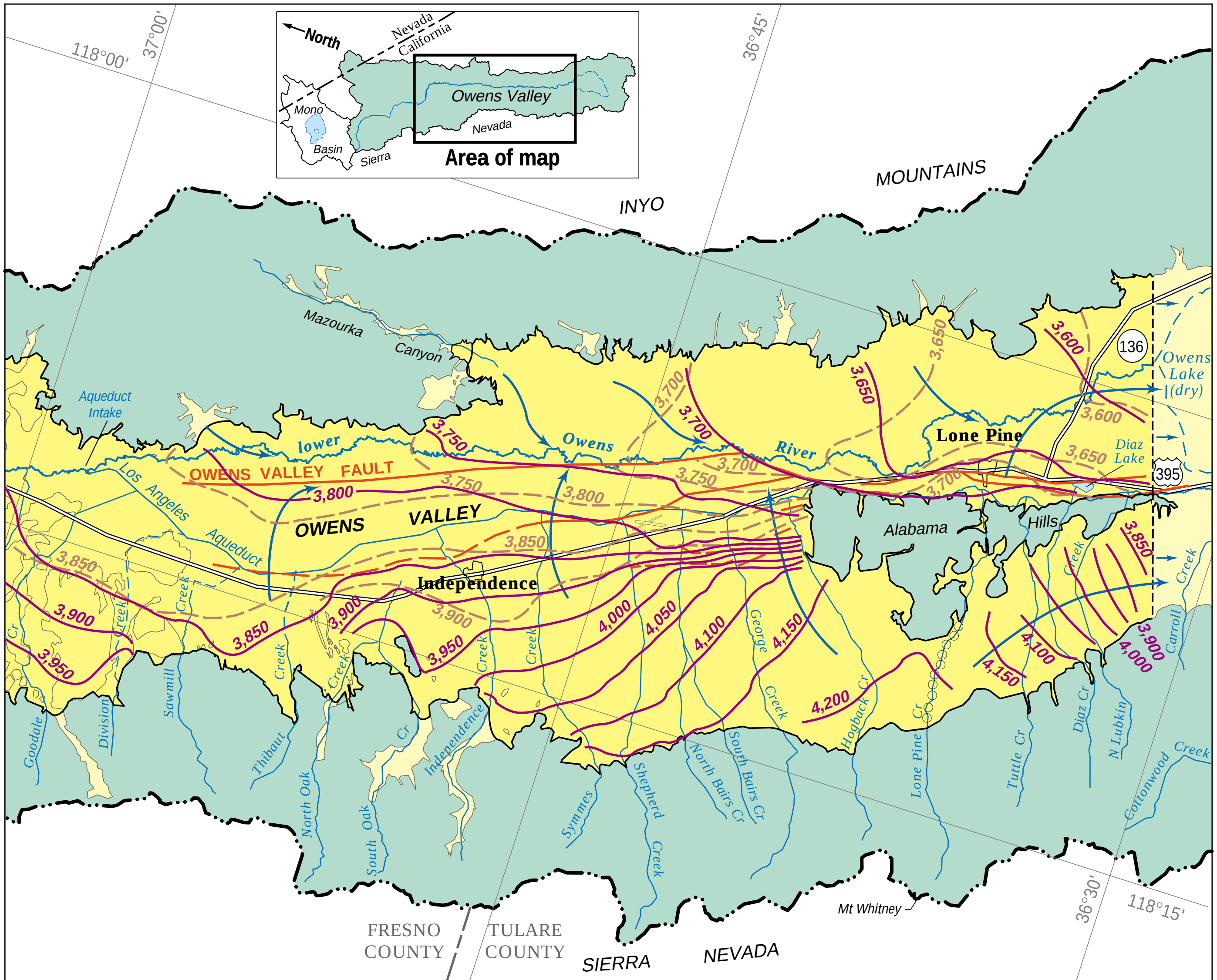


Figure 20. Measured and simulated potentiometric surfaces for hydrogeologic unit 3 (lower model layer) in the Owens Valley, California, spring 1984.



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.

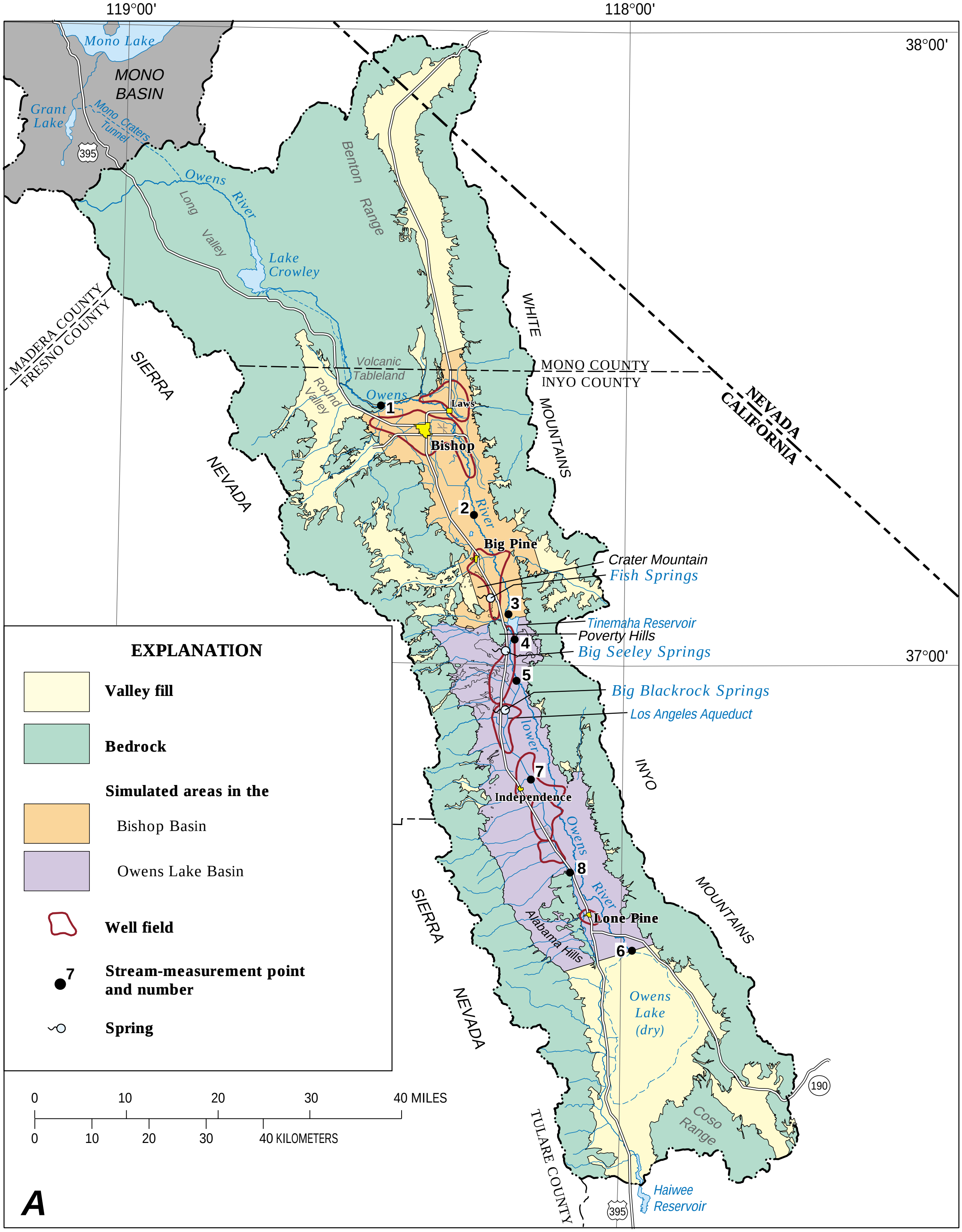
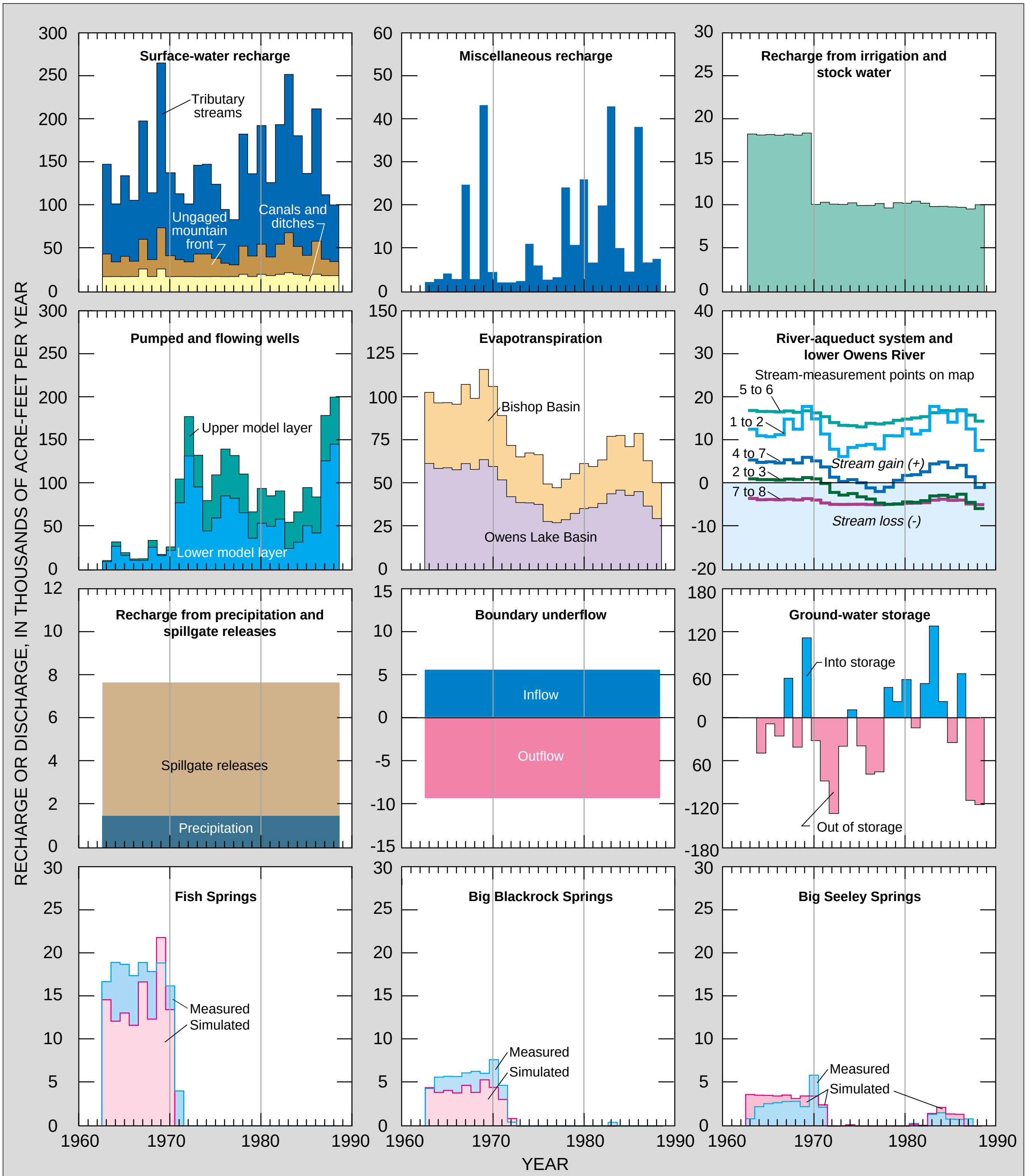


Figure 21. Simulated ground-water recharge and discharge during water years 1963–88 in the Owens Valley, California. Values for each water-budget component are given in table 11.



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.

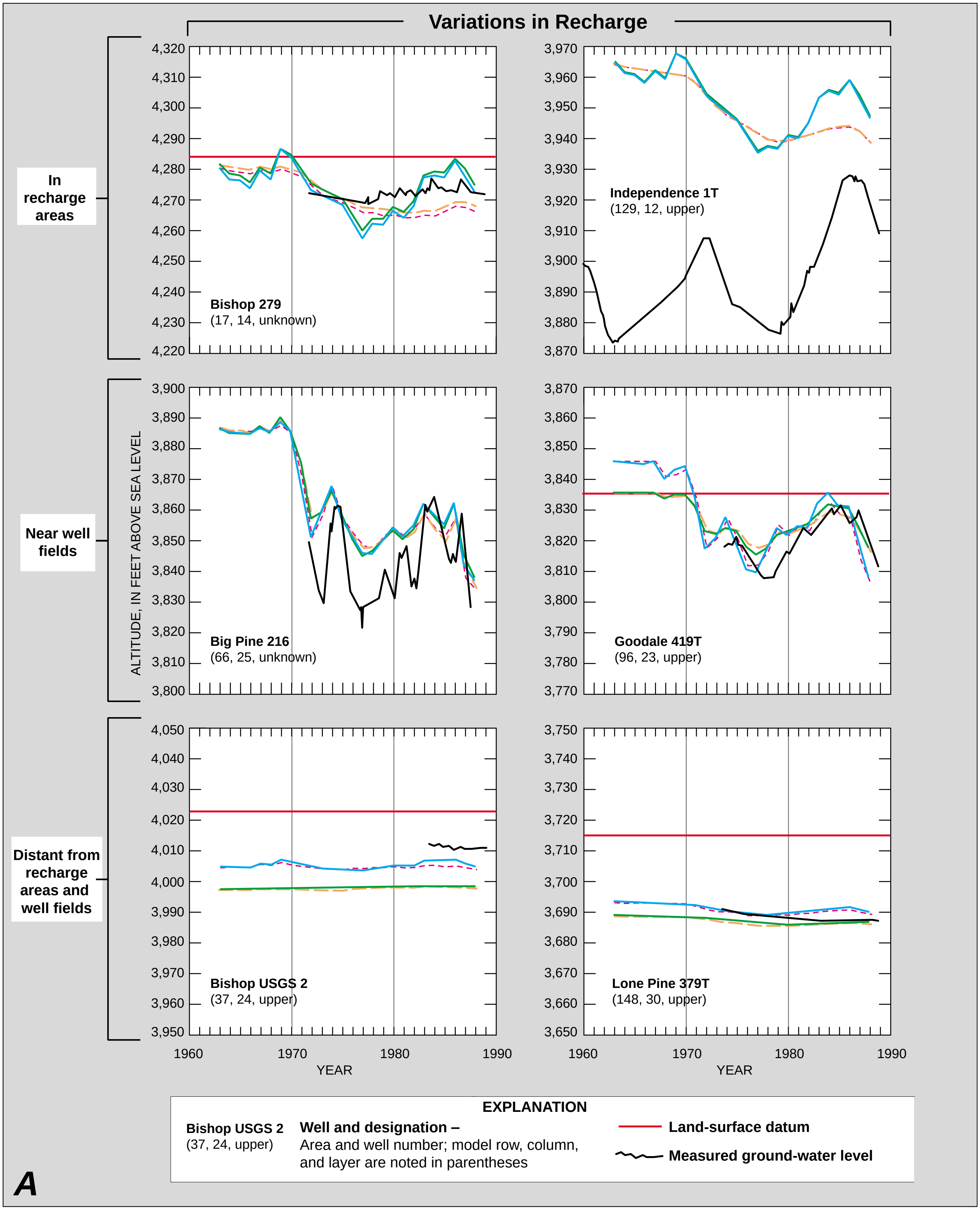


Figure 22. Sensitivity of simulated hydraulic heads in the Owens Valley, California, to variations in recharge ***(A)*** and pumpage ***(B)*** at wells in recharge areas, near well fields, and distant from both. Method of variation is described in text. Well locations are shown on plate 1.

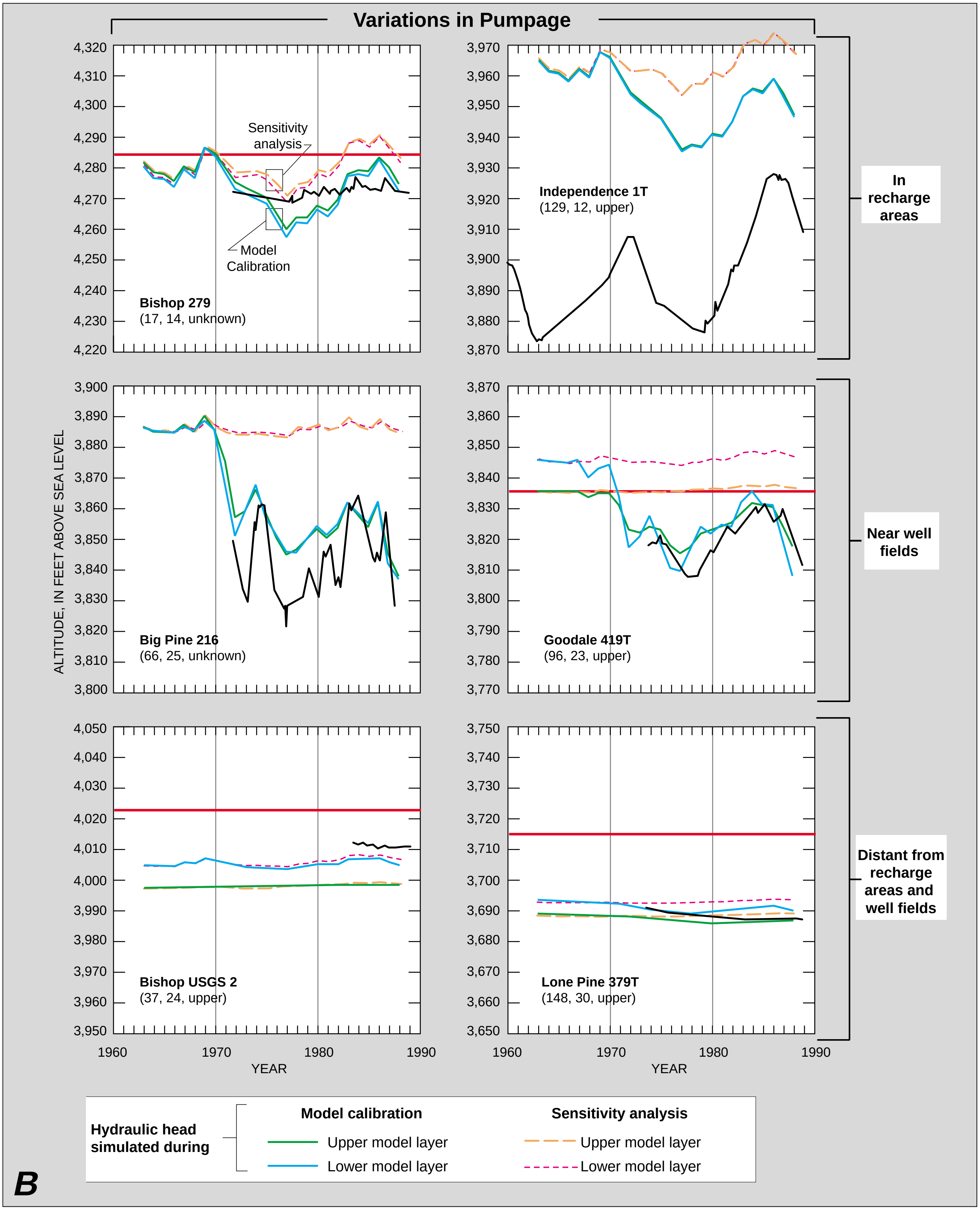


Figure 22. Continued.

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

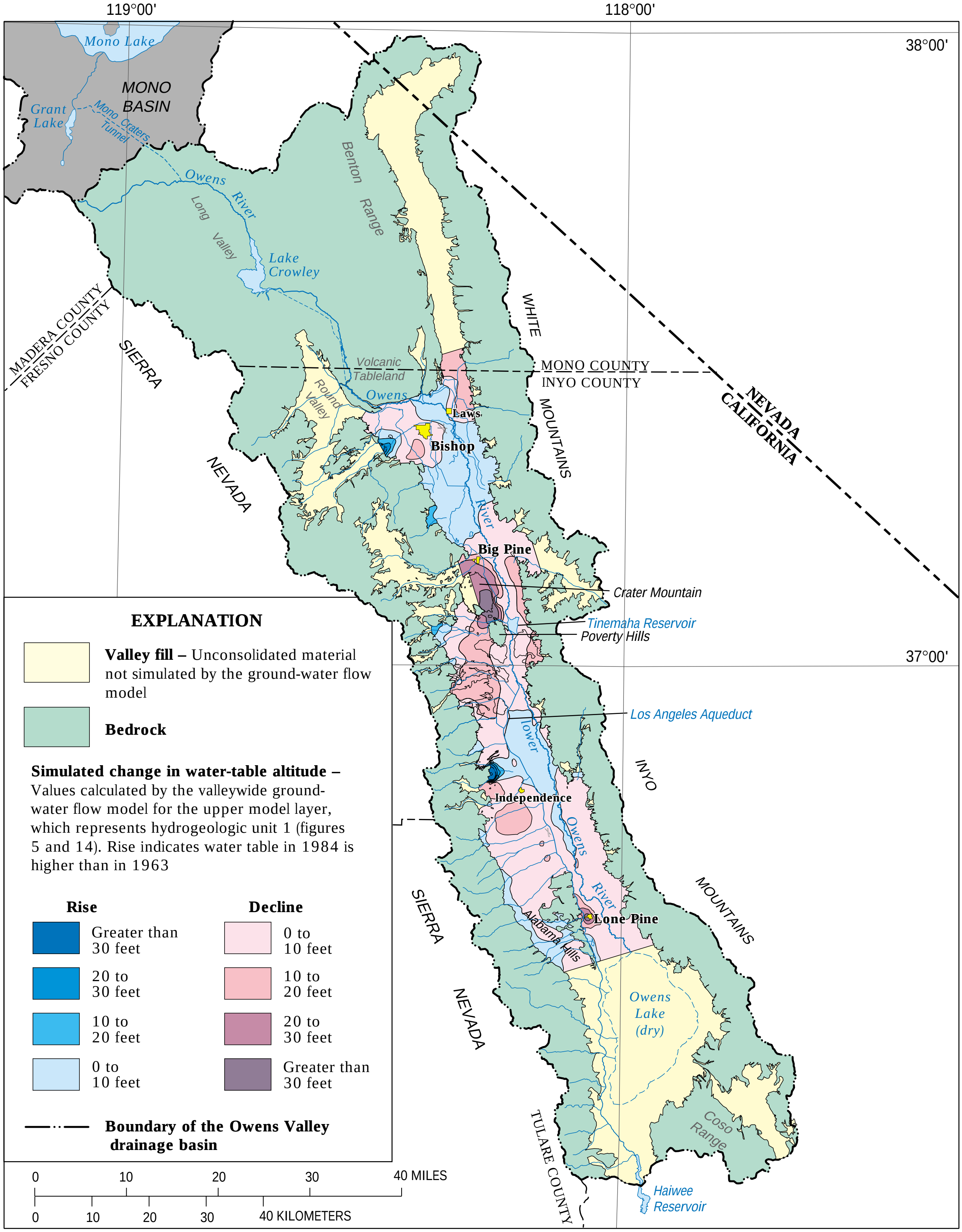


Figure 23. Simulated change in water-table altitude in the Owens Valley, California, between water years 1963 and 1984.

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).

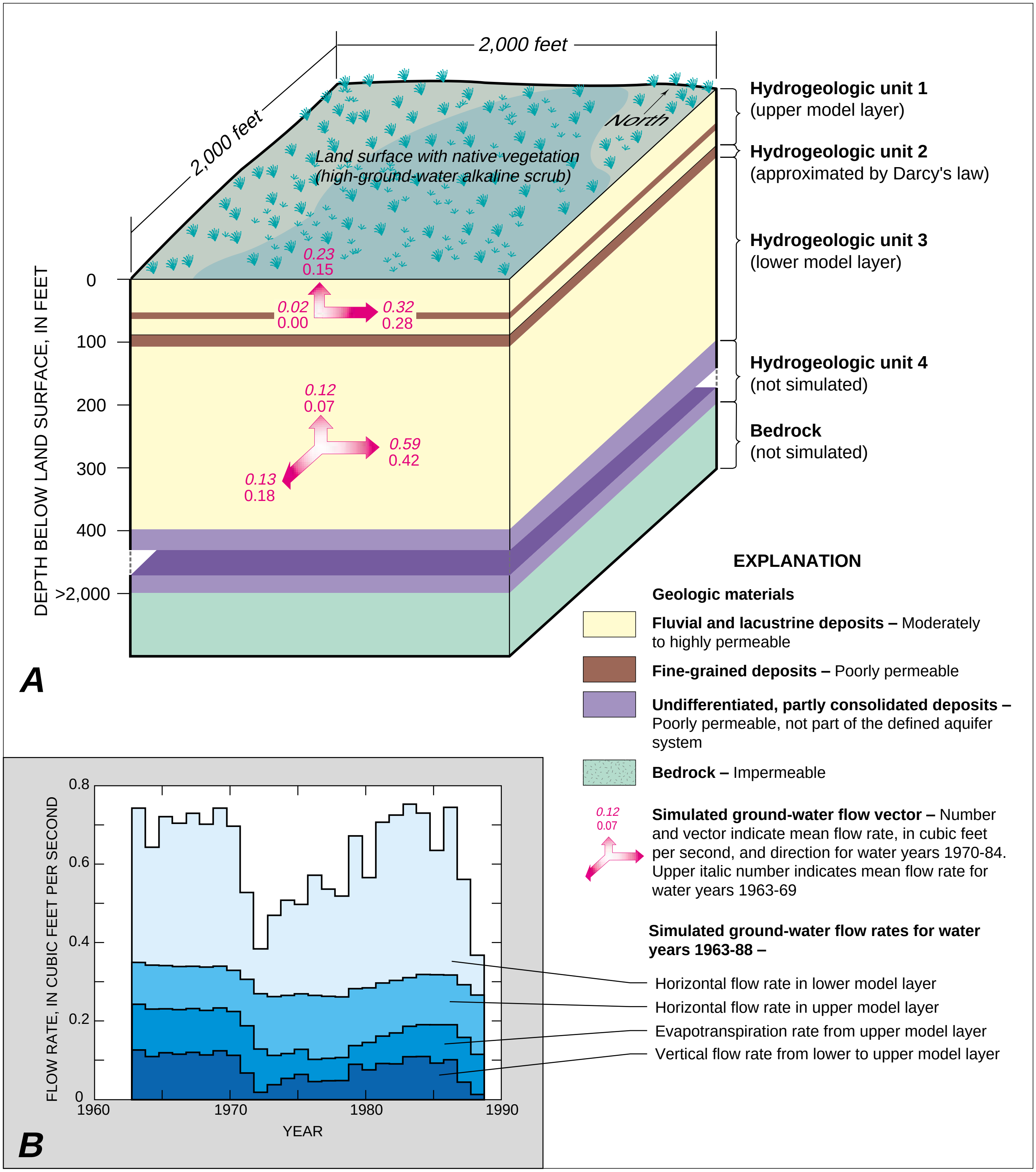


Figure 24. Simulated ground-water flow rates near the fast-drawdown site at Independence, California (figure 2, site K; table 1). ***A,*** average flow vectors for water years 1963–69 and 1970–84 for the ground-water model cell (row 128, column 23) that represents the area surrounding site K. Also refer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

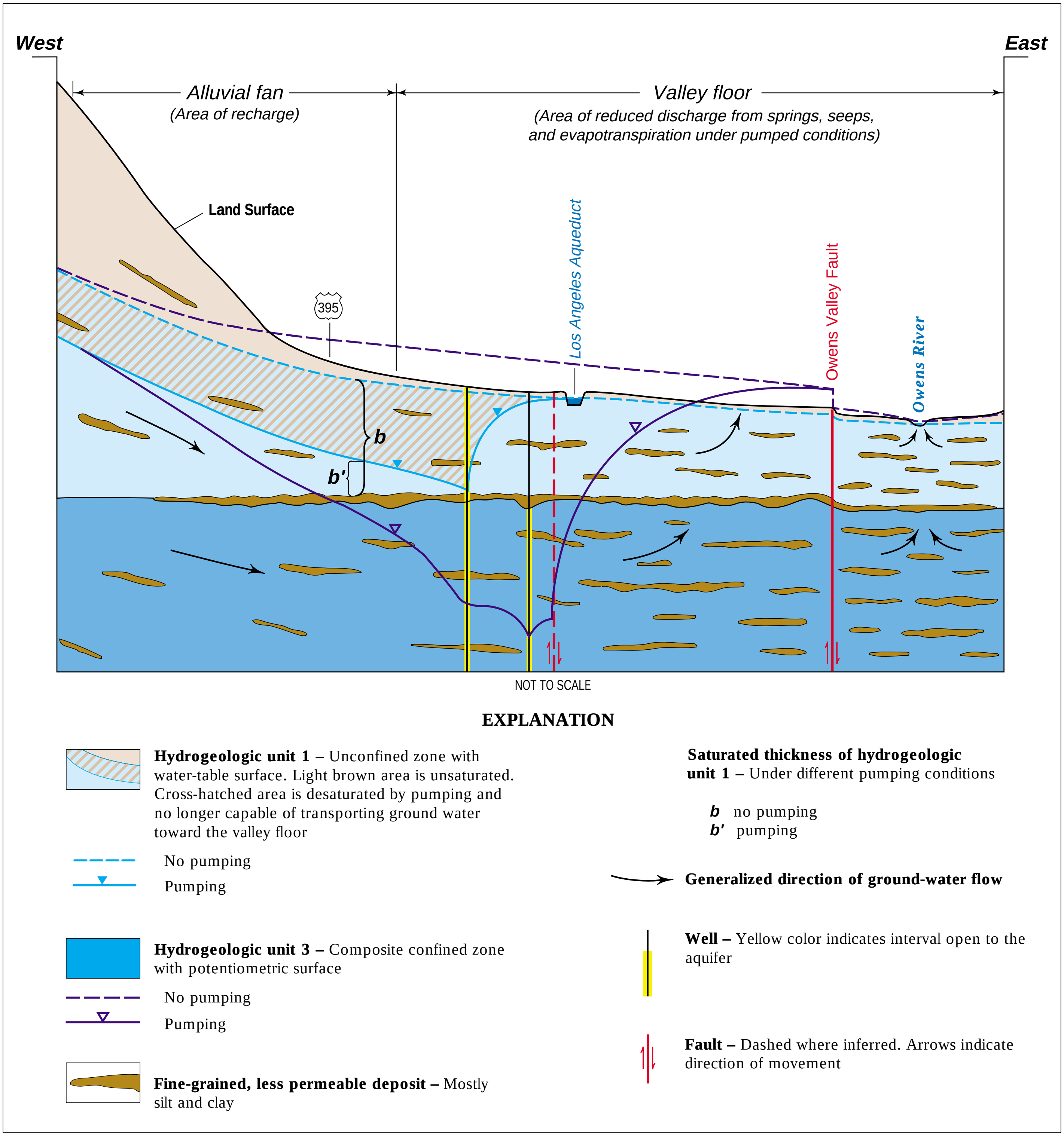


Figure 25. Schematic section across the Owens Valley near Independence, California, showing ground-water flow under different pumping conditions. Saturated thickness of hydrogeologic unit 1 beneath the alluvial fans may decrease markedly (from ***b*** to ***b'***) during pumping and result in significantly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

EVALUATION OF SELECTED WATER-MANAGEMENT ALTERNATIVES

An evaluation of alternative methods of water management involves an appraisal of the present (1988) operating conditions and the physical and social constraints that restrict changes in operations. This evaluation recognizes the social constraints, but focuses on the hydrologic constraints, recognizing that although social constraints might seem to be more encumbering, they often are far less static than the physical constraints presented by precipitation, streamflows, and the aquifer system. Much of the evaluation relies on simulation results from the valleywide ground-water flow model to quantify the likely effects of different management alternatives.

General Water-Management Considerations

Water management of the Owens Valley involves a complex array of conflicting needs and desires. The residents of the Owens Valley need water for local uses such as ranching and domestic supply. Many of the residents desire that water be used for the aesthetic aspects of the valley such as flowing streams and to provide the water needs of native vegetation. The Los Angeles Department of Water and Power, although recognizing these local needs and desires, has continuing needs to export water to Los Angeles. As regional water supplies dwindle and the population of southern California increases, Los Angeles may desire to export additional high-quality water from the Owens Valley. In the difficult task of balancing conflicting needs and desires, the emotional side of water-management issues often tends to take precedence over otherwise purely technical issues.

The goals of water management in the Owens Valley consist of fulfilling both needs and desires. The primary goals include supplying sufficient water for local domestic, ranching, and municipal uses; for native vegetation and aesthetics; and for export to Los Angeles. Secondary goals include mitigation of pumping effects on native vegetation in the immediate area of wells and enhancement of selected areas of the valley. Inherent in achieving these secondary goals, if other water-management practices are continued, is an acceptance of a likely overall decrease in the quantity of native vegetation in other areas of the valley. An ongoing management goal since 1970 has been to decrease consumptive use of water on ranches and lands leased by the Los Angeles Department of Water and Power and to use water more efficiently throughout the valley. Achievement of each of these goals is limited by a variety of considerations that constrain water management in the Owens Valley. The major considerations are described below.

Regional water supplies.—The Owens Valley is part of a much larger network of water supplies, transport, and use. In southern California, water is obtained from a limited number of sources, primarily from northern California, the Colorado River, and the Owens Valley. The use and export of water from the Owens Valley must be viewed within the larger issues of water supply and demand within the arid Southwest, particularly southern California.

Export of surface and ground water.—Water-gathering activities along the aqueduct, primarily north of the Owens Valley in the Mono Basin and the Long Valley, contribute to the total export of water to Los Angeles. A series of reservoirs and ground-water basins along the aqueduct system between the Mono Basin and Los Angeles are used to regulate flow and to store water from one year to the next. Because these storage capacities, in general, are limited, a nearly constant export of water from the Owens Valley is desired. Since 1970, ground-water withdrawals from the Owens Valley have been used to augment surface-water diversions. In an average-runoff year, some ground water typically is exported; however, in a below-average runoff year, the quantity of ground-water exported out of the valley is increased significantly to make up for the shortage in surface water.

Antecedent conditions from the previous water year affect the quantity of export desired by the Los Angeles Department of Water and Power. If antecedent conditions are dry, then less water is stored in reservoirs and ground-water basins along the aqueduct system, and more water is needed from the Owens Valley. As shown in figure 18, the antecedent conditions in turn affect the quantity of ground water that is pumped. If the preceding year has had average or above-average runoff, then ground-water pumpage is less.

The exportation of water from the Owens Valley to Los Angeles has been the subject of many controversies and lawsuits. Historically, California water law has been interpreted to require maximum beneficial use of water (State of California, 1992). In the early 1900's, beneficial use was nearly synonymous with reclamation of the land for farming and for industrial and municipal use. Since about 1970, the historical beneficial