Ground-Water Resources of Honey Lake Valley, Lassen County, California, and Washoe County, Nevada

By Elinor H. Handman, Clark J. Londquist, and Douglas K. Maurer

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CONVERSION FACTORS AND ABBREVIATIONS

Except for water quality, "inch-pound" units of measure are used in this report and may be converted to metric (International System) units by using the following factors:

Multiply	*By*	*To obtain*
Acre	0.4047	Square hectometer (hm^2)
Acre-foot (acre-ft)	1,233	Cubic meter (m^3)
Acre-foot per year (acre-ft/yr)	0.001233	Cubic hectometer per year (hm^3/yr)
Cubic foot per second (ft^3/s)	0.02832	Cubic meter per second (m^3/s)
Cubic foot per second, per mile [(ft^3/s)/mi]	0.01760	Cubic meter per second, per kilometer [(m^3/s)/km]
Foot (ft)	0.3048	Meter (m)
Foot per day (ft/d)	0.3048	Meter per day (m/d)
Foot per day, per foot [(ft/d)/ft]	1.000	Meter per day, per meter [m/d)/m]
Foot per year (ft/yr)	0.3048	Meter per year (m/yr)
Foot squared per day (ft^2/d)	0.09290	Meter squared per day (m^2/d)
Inch (in.)	25.40	Millimeter (mm)
Inch per day (in/d)	25.40	Millimeters per day (mm/d)
Mile (mi)	1.609	Kilometer (km)
Square mile (mi^2)	2.590	Square kilometer (km^2)

For temperature, degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) by using the formula °F = [1.8(°C)] + 32.

EQUIVALENTS

Multiply	*By*	*To obtain*
Cubic foot per second (ft^3/s)	448.83	Gallons per minute (gal/min)
Cubic foot per second (ft^3/s)	724.5	Acre-foot per year (acre-ft/yr)
Foot squared per day (ft^2/d)	7.48	Gallons per day per foot [(gal/d)/ft]
Foot per day (ft/d)	7.48	Gallons per day per square foot [(gal/d)/ft^2]

SEA LEVEL

In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929), which is derived from a general adjustment of the first-order leveling networks of both the United States and Canada, formerly called "Sea-Level Datum of 1929."

GROUND-WATER RESOURCES OF HONEY LAKE VALLEY, LASSEN COUNTY, CALIFORNIA, AND WASHOE COUNTY, NEVADA

By Elinor H. Handman, Clark J. Londquist, and Douglas K. Maurer

ABSTRACT

Honey Lake Valley is a 2,200-square-mile, northwest-trending, topographically closed basin about 35 miles northwest of Reno, Nevada. Unconsolidated basin-fill deposits on the valley floor and fractured volcanic rocks in the northern and eastern uplands are the principal aquifers. Ground water in the Nevada part of the basin is being considered as a potential source of public supply to the Reno area. This report describes the ground-water resources of the basin and the potential hydrologic effects of ground-water development in its eastern part.

During an average year, about 1.1 million acre-feet of precipitation falls on the 1,700-square-mile study area within the valley, and an estimated 86,000 acre-feet enters the study area as streamflow in the Susan River and Long Valley Creek. Almost 90 percent of this water runs off, evaporates, or is transpired by plants before it can infiltrate and become ground-water recharge; this includes about 130,000 acre-feet per year that reaches Honey Lake as surface-water inflow. About 130,000 acre-feet recharges the aquifer system annually, about 40 percent by direct infiltration of precipitation over the study area and about 60 percent by infiltration of streamflow and irrigation water. Balancing this is an equal amount of ground-water discharge, of which about 65 percent evaporates from the water table or is transpired by phreatophytes, about 30 percent is withdrawn from wells, and about 5 percent leaves the basin as subsurface outflow to the east.

A ground-water flow model of the eastern part of the study area, where withdrawals for public supply have been proposed, was used to evaluate components of the water budget and to estimate the long-term hydrologic effects of hypothetical increased development. Results of the evaluation indicate conditions under which a new equilibrium would be established if 15,000 acre-feet of water were withdrawn from the Fish Springs Ranch area annually. The model indicates that, as a result of such an increase in pumpage, water levels would decline more than 100 feet from present (1988) levels in the vicinity of the pumping, and as much as 40 feet at the California-Nevada State line. Evapotranspiration and subsurface outflow to the east would each be reduced by about 60 percent, but hydrologic effects would be minimal at the western boundary of the flow-model area (just east of Honey Lake). Within the modeled area, the simulation indicates that the increased withdrawals would cause the net flow of ground water eastward across the State line to increase from about 700 acre-feet per year to about 2,300 acre-feet per year.

INTRODUCTION

The Reno-Sparks area in western Nevada is one of the fastest growing population centers in the United States. Nearly all economically available surface water has been allocated, and as development continues, so does the demand for water. The principal water-utility company in the area has identified an aquifer system in Honey Lake Valley, about 35 miles northwest of Reno-Sparks, as a possible source of water for the two cities and for the unincorporated areas of Washoe County (Westpac Utilities, 1989). As a result, the county has entered into a public-private venture with Western Water Development Company to develop the potential resource.

Decisions about development of the ground-water resources in Honey Lake Valley are complicated because only about one-fifth of the drainage area is in Nevada, and four-fifths is in California. Development of the aquifer system probably will affect the quantity of interstate flow, and water transported out of the basin would be diverted from its natural discharge areas. An assessment of the potential effects of interstate flow and interbasin transport requires a detailed evaluation of the water budget and the local and regional flow systems in the central and eastern parts of Honey Lake Valley. This study by the U.S. Geological Survey, in cooperation with the California Department of Water Resources and the Nevada Division of Water Resources, is an appraisal of ground-water resources in the basin.

Purpose and Scope of This Report

This report describes and quantifies the water budget for Honey Lake Valley (figure 1). It describes the components of ground-water flow and the characteristics of the aquifer system in Honey Lake Valley on the basis of results of a 3-year study. The ground-water component of the budget for part of the study area east of Honey Lake is further evaluated by means of a three-dimensional, finite-difference mathematical model (figure 1) that simulates the ground-water flow system in unconsolidated basin-fill deposits and consolidated volcanic rocks. The report includes a discussion of the direction and magnitude of ground-water flow at the California-Nevada State line.

The scope of the report is limited to discusssion of the ground-water flow system. Some of the potential hydrological effects of increased ground-water withdrawals are considered. However, an assessment of the potential effects of increased development on vegetation, wildlife, and water quality was not within the scope of this investigation.

Previous Studies

A reconnaissance report (Rush and Glancy, 1967) provided a preliminary estimate of the ground-water budget for a 235-mi^2 area in the eastern part of Honey Lake Valley, primarily in Nevada. Ground-water resources were evaluated from an inventory of water-level measurements at 13 wells and limited additional subsurface information. The report indicated an imbalance of 9,000 acre-ft/yr between estimates of recharge (2,000 acre-ft) and discharge (11,000 acre-ft). According to Rush and Glancy (p. 42) the difference may have resulted from an underestimate of recharge from precipitation or from subsurface inflow to Honey Lake Valley.

An investigation of ground water in the northeastern counties of California included a section on the California part of Honey Lake Valley (California Department of Water Resources, 1963a, p. 205-220). The study considered ground-water recharge, movement, and availability. The valley floor was divided into four zones that were classified in terms of potential for development. The report was updated in 1987 (Pearson, 1987) and water budgets were produced for 1986 (Muir, 1988) and for long-term average conditions (Clements, 1988).

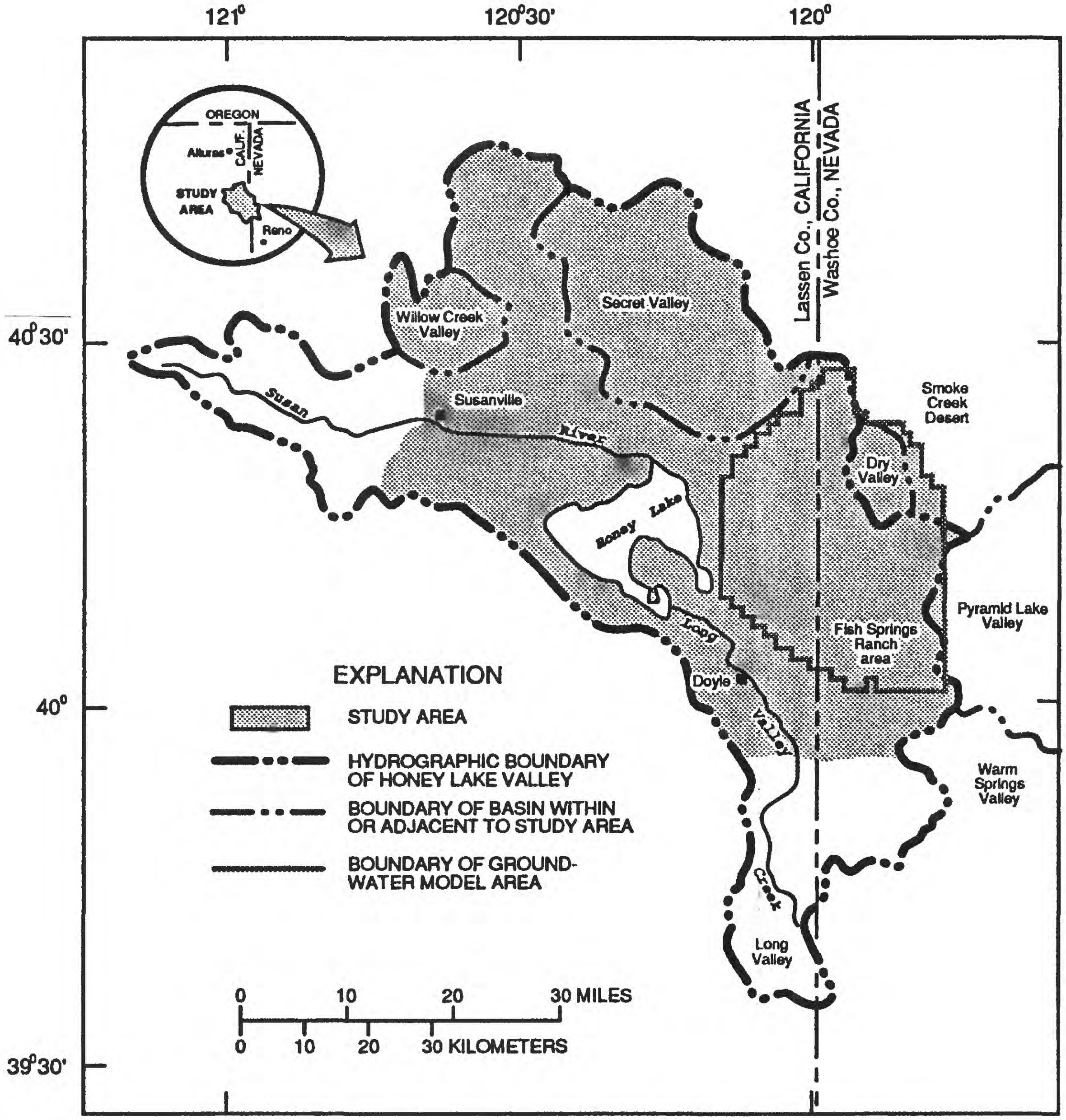


FIGURE 1.--Location of study area.

Water resources were described for a 300-mi² area in the southeastern part of the basin in California (Hilton, 1963). The study used information from approximately 50 wells to determine the altitude of the water table in the vicinity of the Sierra Army Depot near Herlong; chemical analyses of water from 13 wells were tabulated. The general description of water resources in that report did not include estimates of the quantities of ground water available for development.

Water-quality conditions in the California part of the basin were analyzed in two reports (Clawson, 1968; Wormald, 1970). Ground water in the area north of Honey Lake contains arsenic, probably of geothermal origin and related to faults. The geothermal resources of Honey Lake Valley have been the subject of numerous studies (U.S. Bureau of Reclamation, 1982; McNitt and others, 1981; Sanyal and others, 1984; GeoProducts Corporation, 1982, 1984; Juncal and Bohm, 1987; Harding Lawson Associates, 1989a, 1989b). The results of some of the test drilling, geophysical surveys, structural analysis, and geochemical sampling from these studies were used in the evaluation of regional flow.

In 1986, Lassen County published the results of a water budget study for Honey Lake Valley that estimates annual ground-water availability for the basin as 69,000 acre-ft (53,000 in California and 16,000 in Nevada; Walters Engineering, 1986, p. 1). Their budget is based on information compiled from previous investigations.

A study of ground-water availability in the Nevada part of Honey Lake Valley was produced for Westpac Utilities, a division of Sierra Pacific Power Company (William F. Guyton Associates, 1987). The study concludes that ground-water recharge from precipitation may be 1,000 to 2,000 acre-ft/yr greater than that estimated by Rush and Glancy (1967) and that some subsurface inflow may also occur.

A reconnaissance study by R.W. Beck and Associates (1987) for the Regional Water Planning and Advisory Board of Washoe County evaluates importation of ground water from sources in northern and central Washoe County, including Honey Lake Valley, to areas in the southern part of the County. The report states that an estimated 16,500 acre-ft/yr of water could be developed from Honey Lake and Warm Springs Valleys, using only water that originates as precipitation in Nevada. The total includes 4,500 acre-ft of assumed underflow from Warm Springs Valley to Honey Lake Valley.

Location and Physiographic Features

Honey Lake Valley is a northwest-trending, elongated, and topographically closed basin in northwestern Nevada and northeastern California and is on the west edge of the Great Basin Physiographic Region. The largest population center in the basin is Susanville, Calif. The northern part of Honey Lake Valley includes Willow Creek, Secret, and Horse Lake Valleys (plate 1). In this report, the entire Honey Lake Valley topographic basin is referred to as "the basin." The development of topographic features, drainage patterns, and ground-water flow systems of the basin is controlled by the major stratigraphic and structural features of bedrock and by climate. Basin-and-range faults, the principal structural features, provide conduits for ground-water flow in some places but obstruct flow in others.

The Honey Lake Valley study area extends about 60 miles from northwest to southeast (figure 1), and encompasses about 1,700 mi^2. The southeast end of the study area is about 35 miles northwest of Reno, Nev. The study area (figure 1) excludes the Susan River drainage area upstream from the Susanville gage and the Long Valley Creek drainage south of Dry Valley Creek. However, inflow from these areas was measured and included in the water budget for the study area. The study area includes the Willow Creek, Secret, and Horse Lake Valley drainage areas, because inflow to the Honey Lake Valley floor from these basins is partly in the subsurface and cannot be estimated at any one place. For the purposes of modeling ground-water flow, the study area also includes Dry Valley north of Flanigan and a small part of Smoke Creek Desert (figure 1, plate 1).

The valley floor is surrounded by mountains to the west, southwest, and east, and volcanic uplands to the north. The Susan River, a few perennial streams, and several intermittent streams flow toward the center of the basin where water accumulates in Honey Lake. Honey Lake is a shallow lake that dries up at times and has no surface outflow. Alluvial fans, composed of materials eroded from the adjacent mountains and uplands, form sloping areas between the uplands and the flat, central valley floor.

Acknowledgments

Considerable information and assistance in collection and analysis of the data upon which interpretations in this report are based was provided by the California Department of Water Resources, and the Nevada Division of Water Resources, and by the Reno and Susanville offices of the U.S. Soil Conservation Service.

Individual private and public landowners provided access to their property, information about their wells, and permission for drilling of test holes and installation of observation wells and precipitation gages on their property. In particular, the cooperation of Franklyn Jeans (Fish Springs Ranch), the U.S. Army (Sierra Army Depot), and the U.S. Bureau of Land Management is sincerely appreciated.

HYDROGEOLOGIC SETTING

Honey Lake Valley is similar to other valleys of the Great Basin in its physiography and its agricultural land and water use. Most development of the surface water and ground water has been in the western, less arid part of the study area, but ground water also is used to irrigate about 1,800 acres of alfalfa and pasture in the eastern part of the basin at Fish Springs Ranch. Geothermal ground water is used in the Wendel, Amedee, Litchfield, and Susanville areas (plate 1).

Structural Framework

Honey Lake Valley lies at the junction of three geologic and physiographic provinces (plate 1): (1) the northeastern edge of the Sierra Nevada mountain range, (2) the western edge of the Basin and Range Province, characterized by elongate basins surrounded by elevated mountain blocks, and (3) the southeastern edge of the Modoc Plateau, characterized by volcanic cones surrounded by relatively flat-lying volcanic flows (plate 2). A regional fault system, the Walker Lane, extends from Las Vegas through the Walker Lake area and into the study area (plate 2). The fault system has mainly right-lateral offsets similar, and parallel, to those of the San Andreas fault system in California (Bonham, 1969, p. 45).

The topography of the basin was produced primarily by movement along several faults and fault zones of the Walker Lane. This movement began in the middle Miocene time, about 12 million years ago (Bonham, 1969, p. 45), and continues at present. Volcanism, erosion, and sedimentation also have shaped the landscape. The fault zones shown on plate 2 include the Honey Lake and Warm Springs fault zones along the southwest boundary of the basin, and the Antelope Mountains and Eagle-Honey fault zones on the northwest side of the valley floor (Roberts, 1985, p. 43).

Movement along the fault zones has been vertical, accompanied by right-lateral slip (Roberts, 1985, p. 42). Lateral movement is notable on the Warm Springs fault zone, where the estimated horizontal offset has been at least 3.5 miles (Grose, 1984). Movement has been mainly vertical on the Antelope Mountains and Eagle-Honey fault zones, although some evidence suggests strike-slip offset (Roberts, 1985, p. 45).

Description of Hydrogeologic Units

Principal geologic units in Honey Lake Valley are granitic bedrock, volcanic rocks, and unconsolidated to semiconsolidated sediments. The relative ages, thicknesses, and hydrologic properties of the geologic units are summarized in table 1 and their distribution is shown on plate 2. The levels of geologic detail in the five geologic maps used to compile plate 2 (Lydon and others, 1960; Burnett and Jennings, 1962; Bonham, 1969; Grose, 1984; and Grose and others, 1989) are different enough to prevent reconciliation of across-the-join discrepancies without further field mapping (which was beyond the scope of this project).

Granitic Bedrock

Relatively impermeable bedrock forms a lower boundary to most ground-water flow within the basin. Although granitic bedrock is exposed mainly on the southwest and south sides of the basin (in the Diamond Mountains of the Sierra Nevada and in the Fort Sage Mountains), it also crops out in the Virginia Mountains (plate 2). Geothermal studies indicate that some ground water moves along fault zones in the bedrock to sustain hot springs near Wendel and Amedee (Juncal and Bohm, 1987).

Movement along fault zones has displaced the granitic bedrock downward on the northeast side of the basin to depths greater than 5,000 feet below land surface. Granitic rocks have been found 5,000 feet below land surface north of Honey Lake near Wendel, in holes drilled for energy exploration (plate 3A). Interpretation of seismic-refraction data (Fuis and others, 1987, p. 57) and telluric electrical-resistivity soundings (Pierce and Hoover, 1988, p. 4) show that bedrock (perhaps granitic) underlies the northern part of the basin at depths of 5,000-6,000 feet below land surface north of Eagle Lake and in Secret Valley.

Volcanic Rocks

Volcanic rocks (plate 2) ranging in age from Oligocene to Miocene, which have been dated from about 35 million to 12 million years, overlie the granitic rocks in the Sierra Nevada, the Fort Sage Mountains, and the Virginia Mountains (Bonham, 1969, p. 23 and 28). In the Virginia Mountains, their thickness increases toward the north to more than 1,500 feet beneath the basin fill near Fish Springs Ranch (Pierce and Hoover, 1988, p. 4). The thickness of the volcanic rocks also increases toward the east to about 7,000 feet beneath the crest of the Virginia Range and to more than 4,000 feet in the adjacent Pyramid Lake Valley (Bonham, 1969, p. 45).

TABLE 1.--*Characteristics of principal hydrogeologic units* [1]

Era	Period	Epoch	Unit	Approximate thickness (feet)	Lithology	Occurrence	General hydrologic properties
Cenozoic	Quaternary	Pleistocene and Holocene	Unconsolidated basin-fill deposits	0-700	Alluvial gravel, sand, and silt with some clay lenses; poorly to well sorted.	Forms perimeter of basin-fill deposits.	Moderate to high hydraulic conductivity at perimeter of basin and in northwest and southwest corners of basin where sand units are thickest; low hydraulic conductivity in center and eastern side of basin.
					Sand, silt and clay of pluvial, fluvial, and deltaic deposition; includes nearshore and offshore sediments of Pleistocene Lake Lahontan, and Holocene sediments.	Surficial unit beneath central valley floor.	
		Pleistocene	Volcanic rocks	0-300	Basalt flows, jointed, with scoriaceous tops and bottoms, dense interiors.	West, north, and east of Susanville and in low foothills.	Avenue of recharge to basin-fill aquifers in northwest part of Honey Lake Valley.
	Tertiary	Pliocene	Volcanic rocks	0-300	Basalt flows, jointed, with scoriaceous upper and lower flow surfaces.	Surrounds northern mountain blocks from Willow Creek Valley to area east of Skedaddle Mountains	Moderate to high hydraulic conductivity. Avenue of recharge to basin-fill aquifers from northern mountain blocks.
			Sedimentary and pyroclastic deposits	0-5,000	Tuffaceous silt, clay, diatomite, and sand; also includes pyroclastic air-fall and water-laid volcanic tuffs; semiconsolidated.	Forms floor of Secret Valley; also exposed in small outcrops southeast of Honey Lake; probably constitutes principal basin-fill deposit at depth.	Generally low hydraulic conductivity. Could transmit recharge to basin-fill aquifers from northern mountain blocks.
		Oligocene to Miocene	Volcanic rocks	0-5,000 in north; 0-7,000 in Virginia Mountains	Basalt, andesite, and rhyolite flows and flow breccias, jointed and fractured; commonly with scoriaceous upper and lower flow surfaces.	Forms high north and northeast mountain blocks, and caps granitic rocks on south and southwest mountain blocks.	Moderate to high hydraulic conductivity. Avenue of recharge to basin-fill aquifers from north and northeast mountain blocks.
Mesozoic	Cretaceous		Granitic bedrock	--	Massive granodiorite, locally weathered and decomposed.	Forms southwest mountain blocks; small exposures in Virginia Mountains; granitic bedrock may be present beneath much or all of basin.	Generally impermeable.

[1] Distribution of hydrogeologic units are shown on plate 2.

In the low pass that separates the Fort Sage Mountains from the Virginia Mountains, the rocks are rhyolitic ash-flow tuffs and volcanic-flow breccias that have been dated by the potassium-argon method as 30 million to 23 million years old (Grose, 1984). These volcanic rocks are relatively impermeable to ground-water flow except where fractured.

In the Virginia Mountains, the volcanic rocks are composed of basaltic and andesitic flows. They are included in the Pyramid sequence by Bonham (1969, p. 29) and are shown by Grose (1984) to overlie the older rhyolites and flow breccias to the west. They have been dated by the potassium-argon method as ranging from 15 million to 12 million years. These rocks are the main water-bearing unit near Fish Springs Ranch, forming an avenue for recharge to the basin-fill deposits of Honey Lake Valley.

The northern tip of the Virginia Mountains near Astor pass is composed of volcanic rocks that are more massive than those found near Fish Springs Ranch (Larry J. Garside, Nevada Bureau of Mines and Geology, oral communication, 1989). These massive rocks are considered to be less permeable to ground-water flow than the more fractured volcanic rocks near Cottonwood Canyon to the south.

Volcanic rocks (plate 2) in the north part of the basin, and on the east side north of Astor Pass, range in age from about 12 million years (Miocene) to 1 million years (Pleistocene; Grose and others, 1989). These rocks are volcanic flows of the Modoc Plateau. The plateau is characterized by small- to medium-size eroded volcanoes surrounded by relatively flat-lying basalt and andesite flows. The rocks differ greatly in thickness, with layered sections hundreds to thousands of feet thick consisting of individual flows 10-30 feet thick. Pleistocene-age rocks near Susanville and Pliocene-age rocks, which occur as flat-lying, layered flows surrounding Shaffer and Skedaddle Mountains, are as thick as 300 feet (Roberts, 1985, p. 25-26). Miocene-age rocks are more than 5,000 feet thick near volcanic centers such as those in the Skedaddle Mountains (Diggles and others, 1988, p. C15). The rocks are highly permeable in Willow Creek Valley in the northwest part of the study area and are an important source of water to wells along the entire north side of the basin. These rocks form an avenue for southward ground-water flow toward the floor of Honey Lake Valley from Willow and Secret Valleys.

Unconsolidated and Semiconsolidated Sediments

Basin-fill deposits of Pliocene to Holocene age, consisting of unconsolidated and semiconsolidated sediments and pyroclastic volcanic rocks, partly fill the structural depression underlying Honey Lake Valley. Generally fine-grained sedimentary and pyroclastic volcanic deposits of Pliocene age underlie, interfinger with, and overlie the consolidated volcanic rocks along the entire north and northeast margins of the basin. These semiconsolidated deposits consist of thick layers of volcanic tuff and ash that typically were deposited in shallow lakes along with lacustrine and fluvial deposits of clay, silt, and minor amounts of sand. Logs of geothermal wells in the Wendel area indicate that a basal conglomerate unit about 500 feet thick overlies bedrock. The semiconsolidated unit is called a lake deposit by Lydon and others (1960) and by the California Department of Water Resources (1963a, table 1), but is described as a volcanic tuff by Grose and others (1989). The most significant hydrologic characteristic of the unit is its low permeability. For this study, the fine-grained sedimentary and pyroclastic deposits are treated as a single unit. The unit is present over most of the study area northeast of the Sierra Nevada and is exposed on the floor of Secret Valley and at a few places in the center of Honey Lake Valley (plate 2). Most of the basin-fill material consists of this unit.

Honey Lake occupies part of an area previously covered by a much larger, prehistoric water body known as Lake Lahontan. During the late Pleistocene time, between 14,000 and 12,500 years ago, Lake Lahontan inundated as much as 8,600 mi^2 in northwestern Nevada and in the California part of Honey Lake Valley (Benson and Mifflin, 1986, p. 1). The water level in the huge lake attained a maximum altitude of about 4,365 feet above sea level, almost 400 feet above the present-day level of Honey Lake. Quaternary-age sediments deposited in Lake Lahontan are an important aquifer in the northwestern and southern parts of the study area where sands and gravels from the Susan River and Long Valley Creek dominate. On the eastern side of the basin, however, the deposits consist mostly of fine-grained silt and clay that have low hydraulic conductivity.

Alluvial fans of Quaternary age, consisting of poorly sorted deposits ranging in size from clay to boulders, have accumulated along the base of the mountain fronts and interfinger with the dominantly fine-grained lake deposits toward the center of the basin. These alluvial sediments have moderate to high permeability and are an important aquifer at the edge of the valley floor.

Thickness of Geologic Units

The volume of ground water in storage in the study area depends in part on the thickness of sedimentary and volcanic rocks overlying the granitic bedrock. Plate 3 and figure 2 show the estimated thicknesses of volcanic rocks and sedimentary units, and the total depth to bedrock beneath the eastern part of the basin. Because of similar hydrogeologic properties, the unconsolidated deposits of Quaternary age and the semiconsolidated sedimentary and pyroclastic deposits are shown as one basin-fill sedimentary unit on plate 3, and the volcanic rocks of Oligocene to Pliocene age are shown as one volcanic unit on plate 3 and figure 2. Types of data that were collected for this study and used to compile the maps and geologic sections include existing surface geologic maps, gravity and aeromagnetic surveys, telluric electrical-resistivity soundings, seismic-refraction profiles, and lithologic logs from deep wells penetrating more than about 20 feet of granitic bedrock.

Gravity measurements at about 700 stations were obtained from the data base of the National Oceanic and Atmospheric Administration (written communication, 1987). Aeromagnetic data are from the U.S. Department of Energy (1983a, 1983b). About 50 additional gravity and magnetic stations were established during this study to fill areal gaps in the existing data, mainly along and east of the California-Nevada State line.

Computer programs developed by the U.S. Geological Survey (Cordell, 1970; Webring, 1985) were used to estimate the thickness of geologic units from the gravity and magnetic measurements. Results of these programs matched observed depths to granitic bedrock in drill-holes on the southwest side of the basin. In contrast, computer results were inconsistent with observed depths to granitic bedrock in drill holes on the northeast side of the basin near Wendel. To provide more information in the northeast area, telluric electrical-resistivity soundings and seismic-refraction profiles were made. Results of the telluric soundings are summarized by Pierce and Hoover (1988) and are shown on plate 3 along with results of the seismic-refraction profiles.

Plate 3A shows that the depth to granitic bedrock increases greatly toward the north and is more than 5,000 feet near Wendel. Lines of equal depth to bedrock are parallel to the strike of the Honey Lake and Warm Springs fault zones (plate 2) and indicate down-dropping of the bedrock, generally to the northeast.

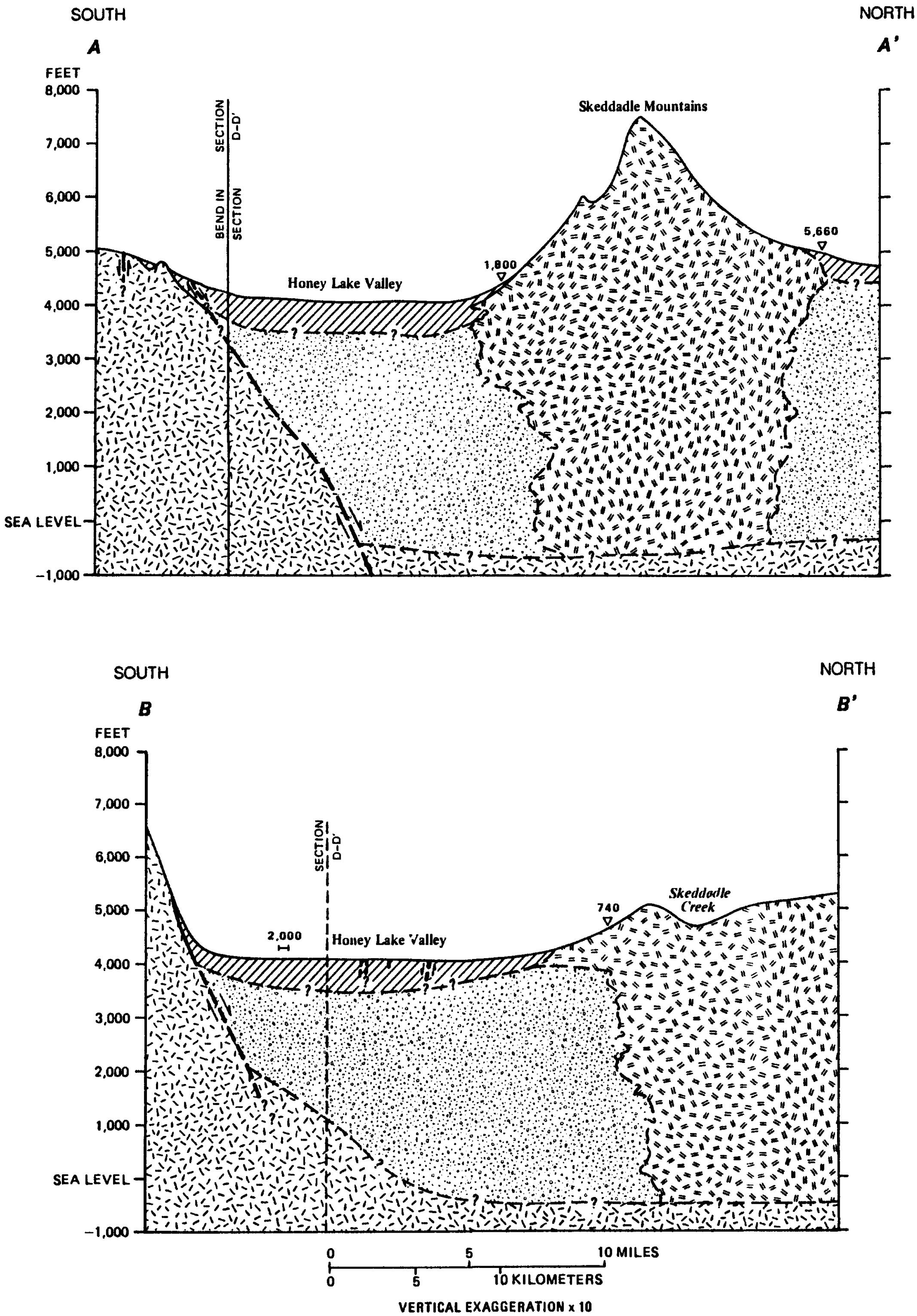
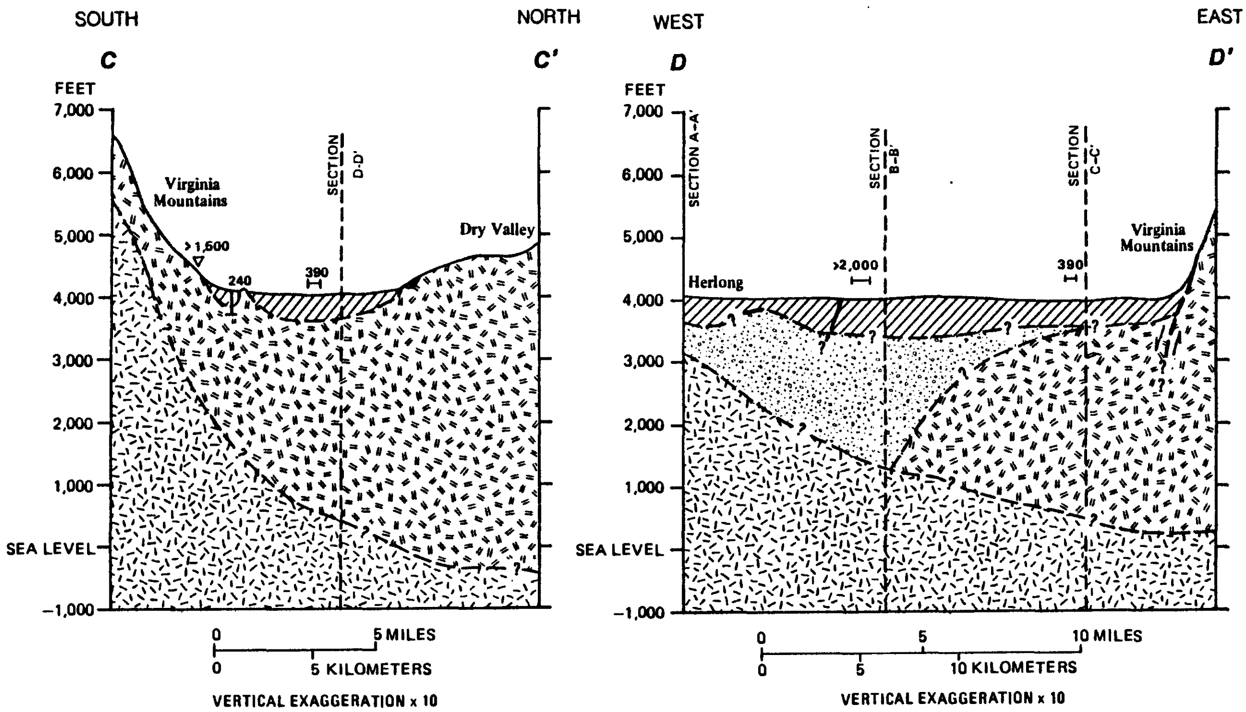


FIGURE 2.--Generalized geologic sections showing the estimated thickness of sedimentary deposits and volcanic rocks in eastern Honey Lake Valley. Lines of section are on plate 2.



EXPLANATION

HOLOCENE AND PLEISTOCENE BASIN-FILL SEDIMENTARY DEPOSITS

PLIOCENE SEDIMENTARY AND PYROCLASTIC DEPOSITS

MIOCENE TO OLIGOCENE VOLCANIC ROCKS

CRETACEOUS GRANITIC BEDROCK

— — ? — — CONTACT- - Queried where uncertain

? ? FAULT - - Dashed where inferred, queried where uncertain. Arrows indicate relative direction of movement

1,800 TELLURIC SOUNDING - - Number indicates thickness of volcanic unit, in feet

>2,000 SEISMIC SOUNDING - - Number indicates thickness of sedimentary unit, in feet

240 WELL - - Number indicates thickness of sedimentary unit, in feet. Total depth of well is 246 feet

FIGURE 2.--Continued.

Plate 3B shows that the thickness of unconsolidated and semiconsolidated basin-fill deposits of Holocene to Pliocene age increases toward the north with increasing depth to bedrock, and decreases toward the east where volcanic rocks make up the largest part of the geologic section overlying bedrock. In the deeper parts of the basin, most of the total thickness probably is fine-grained Pliocene lake deposits (California Department of Water Resources, 1963a, p. 207), except along the north margin of the basin where pyroclastic tuffs or ash flows predominate.

Plate 3C shows the thickness of volcanic rocks of Oligocene to Pliocene age that constitute aquifers on the north and east sides of the basin. The unit forms a wedge on the extreme east side of the basin, increasing from less than 1,000 feet thick near the head of Cottonwood Creek to more than 4,000 feet thick beneath Dry Valley north of Flanigan. Volcanic rocks are absent in drill holes completed to granitic bedrock beneath The Island area (south of Honey Lake) and the area southeast of Standish, and are present as only thin units near Wendel. A drill hole north of Wendel penetrated volcanic rocks in the upper 2,500 feet of basin fill (GeoProducts Corporation, 1982, p. 17; 1984, p. 22), and telluric soundings near Viewland and north of Herlong Siding indicate volcanic rocks overlie tuffaceous or sedimentary rocks (Pierce and Hoover, 1988, p. 6 and figure 8). Although data are absent beneath the Skedaddle Mountains, Diggles and others (1988, p. C15) believe extensive thicknesses of sedimentary layers do not exist there.

The generalized geologic sections in figure 2 show the estimated thicknesses of sedimentary and volcanic rock units overlying bedrock along profiles in the eastern part of Honey Lake Valley. Because the estimated thicknesses are uncertain due to lack of data points in some areas, the profiles in figure 2 should be considered approximations, rather than exact depictions, of the subsurface geology.

Hydrologic Features

Surface Water

More than 40 streams flow from the Diamond, Fort Sage, and Virginia Mountains and the northern volcanic uplands toward the center of the topographically closed basin. Most are intermittent and reach the valley floor only in wet years. The largest streams in the study area are the Susan River, Willow Creek, Long Valley Creek, and Gold Run Creek (plate 1). The most prominent surface-water feature in the basin is Honey Lake, which fluctuates greatly in area and volume, but on average has a surface area of about 47,000 acres and contains about 120,000 acre-ft of water, derived from a combination of lake-surface precipitation, stream inflow (mostly from the Susan River), and ground-water inflow. Water accumulates in Honey Lake during periods of rapid snowmelt, but most streamflow is diverted or seeps into alluvial-fan deposits before it reaches the valley floor and the lake. Surface-water conditions determined for this study are discussed more fully in a separate report by Rockwell (in press).

Ground Water

Ground water in Honey Lake Valley mainly originates as precipitation in the study area and in the drainage areas of the Susan River and Long Valley Creek. Precipitation infiltrates through unconsolidated deposits and faults and fractures in consolidated rocks to become ground water. Ground water flows downgradient from recharge areas in or near the mountains to discharge areas near the central axis of the basin.

Thermal water is found in several places in the basin, most notably in the Wendel and Amedee areas (plate 1). According to Juncal and Bohm (1987), the geothermal water is part of a flow system in fractured bedrock and is related to the Honey Lake range-front fault zone (plate 2, inset map) and the Walker Lane fault system. Recharge for the system is from precipitation in the Diamond Mountain range of the Sierra Nevada. Meteoric water infiltrates and circulates deeply in granitic bedrock beneath the valley floor. It is heated by above-average regional heat flow related to volcanism, and rises along the north-northeast-striking faults. Hot-spring locations might be controlled by the intersection of the north-striking and northwest-striking faults.

Water Quality

Analyses of about 500 surface- and ground-water samples from the Nevada Division of Health, the California Department of Water Resources, the Washoe County Department of Public Works, the Sierra Army Depot, and several published reports (Hilton, 1963; Rush and Glancy, 1967; Clawson, 1968; William F. Guyton Associates, 1987) indicate that the quality of water in much of Honey Lake Valley is suitable for irrigation, stock watering, industrial, commercial, and domestic uses. In the eastern part of the basin, calcium, sodium, and bicarbonate ions predominate in streams fed by mountain springs, sodium and bicarbonate ions predominate in most ground-water samples, and the dissolved-solids concentrations are low, generally less than 500 milligrams per liter. In the central part of the basin, sodium and chloride ions predominate and dissolved-solids concentrations are higher. Geothermal areas also are characterized by high dissolved-solids concentrations, dominated by sodium and sulfate ions. Areas in the basin where ground water contains elevated concentrations of dissolved solids, boron, fluoride, and nitrate have been delineated by the California Department of Water Resources (1963b, plate 32). Water from thermal springs at Amedee and Wendel, and from several wells near Standish and elsewhere in Honey Lake Valley, contain elevated concentrations of arsenic (Wormald, 1970).

In general, the dissolved-solids concentration in ground water increases with depth and with distance from the recharge area because longer flow paths allow more contact with soluble minerals of the aquifer. In the central parts of topographically closed basins, such as Honey Lake Valley, deep water moves upward under artesian pressure into shallower aquifers and continues to dissolve minerals along its flow path. Concentrations of dissolved solids in water in shallow aquifers are increased further by evapotranspiration near the surface. Thus, concentrations of dissolved solids in water in the upper parts of aquifers in some discharge areas (along the central axis of the basin, including Honey Lake and the playa areas) may decrease with depth. Actual flow paths are more complicated than indicated by this simple concept and involve recirculation and mixing of water from different source areas due to density differences caused by differences in temperature or chemical concentrations.

GROUND-WATER FLOW SYSTEM

Aquifers and the water that is stored in and moves through them constitute the ground-water flow system. Analysis of the flow system includes assessment of the hydraulic characteristics of aquifer materials, quantification of components of the ground-water budget (recharge, storage, and discharge), and evaluation of rates and directions of flow.

Aquifer Properties

Properties of aquifer materials that control storage and movement of ground water are hydraulic conductivity (the capacity of rocks and sediments to transmit water) and specific storage and yield (the amount of water that is stored in and released by aquifers in response to changes in head).

Hydraulic Conductivity

Hydraulic conductivity is expressed as the volume of water at the existing kinematic viscosity that will move in a unit time, under a unit hydraulic gradient, through a unit area measured at right angles to the direction of flow. Coarse-grained, well-sorted sediments transmit water more readily than fine-grained or poorly sorted sediments, and layers of fine-grained material impede vertical flow. Flow through consolidated rock depends on the size, distribution, orientation, and interconnection of fractures and other openings. Therefore, hydraulic conductivity differs for different materials and from place to place within an aquifer.

Horizontal Hydraulic Conductivity

Horizontal hydraulic conductivity of basin-fill deposits and volcanic rocks was estimated at well sites for which drillers' logs, geologists' logs, or aquifer-test data were available. Drillers' logs were available for 140 wells in the basin, and 36 of these also had reports on production tests. Hydraulic conductivities were calculated by two methods for the 36 sites:

1. Specific-capacity data from the production tests were used to estimate hydraulic conductivity by the Theis (1963) method.
2. Lithologic descriptions were used to estimate hydraulic conductivity by a method described by Maurer (1986, p. 27-28) and referred to in this report as the "Percent Coarse" method. For this method, the percentage of coarse-grained materials in the aquifer section penetrated by a well was estimated from the driller's or geologist's log. Examples of coarse and fine lithologic designations are listed by Plume (1989, p. A10).

Hydraulic conductivity for the Percent Coarse method was assumed to be 1 ft/d for fine-grained materials, and 20 ft/d for coarse-grained materials, on the basis of typical values for silt to fine sand and for coarse sand and porous lava flows (Heath, 1983, p. 13).

Results of the Theis and Percent Coarse methods of estimating hydraulic conductivity were used in regression and correlation analyses to determine whether a relation between the two methods could be shown. The Theis method is considered to be more accurate because it is based on actual performance of a well during pumping, but it can be applied only where production-test data are available. If the methods relate well, more accurate hydraulic conductivities can be estimated at numerous sites where only drillers' logs are available.

Comparison of the results indicates that the median hydraulic conductivity for the 36 sites estimated by both methods was about 8 ft/d and that equal numbers of values were above and below the line of equality (the line that defines hydraulic-conductivity values from both methods as being equal). However, hydraulic conductivities estimated by both methods for the same site generally were not in agreement. Correlation was poor (correlation coefficient 0.04) probably because (1) specific capacities were based on short-duration production tests (generally much less than 1 day) made to verify yields of new wells, rather than to determine aquifer characteristics; (2) wells may not have been fully developed at the time of testing; (3) descriptions of the same geologic materials by different drillers probably differ; (4) information on packing and sorting of sediments, which also affect hydraulic conductivity, was not available; and (5) the range of possible values resulting from the Percent Coarse method was restricted. Because correlation of the Theis and Percent Coarse methods is poor, the more accurate Theis method was used to estimate hydraulic conductivities wherever possible and additional factors, as discussed in the following paragraphs, were considered in other places.

To estimate aquifer characteristics for areas where well data are sparse or unavailable, another relation was considered. Hydraulic conductivity of basin-fill deposits is a function of their source (volcanic or granitic rocks) and depositional environment (for example, alluvial fan, nearshore in a lake, and playa). In general, the upstream parts of alluvial fans at higher altitudes primarily contain coarse sands and gravels, whereas sediments beneath the valley floor are dominated by fine sands, silts, and clays. Therefore, grain size, and the corresponding hydraulic conductivity, can be related to land-surface altitude. The percentages of coarse-grained materials described in drillers' logs for 88 wells in the basin are grouped by range of land-surface altitude and shown in figure 3. The median of 18 sites where land-surface altitude is at or above 4,140 feet (the highest altitude group) was 56 percent coarse-grained materials, whereas the median of 24 sites below 4,000 feet (the lowest altitude group) was 29 percent. Most wells and most water use in the basin are near 4,000 feet. At altitudes below this, playa deposits produce smaller yields and water quality and soils are unacceptable for agricultural use; at much higher altitudes, the depth to water is greater and soils tend to be thin and rocky.

Aquifer tests were made at two wells at the California Correctional Center north of Lake Leavitt (California Department of Water Resources, 1988), at five wells in the southeast part of the basin (William F. Guyton Associates, 1987, p. 12; Michael Widmer, Washoe County Utility Division, written communication, 1989), and at two wells in the southeast as part of this study. The results indicate that hydraulic conductivities of aquifers in the southeastern part of the basin ranged from about 10 ft/d for basin-fill deposits to greater than 100 ft/d for fractured volcanic rocks. The hydraulic conductivities of basin-fill sediments may differ over short distances. The hydraulic conductivity of fractured volcanic rocks also differs, but generally is greater than that of other aquifer materials in the southeastern part of the study area. No aquifer tests are available from wells in granitic bedrock, which is relatively massive and considered capable of yielding only small quantities of water in Honey Lake Valley (California Department of Water Resources, 1963a, p. 29).

The median horizontal hydraulic conductivity of aquifer materials throughout Honey Lake Valley is estimated to be 8 ft/d, on the basis of analyses of production tests and descriptions of geologic materials. In general, the hydraulic conductivity of unconsolidated sediments decreases with decreasing altitude from a maximum on upper alluvial fans to a minimum in sediments underlying the playa on the valley floor. It also decreases with depth as a result of compaction. The hydraulic conductivity of volcanic rocks differs areally and with depth, and depends on the number, spacing, and degree of connection of fractures. The greatest hydraulic conductivities are in fractured volcanic rocks in the southeastern part of the basin; the smallest are probably in massive granitic bedrock.

Vertical Hydraulic Conductivity

Unconsolidated basin-fill sediments are vertically anisotropic because of layering and compaction. Freeze and Cherry (1979, p. 32, 34) report that the ratio of horizontal to vertical anisotropy in individual core samples is less than 10 to 1, due to orientation of clay minerals, and that a typical regional ratio of horizontal to vertical anisotropy is 100 to 1 or greater, due to layering of sediments. The 100 to 1 ratio is considered representative of sediments in basins similar to Honey Lake Valley (Morgan, 1988, p. 18) and is probably reasonable for unconsolidated aquifers in the basin.

Volcanic rocks also may be vertically anisotropic because flows are layered; some layers or parts of layers are more welded, dense, vesicular, or fractured than others. Ground-water flow through volcanic rocks is predominantly horizontal, but water also moves vertically through joints and fractures. The ratio of horizontal to vertical hydraulic conductivity of basalts in other areas has been estimated as being greater than 100 to 1 (John J. Vaccaro, U.S. Geological Survey, written communication, 1989).

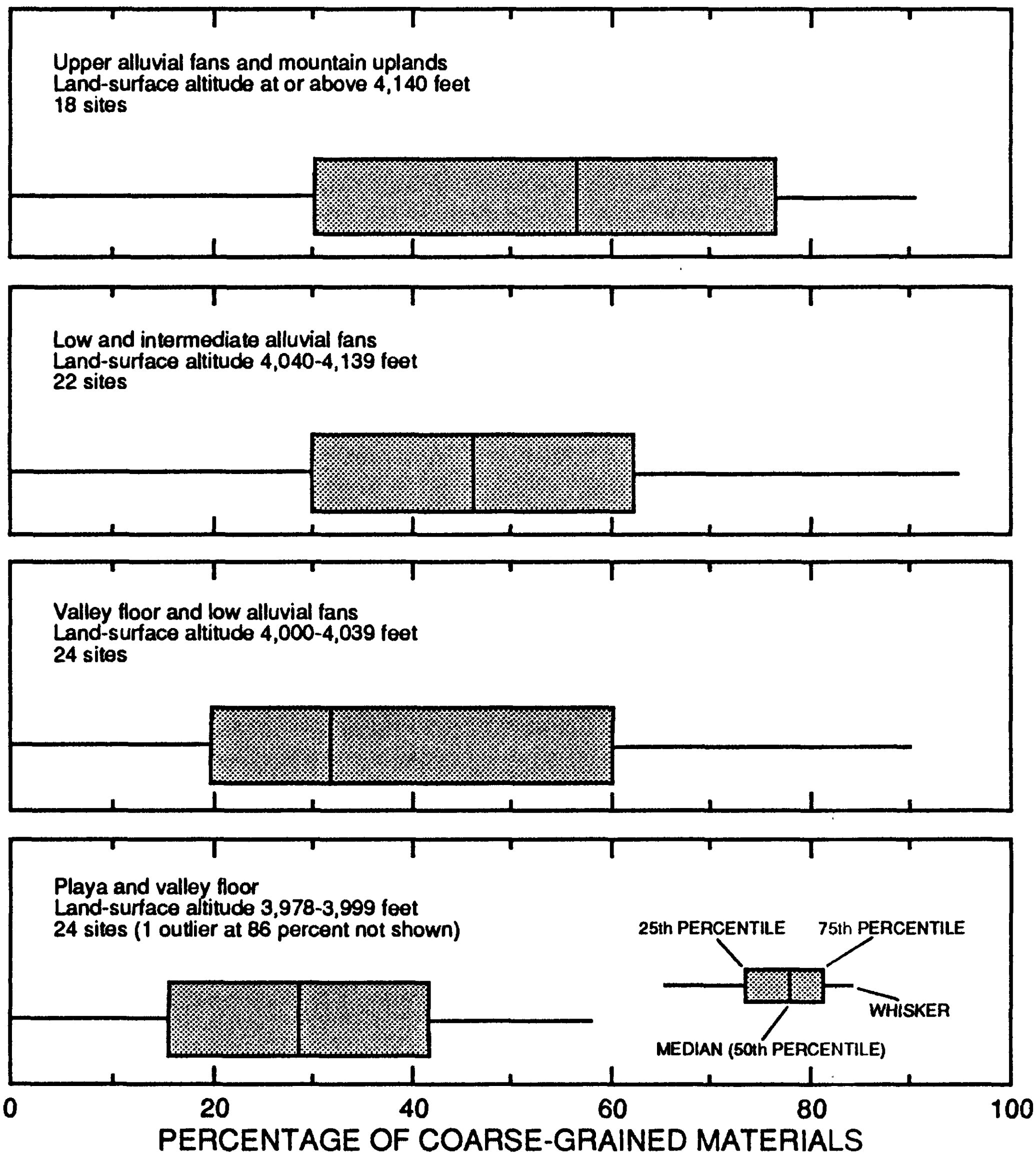


FIGURE 3.--Percentage of coarse-grained materials at different altitudes, estimated from lithologic logs for 88 well sites.

Storage Coefficient and Specific Yield

The storage of water in an aquifer is expressed in terms of a storage coefficient or specific yield. Storage coefficient is the volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. Specific yield is the ratio of the volume of water that will drain under the influence of gravity, to the total volume of saturated aquifer material from which the water drains. For confined aquifers, the water released from storage comes from expansion of the water and compression of the aquifer. For unconfined aquifers, the water released from storage comes from drainage of the sediments by gravity, and the storage coefficient is equivalent to the specific yield. The storage coefficient of unconfined aquifers is 100 to 10,000 times greater than the storage coefficient of confined aquifers because of these differences (Heath, 1983, p. 29).

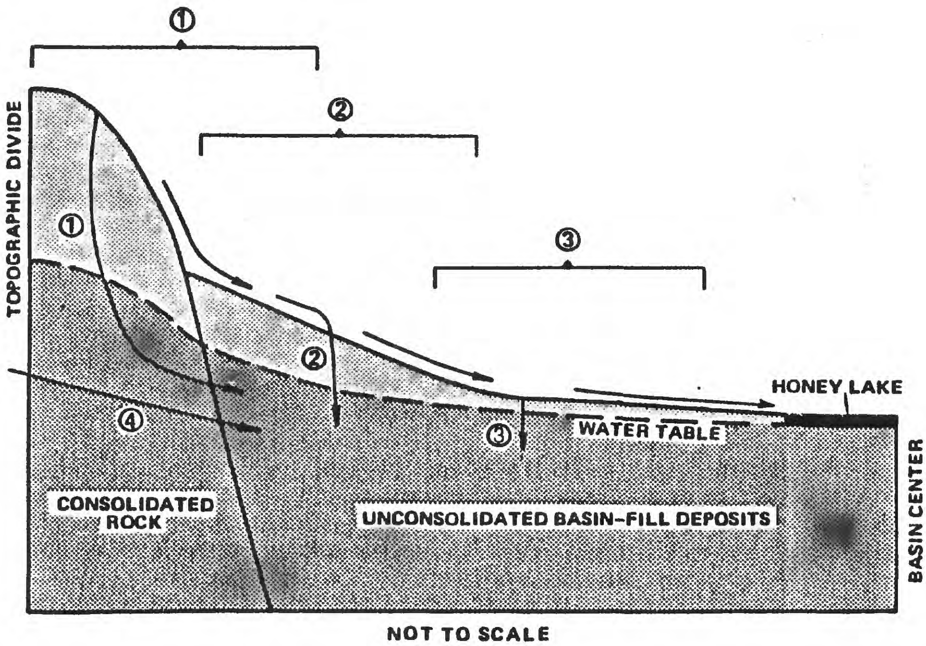
The specific yield of unconfined basin-fill deposits (assumed for this study to be the upper 100 feet of saturated sediments in Honey Lake Valley) was estimated from aquifer tests and from lithologic logs and reported values for different types of geologic materials. Specific yield, which depends on grain size, sorting, and porosity, ranges from about 2 percent for clay to nearly 30 percent for uniform coarse sand (Johnson, 1967, p. 70). A specific yield of 15 percent has been used as the average for ground-water flow models of other valleys in the Great Basin (Thomas and others, 1989, p. 14), and can be considered representative for primarily coarse-grained (upper-fan) deposits, included as perimeter deposits in table 1. A specific yield of 10 percent is typical of mixed coarse- and fine-grained deposits, and about 6 percent is typical for fine-grained deposits (Thomas and others, 1989, p. 14-15). The former correspond to nearshore deposits in Honey Lake Valley and the latter correspond to offshore deposits beneath the central valley floor (table 1).

Storage coefficients for confined aquifers are directly proportional to the saturated thickness of the aquifer, and can be estimated as 0.000001 per foot, times the thickness of the aquifer (Lohman, 1972, p. 53). The storage coefficient calculated from results of aquifer tests at the California Correctional Center (California Department of Water Resources, 1988) was 0.00012 for confined basin-fill deposits about 270 feet thick. Storage coefficients determined by aquifer tests in the southeast part of the basin ranged from 0.024 for semiconfined basin-fill deposits to 0.0005 for volcanic rocks (Michael Widmer, Washoe County Utility Division, written communication, 1989). Values for fractured volcanic rocks were more variable than those for sediments.

Recharge

Recharge to the ground-water system in Honey Lake Valley is from (1) direct infiltration of precipitation and snowmelt into consolidated rock and unconsolidated basin-fill deposits, (2) infiltration of water from streams, (3) seepage of irrigation water, and (4) subsurface inflow from adjacent areas. The major sources are direct infiltration of precipitation in upland areas and infiltration of streamflow in alluvial-fan areas (figure 4).

To ascertain how much water recharges the ground-water system, direct infiltration of precipitation was estimated using a Deep Percolation Model (Bauer and Vaccaro, 1987), and surface-water infiltration was estimated using streamflow data, measurements of variations in streamflow resulting from seepage along stream channels, and information on irrigation-water use.



EXPLANATION

SOURCES OF GROUND-WATER RECHARGE

① **Direct infiltration of precipitation and snowmelt in mountain and upper alluvial-fan areas**

② **Infiltration of streamflow through stream bottom in alluvial-fan areas**

③ **Infiltration of irrigation water from surface-water and ground-water sources in low alluvial-fan and valley-floor areas**

④ **Subsurface inflow from adjacent basins**

FIGURE 4.--Schematic cross section showing sources and relative amounts of ground-water recharge in Honey Lake Valley.

Infiltration of Precipitation

Most precipitation that falls on the basin evaporates or is transpired by vegetation before it infiltrates to the water table. The small part that does infiltrate is the major source of ground-water recharge.

Areal variations in recharge to aquifers in Honey Lake Valley result in large part from differences in the areal distribution of precipitation and streamflow. Precipitation is much greater in the mountains than on the valley floor, ranging from more than 20 inches in the Diamond Mountains in the Sierra Nevada, to less than 8 inches over large areas of the valley floor (figure 5). Mean annual precipitation over the study area, estimated from amounts shown in figure 5, is about 1.1 million acre-ft. The areal distribution determined for this study is the same as that shown by the California Department of Water Resources (1963b, plate 2), but the mean annual amounts are estimated to be 2 inches greater everywhere, on the basis of 23 additional years (1963-88) of recorded observations since the original map was compiled. Mean annual precipitation at each of seven precipitation-measurement sites in the basin was about 2 inches greater. The difference may be even greater at high altitudes, but no data are available to quantify the amount. Therefore, a conservative 2 inches was added to the entire distribution. The revised quantities of precipitation are shown in figure 5.

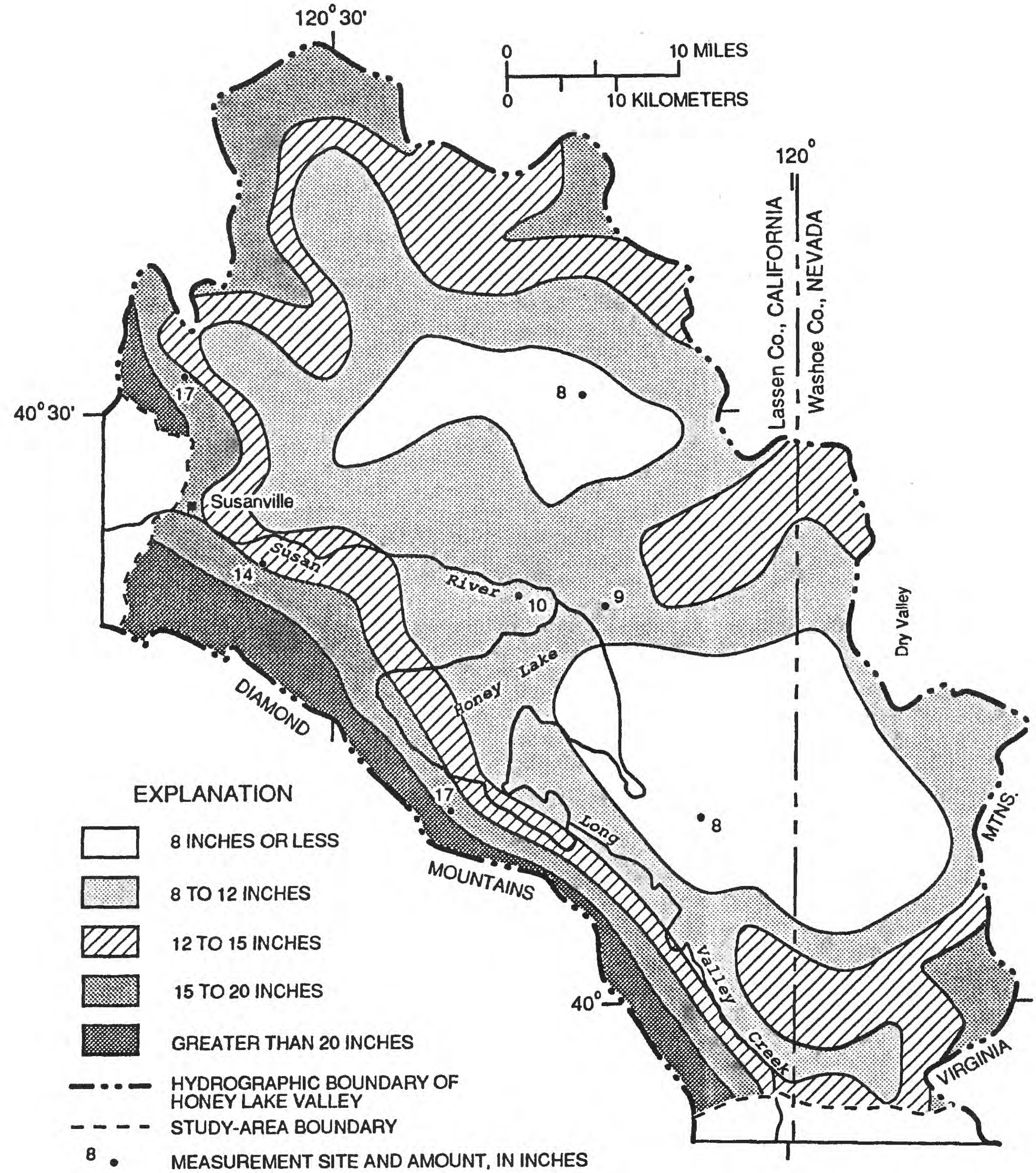


FIGURE 5.--Mean annual precipitation. Distribution from California Department of Water Resources (1963b, pl. 2); amounts from California Department of Water Resources data for 1958-86 (written communication, 1987).

Precipitation is recorded at many low-altitude sites in the basin, but little direct information is available for higher altitudes. To collect high-altitude data, three precipitation-storage gages were installed in September 1987 as part of this study, and a fourth gage was added by Washoe County in April 1988 (plate 1). Altitudes ranged from 5,260 to 7,770 feet. Although the periods of record through mid-1989 were too short to permit an accurate estimation of mean annual precipitation, the data clearly show that precipitation increases with altitude (table 2).

TABLE 2.--*Precipitation data for high-altitude storage gages in and adjacent to study area, 1987-89*

[--, no data available]

Precipitation site (plate 1)	Land-surface altitude (feet above sea level)	Cumulative precipitation for listed period (inches of water)	
		Sept. 1987 to April 1988	April 1988 to mid-1989[1]
1. Spanish Flat[2]	7,770	--	13.4
2. Shaffer Mountain[3]	6,660	5.6	--
3. Fort Sage Mountains	6,600	5.1	10.2
4. Skedaddle Mountains	5,260	2.4	6.4

[1] Site 1 visited in May 1989; sites 3 and 4 visited in April 1989.

[2] Washoe County gage, installed in April 1988.

[3] Destroyed by vandals between April and June 1988.

At the Susanville Airport (altitude, 4,148 feet), the climatological station in the basin for which the longest period of record is available, mean annual precipitation for the 30-year period 1951-80 was 14.3 inches (National Climatic Center, 1982). The rate of precipitation varied seasonally, from a mean of about 0.3 inch per month during July, August, and September to about 2.7 inches per month during December and January, as shown in figure 6. During 1931-88, total annual precipitation at the Susanville Airport ranged from 4.2 inches in 1949 to 28.7 inches in 1940 (figure 7).

Deep-percolation estimate.--Direct infiltration of precipitation was estimated by use of the Deep Percolation Model (DPM), a set of computer programs that uses long-term data on daily precipitation, daily temperature, soil characteristics, and vegetative cover to determine values for evapotranspiration (evaporation plus transpiration by plants), runoff, and recharge. The model, which calculates the areal distribution and average volume of water that percolates past the root zone to become recharge, was developed for use in eastern Washington, an arid to semiarid area similar in climate to the Honey Lake basin. The algorithms that the model uses to calculate potential evapotranspiration and surface runoff are directly applicable to the study area.

The model produces an estimate of recharge on the basis of physical processes; it uses data from the basin under investigation, and it calculates an energy and moisture budget. Hydrologic factors included in the model are precipitation; air temperature; solar radiation; interception of precipitation by plant foliage; accumulation, sublimation, and melting of snow; accumulation and evapotranspiration of soil moisture; and surface-water runoff.

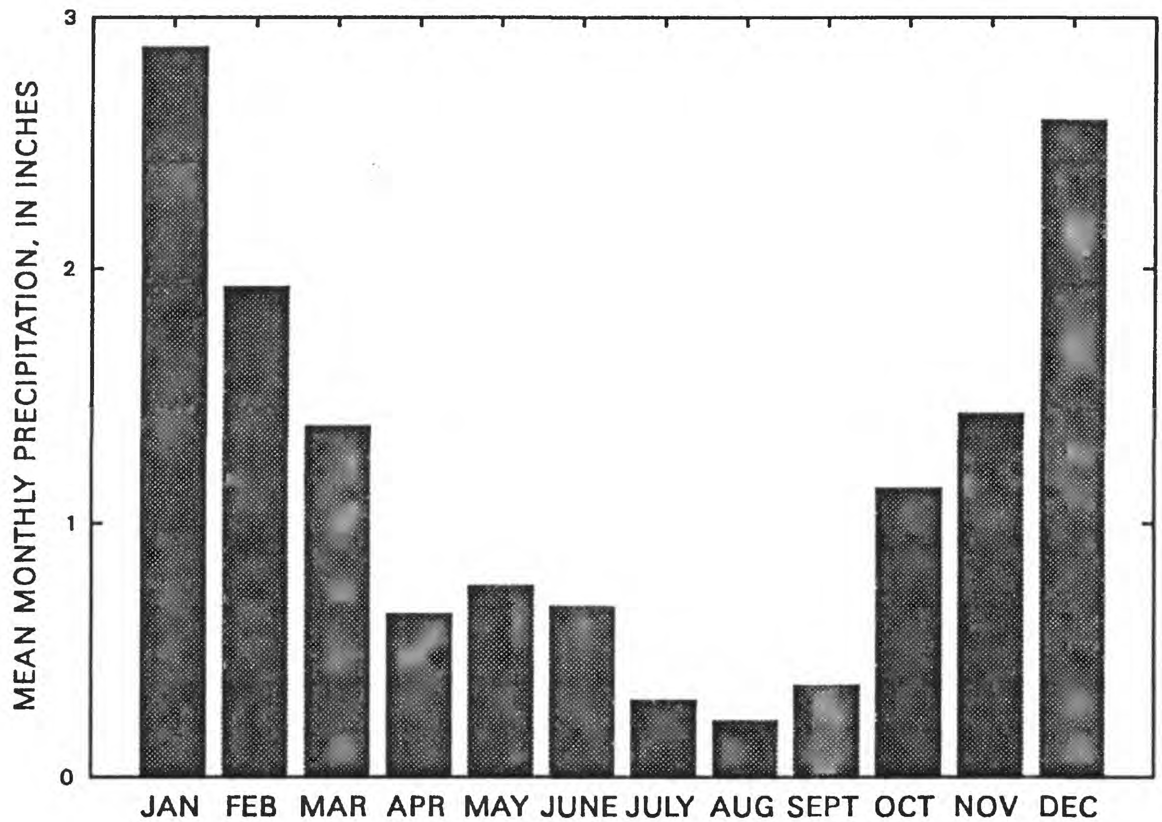


FIGURE 6.--Mean monthly precipitation at Susanville Airport, 1951-80 (data from National Climatic Center, 1982).

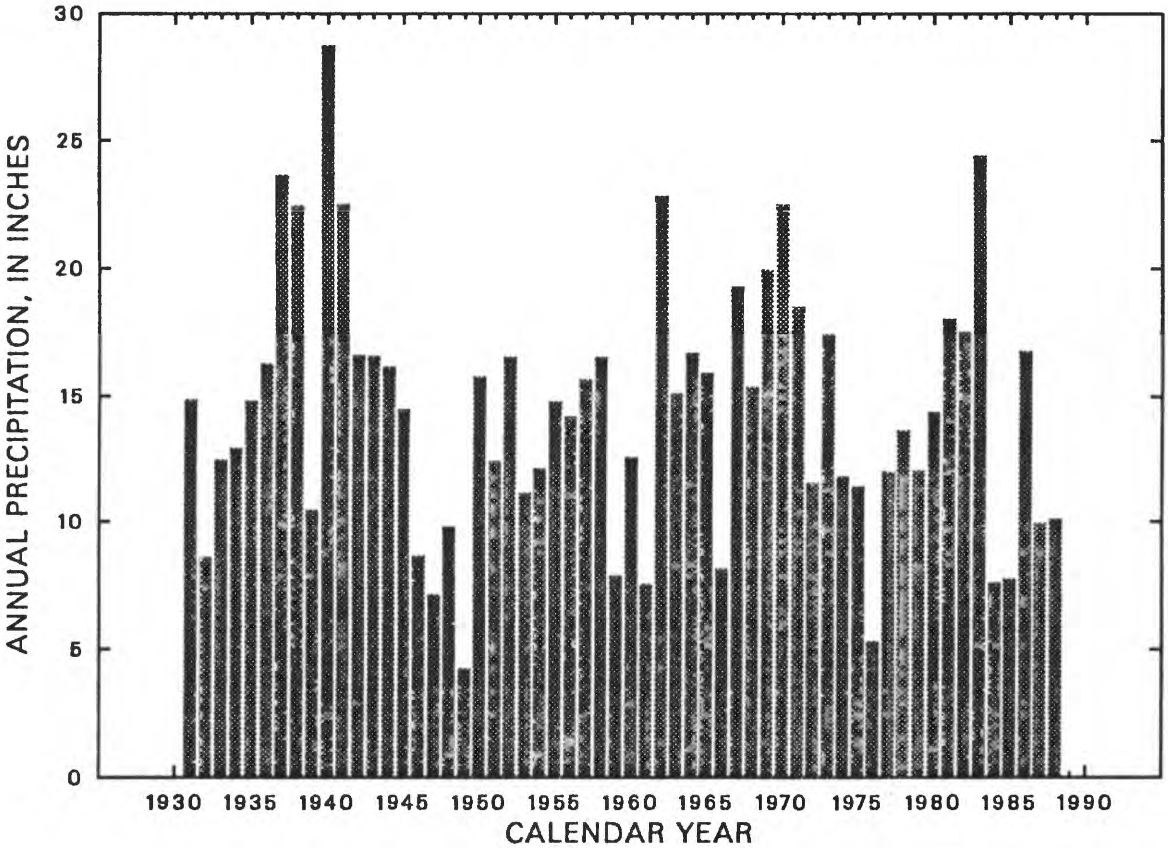


FIGURE 7.--Annual precipitation at Susanville Airport, 1931-88 (data from National Weather Service).

Limitations.--As with all models, the DPM can only approximate the complexity of a real system; simplifying assumptions must be made. For example, a single set of values is assumed to represent conditions over a heterogeneous tract of land (a model-grid cell). The DPM does a separate budget calculation for each grid cell and does not route runoff from one cell to another. It estimates only the recharge that is derived from precipitation on the cell area; recharge from seepage of streamflow or irrigation water is not included unless daily streamflow and irrigation values throughout the basin are supplied. This information is unavailable for most of Honey Lake Valley. The DPM does not consider the possible effects of the permeability of consolidated rock beneath the soil zone, or the contributions of ground water to soil moisture where the water table is shallow. Furthermore, it does not account for infiltration of runoff that may migrate to adjacent cells or for the delay of runoff when the temperature is above freezing but precipitation is temporarily stored in a residual snowpack. The methods, equations, and assumptions used to simulate processes in the DPM are described in detail by Bauer and Vaccaro (1987).

Method.--A north-south-oriented grid was superimposed on a map of the study area within Honey Lake Valley to divide the area into discrete cells (figure 8). The DPM grid for this study consists of 1,739 cells, each of 1 mi^2, including 74 inactive cells over Honey Lake. The model calculates daily, monthly, and annual energy and moisture budgets at each grid cell for the simulation period. It averages the monthly and annual values to determine mean monthly and long-term annual volumes of water for each component of the moisture budget, including water that percolates past the root zone to become recharge.

The energy-budget calculations are based on solar radiation and thermal energy at each grid cell. Energy values are determined from the daily maximum and minimum air temperatures at weather stations, and are adjusted for the distance to the grid cell from the station, the temperature lapse rate (vertical temperature gradient) in the region, and the altitude, slope, and aspect (land-slope orientation) of each grid cell. The model applies available energy to either snowmelt or potential evapotranspiration (the amount of evapotranspiration that would occur if unlimited water were available). Potential-evapotranspiration energy is used for: evaporation of moisture intercepted by the plant canopy, snow sublimation, soil evaporation, and plant transpiration. Excess energy becomes sensible heat or reflected radiation.

The daily moisture-budget calculations are based on incident precipitation at each grid cell. Daily precipitation recorded at weather stations is adjusted by the distance to the grid cell and weighted by the average annual precipitation at each grid cell. Depending on whether the daily temperature is below or above freezing, the moisture is added to either snowpack storage or interception storage in the plant canopy. Storage is reduced by sublimation from the snow pack or evaporation from the plant canopy, surface runoff is subtracted, and the remaining moisture is added to soil-moisture storage. Soil-evaporation and plant-transpiration rates for the soil and vegetation types specified in the cell are subtracted from the soil moisture for each soil layer. The remaining water (the part in excess of water that can be held in the soil or evapotranspired from it) is the deep percolation, or recharge.

Potential evapotranspiration is calculated using the Jensen-Haise method (Jensen, 1973), which is suitable for arid to semiarid climates. Surface runoff is calculated by the modified Soil Conservation Service (SCS) method of Wight and Neff (1983), which was developed for rangeland in eastern Montana. This method produced results that correlate well with measured runoff in areas that receive less than 10 inches of annual rainfall (John J. Vaccaro, U.S. Geological Survey, oral communication, 1989), similar to a large part of the study area (figure 5).

Data.--Information required for energy- and moisture-balance equations consists of three types: data on attributes of each cell, time-series data from specific weather stations, and basinwide data. Cell data include location (latitude and longitude), land-surface characteristics (altitude, slope, and aspect), soil characteristics (texture, available water capacity, and thickness), mean annual precipitation, and type of vegetative cover. Weather-station data consist of daily values of precipitation and maximum and minimum air temperatures, and average July maximum and minimum temperatures, for the 20-year simulation period (1961-80). General basinwide data are average monthly maximum and minimum temperature-lapse rates, snowmelt coefficient and sublimation rates, and minimum potential evapotranspiration rates.

Daily precipitation data for 1961-80 from 5 weather stations were used for this study (the National Weather Service site at the Susanville Airport and the California Department of Water Resources stations at Milford, Wendel, Fleming, and Willow Creek Valley; figure 8). Stations were selected on the basis of diversity of setting (location and altitude) and availability of long-term records. The annual precipitation at each grid cell was obtained by digitizing the precipitation map (figure 5) and using a gridding program (Webring, 1981) to calculate the value at each cell.

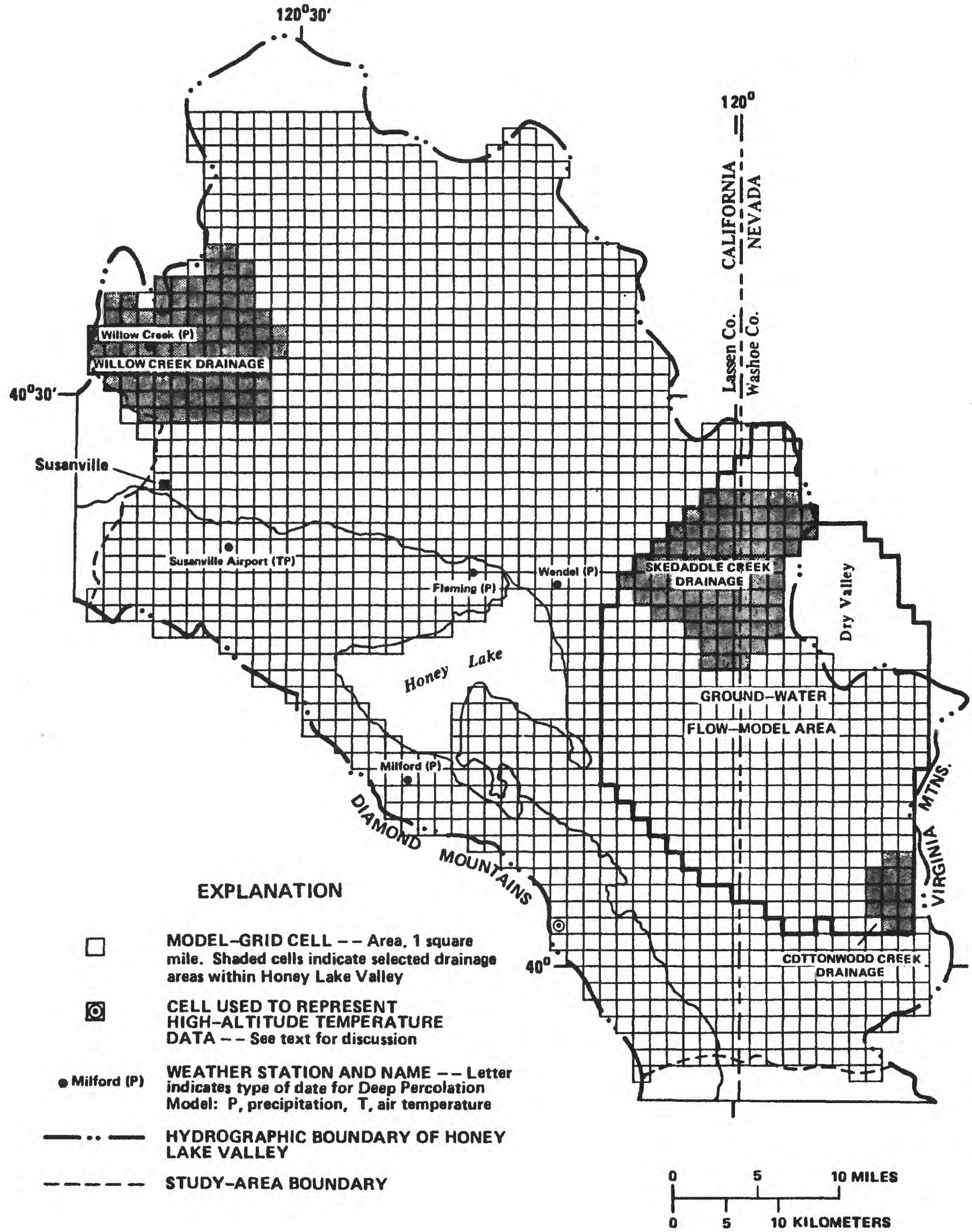


FIGURE 8.--Grid and meteorological sites used for the Deep Percolation Model.

Daily maximum and minimum air temperatures for the entire simulation period were available for only one weather station in the basin, the Susanville Airport station (figure 8). During 1951-80, air temperatures at Susanville Airport varied from -23 °F on February 1, 1956, to 103 °F on July 14, 1972. Air temperatures at Susanville Airport are summarized in figure 9. By use of temperature data from only the Susanville Station, the DPM produced unreasonable simulations of snowpack duration and recharge in comparison with observed snowpack and recharge periods. The temperatures at this relatively low-altitude (4,148 feet) station were too high to maintain a snowpack at the higher altitudes. Therefore, to represent daily maximum and minimum temperatures at a higher altitude, data from the Truckee Ranger Station, a Sierra Nevada weather station in California about 50 miles south of Honey Lake Valley, were used in the DPM grid at a cell that has similar altitude (about 6,000 feet), slope, and aspect (figure 8). Only temperature data, not precipitation data, from the Truckee Station were used. The lapse rates (vertical temperature gradient) used by Bauer and Vaccaro (1987, p. 75) for minimum and maximum monthly temperatures in southeastern Washington, and by Morgan (1988, p. 87) in the Goose Lake basin about 120 miles north of Honey Lake, were used for this study. Sublimation and snowmelt rates were adjusted to match simulated snowpack duration and maximum depth with observed snowpack conditions at the Truckee station. A sublimation rate of 0.1 in/d and snowmelt rate of 0.4 (in/°C)/d resulted in the closest approximation of simulated to observed snowpack.

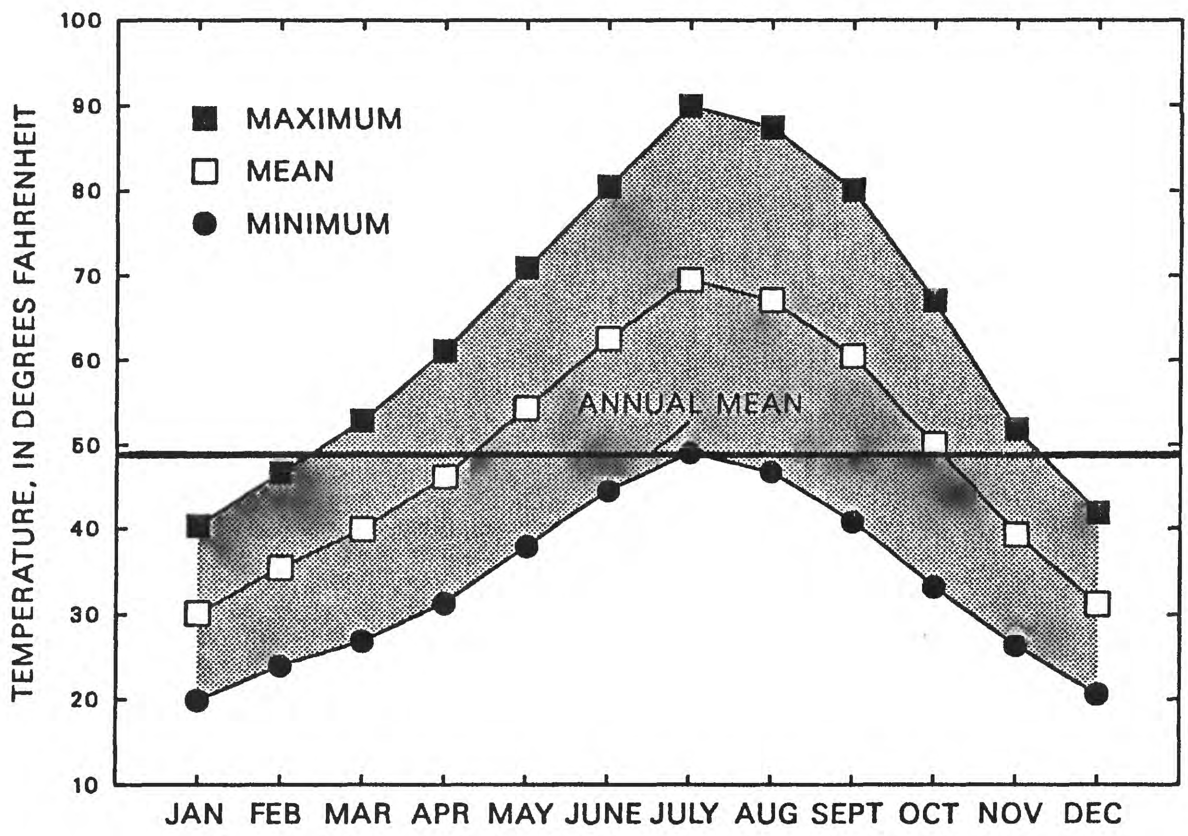


FIGURE 9.--Average minimum, mean, and maximum monthly air temperatures at Susanville Airport, 1951-80 (data are from National Climatic Center, 1982).

Land-surface altitude, slope, and aspect were calculated at each grid cell by digitizing altitude contours from topographic maps and applying a computer gridding routine (Webring, 1981). Land-surface altitude for each cell was obtained by gridding the digitized altitude contours at 1-mile intervals. Altitudes were then gridded at 1/3-mile intervals to produce nine values for each 1-mi^2 cell, and programs developed by Graham and others (1986) were used to determine slope and aspect from these values.

The predominant soil type or association for each grid cell was determined from 7.5-minute soil maps (Baumer, 1983; U.S. Soil Conservation Service, written communication, 1988). Additional historical soil information was obtained from Guernsey and others (1917). Average texture, available water-holding capacity, and root-penetration depth or soil thickness of each type was computed from data obtained from the U.S. Soil Conservation Service (written communication, 1988). The predominant soil types and associations were grouped into 24 categories on the basis of these characteristics (table 3), and each grid cell was assigned a number from 1 to 24. The distribution of soil groups used in the DPM is shown in figure 10.

Land cover at each grid cell was estimated from land-use and land-cover maps (U.S. Geological Survey, 1979a, 1979b, 1980a, 1980b, 1983), and assigned to one of six categories: forest, grass (includes residential and other built-up areas), sage (to represent rangeland), alfalfa (to represent agricultural land), surface water and wetlands, or bare soil (figure 11). Land-use and land-cover categories are defined and classified by Anderson and others (1976). Maximum values for root depth, percentage of foliage cover, and interception capacity for DPM land-cover categories in this study (table 4) are the same as those used by Morgan (1988, p. 89) in the Goose Lake area, and are based on information from the U.S. Soil Conservation Service and the U.S. Forest Service. Root depths are for vegetation that uses soil moisture derived from precipitation. Ground-water use by phreatophytes is not included in the DPM.

Data on streamflow and application of irrigation water can be included in DPM calculations for individual cells. The largest source of streamflow and irrigation water in the Honey Lake basin is the Susan River. Most surface water ultimately is transpired by crops or evaporated from the surface of Honey Lake, but during part of the year some streamflow and irrigation water infiltrates to the water table. These sources were not included in the DPM for Honey Lake Valley because daily streamflow and irrigation records for the simulation period are unavailable for most of the study area. To account for these contributions to deep percolation, separate estimates were made. These are discussed in the sections titled "Infiltration of Streamflow" and "Infiltration of Irrigation Water."

Calibration.--The Deep Percolation Model was calibrated by adjusting sublimation and snowmelt rates to achieve a satisfactory seasonal distribution of recharge throughout the basin and to match the annual duration of snowpack at an altitude of about 6,000 feet. Part of the calibration process was the addition of temperature data from the Truckee Ranger Station. Without the inclusion of daily minimum and maximum temperatures from the Truckee Station, the maximum simulated recharge from deep percolation occurred during the months of maximum precipitation (December, January, and February; figure 6). The cold temperatures provided by the Truckee data increased simulated snowpack thickness and duration, thus delaying maximum recharge.

TABLE 3.--*Characteristics of soil groups used for Deep Percolation Model calculations*

[Summarized from soil maps and data from Baumer (1983) and U.S. Soil Conservation Service (written commun., 1988). Distribution of soil groups is shown in figure 10.]

Soil group number	Texture[1]		Available water capacity[2]		Thickness[3]		Number of 6-inch layers in model cell	Number of soil types in group
	Range	Average	Range	Average	Range	Average		
1	1.0-1.6	1.2	0.3-0.7	0.4	21-46	30	5	10
2	1.0-1.7	1.3	0.4-0.7	.5	60+	60+	10	17
3	1.5-1.6	1.6	0.9-1.2	1.0	21-30	25	4	2
4	1.8-1.9	1.8	0.5-0.6	.6	9-28	18	3	8
5	1.8-1.9	1.8	0.4-0.6	.6	30-50	38	6	19
6	1.8-1.9	1.8	0.3-0.6	.5	60+	60+	10	22
7	1.8-1.9	1.8	0.7-0.9	.8	10-25	15	2	9
8	1.8-1.9	1.8	0.7-0.8	.8	30-53	40	7	11
9	1.8-1.9	1.8	0.7-0.9	.8	60+	60+	10	14
10	1.8-1.9	1.8	1.0-1.2	1.1	23-30	27	4	4
11	2.0	2.0	0.4-0.6	.6	10-29	22	4	27
12	2.0	2.0	0.5-0.6	.6	30-48	35	6	12
13	2.0-2.1	2.0	0.3-0.6	.5	60+	60+	10	9
14	2.0-2.1	2.0	0.7-0.9	.8	16-26	21	4	21
15	2.0-2.1	2.0	0.7-0.9	.8	30-58	40	7	15
16	2.0-2.1	2.0	0.7-0.9	.7	60+	60+	10	10
17	2.0	2.0	1.0-1.1	1.0	20-25	22	4	2
18	2.1	2.1	1.2	1.2	50	50	8	1
19	2.0-2.1	2.0	1.0-1.1	1.0	60+	60+	10	5
20	2.2-3.0	2.5	0.6-0.8	.8	19-28	22	4	10
21	2.6-3.0	2.9	0.6-0.9	.8	30-51	35	6	11
22	2.3-2.9	2.7	0.5-0.9	.8	60+	60+	10	16
23	2.7-3.0	2.9	1.1-2.1	1.4	30-50	42	7	5
24	2.3-3.0	2.6	1.0-1.2	1.1	60+	60+	10	3

[1] Total range (dimensionless) is from 1.0 (sand) to 3.0 (clay), on the basis of the uppermost layer of soil.

[2] Inches of water per 6-inch thickness of soil.

[3] Average thickness of soil, in inches, from land surface to impervious layer (hardpan or bedrock).

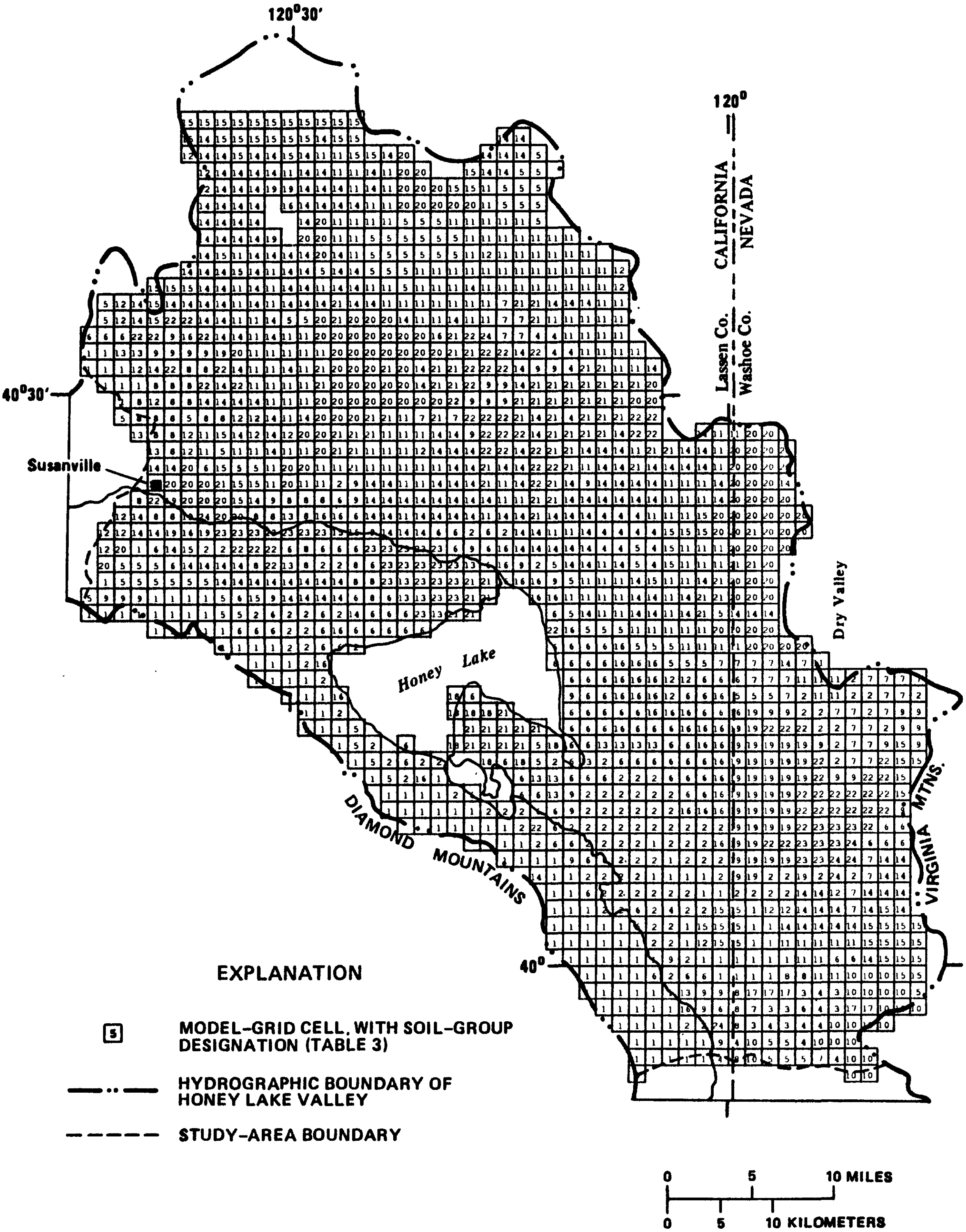


FIGURE 10.--Distribution of soil groups for the Deep Percolation Model.

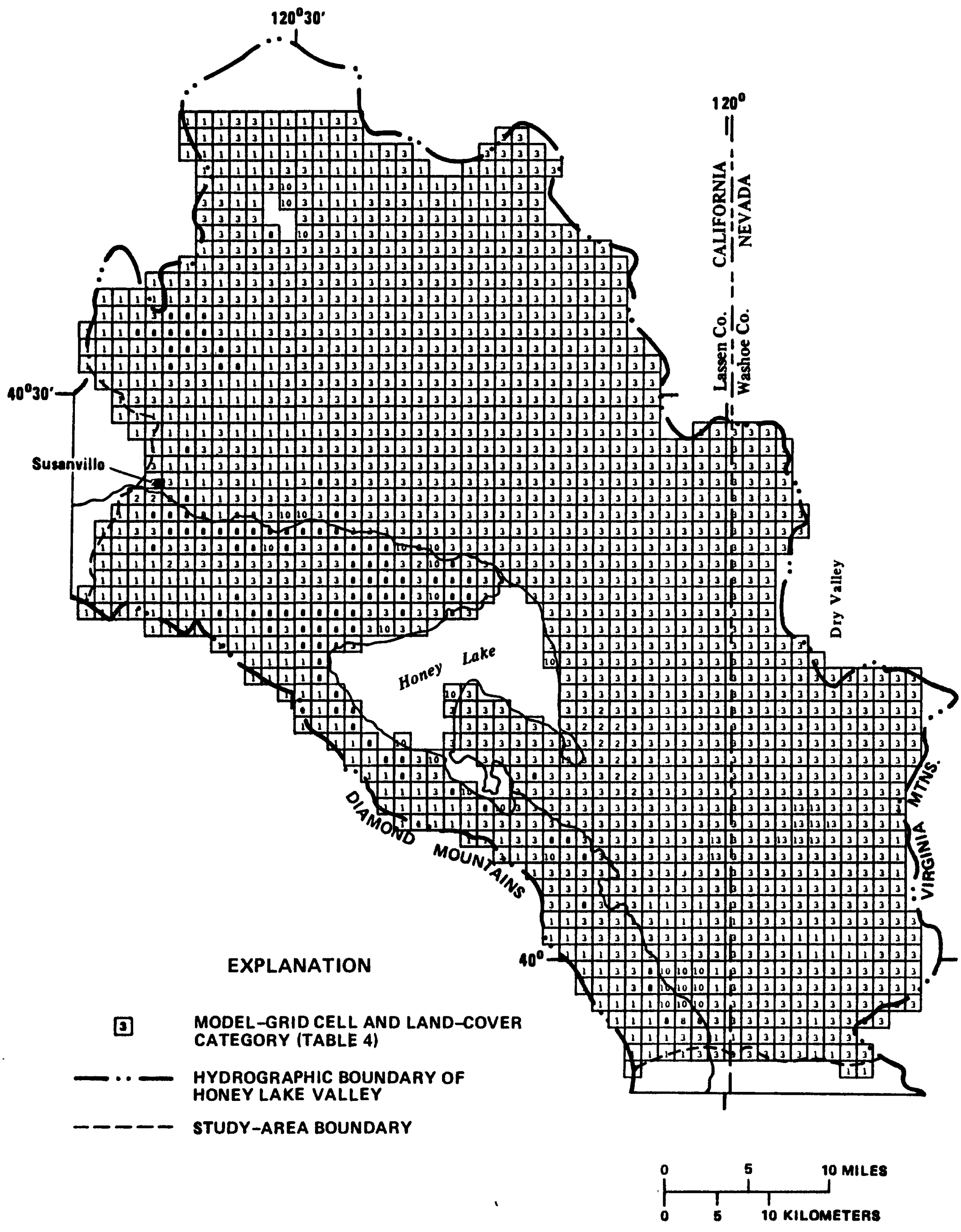


FIGURE 11.--Distribution of land cover used for the Deep Percolation Model.

TABLE 4.--*Land-cover characteristics used for Deep Percolation Model calculations*

[From Morgan, 1988, p. 89]

Category (figure 11)	Cover type	Maximum root depth (feet)	Maximum foliar cover (percent)	Maximum precipitation-interception capacity (inches)
1	Forest	3.0	80	0.4
2	Grass[1]	2.0	100	.06
3	Sage[2]	4.6	[3]35	.06
8	Alfalfa[4]	5.0	100	.11
10	Water and wetlands	0	0	0
13	Barren land	0	0	0

[1] Includes residential and other built-up areas.

[2] Represents all rangeland.

[3] Maximum cover is lower than 35 percent in northern and eastern parts of basin, and higher in south and west.

[4] Represents all agricultural land.

To calibrate the model, snow-course data from the Truckee Ranger Station were compared with the DPM results for each year of the simulation period (1961-80). The dates of first and last days of measureable snowpack and the date of maximum observed snow depth at the Truckee station were compared with DPM-simulated snowpack dates at a grid cell that is similar to the Truckee station in altitude, slope, and aspect. The Truckee station and the grid cell represent forested areas at about 6,000-feet altitude. Analysis of 11 years of data from the Central Sierra Snow Laboratory near Soda Springs, California, indicates that the mid-altitude (6,200 to 7,200 feet) Sierra Nevada snow zone has a mid- to high-density forest canopy, which extends the period of snowmelt runoff as much as 4 weeks compared to open areas (Bergman, 1985). This geographic setting and extended snowpack period applies to the real and simulated Truckee sites as well. The DPM sublimation rate and snowmelt coefficient were adjusted to obtain satisfactory matches between simulated and observed snowpack duration and simulated and observed dates of maximum snow depth. The results for each year of the 20-year simulation period are shown in figure 12. For this period, the maximum simulated recharge from deep percolation occurs in March. Most of the recharge is at higher altitudes, where water-level data are unavailable for comparison.

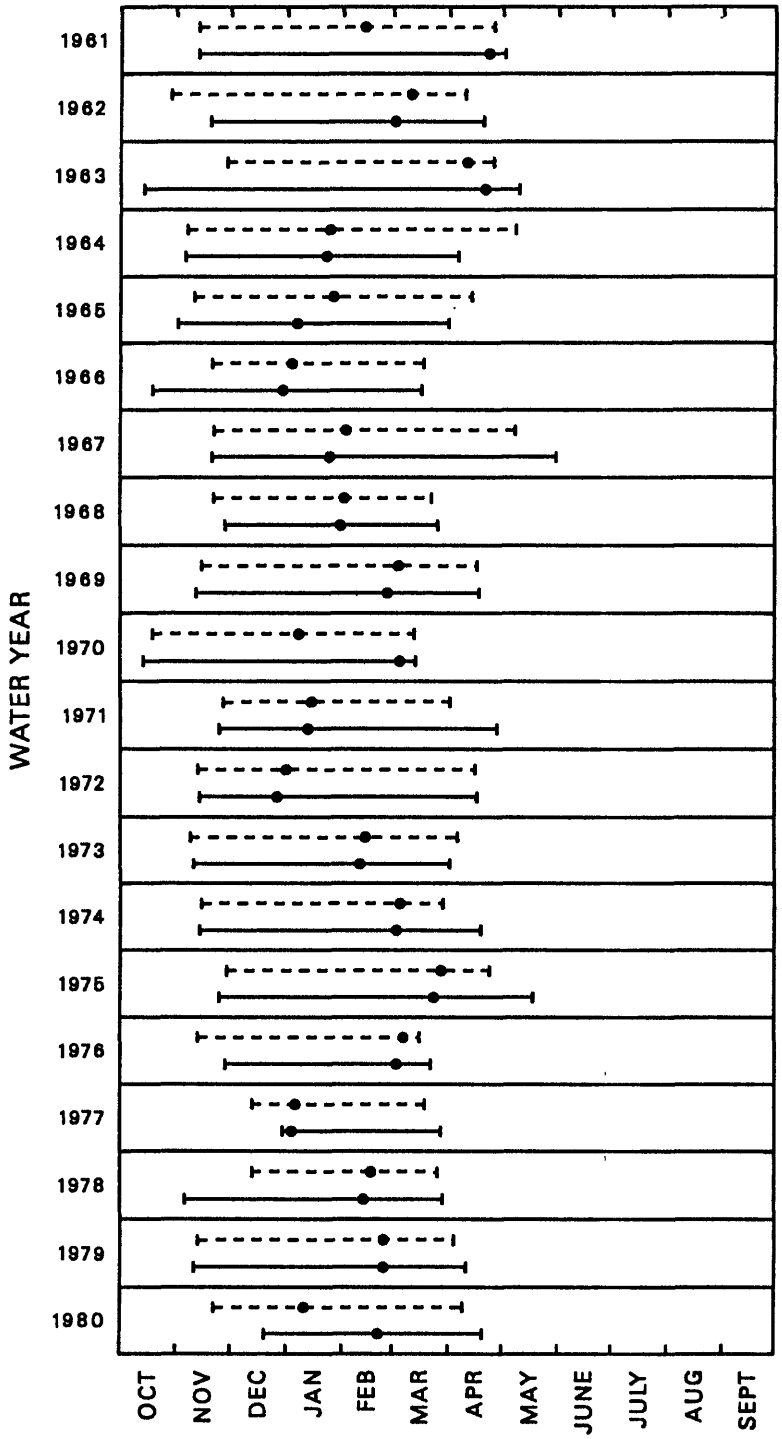


FIGURE 12.--Simulated and observed snowpack duration (upper line and lower line, respectively, of each pair), 1961-80, at the high-altitude meteorological site (fig. 8). Lines extend between dates of first and last measurable snowpack; intermediate point is date of maximum snow depth.

Results.--Mean monthly water budgets (figure 13) calculated by the DPM show seasonal variations in evapotranspiration, runoff, and recharge. Evapotranspiration shown in figure 13 is the sum of soil evaporation, plant transpiration, evaporation of water intercepted by plants before it reaches the ground, and snow sublimation. Actual evapotranspiration from available moisture shown in figure 13, is much less than potential evaporation (unlimited moisture) during most of the year because precipitation is scant during the months when potential evapotranspiration is greatest. The average budget balances although the monthly budgets do not. During the summer, evapotranspiration plus runoff exceed precipitation, and no recharge occurs. During the winter, precipitation exceeds the total evapotranspiration, runoff, and recharge because some precipitation is stored as snowpack and soil moisture. Simulated runoff is greatest in the winter months because precipitation is greatest during those months and because the DPM assumes that rain on snow immediately runs off, although in reality it may be absorbed by the snowpack, thus delaying runoff. Estimated ground-water recharge by deep percolation is greatest in February and March, when mean air temperature is above freezing in much of the basin, but the growing season has not as yet begun.

Long-term annual recharge, as calculated by the DPM, differs with location as well as with season. Estimated annual recharge ranged from nil over large parts of the valley floor to more than 4 inches along the crest of the Diamond Mountains, as shown in figure 14. The DPM estimate of mean annual recharge from precipitation for the study area totals about 55,000 acre-ft.

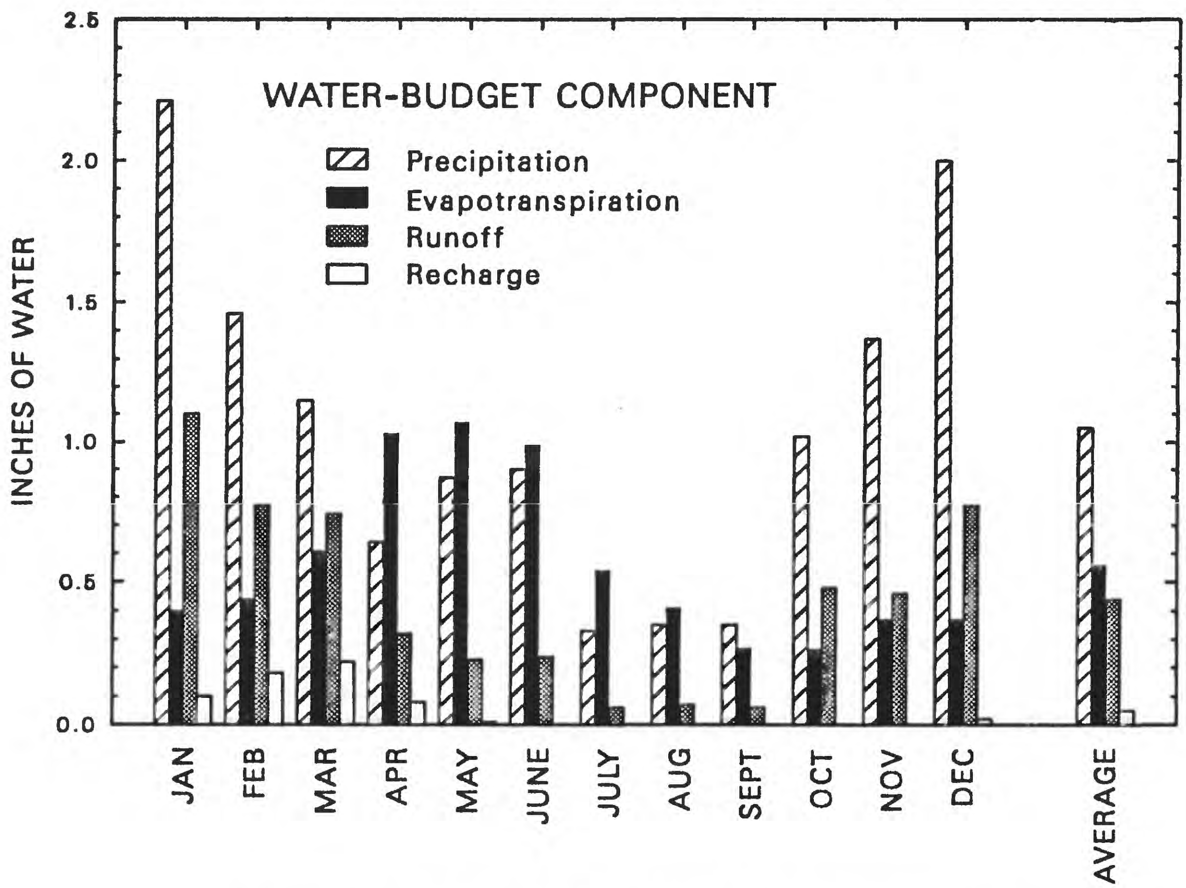


FIGURE 13.--Mean monthly values for water-budget components, 1961-80, as simulated by use of the Deep Percolation Model.

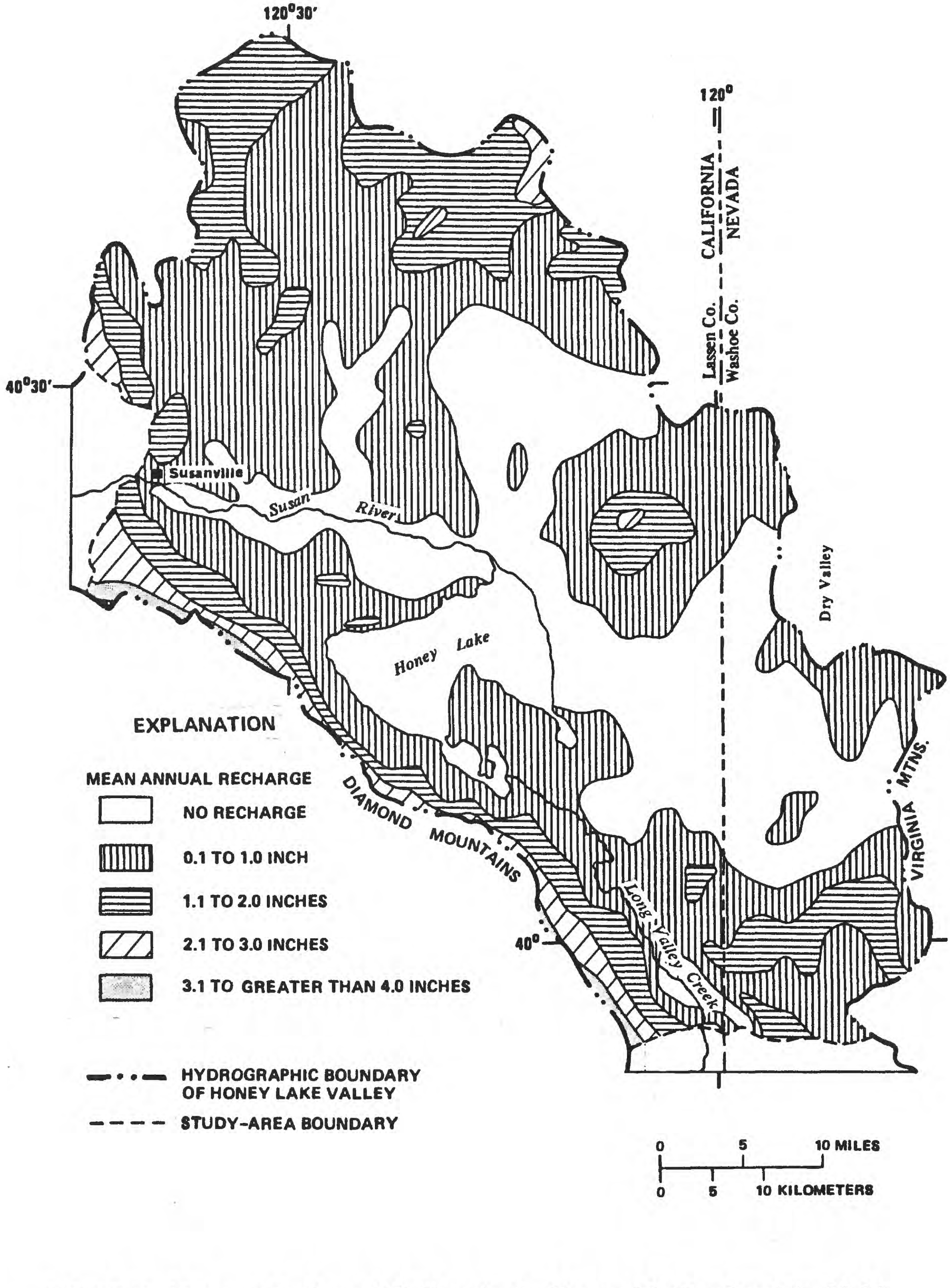


FIGURE 14.--Mean annual recharge, 1961-80, as simulated by use of the Deep Percolation Model.

Total runoff was calculated by the model for the Willow, Skedaddle, and Cottonwood Creek drainages (figure 8) and compared with measured or estimated streamflow from each of the three basins to evaluate DPM results (table 5). The DPM calculates runoff from each grid cell, but it does not route the flow from cell to cell. Therefore, downgradient infiltration losses are not calculated, and runoff from many areas could be overestimated.

The three basins (figure 8) were selected on the basis of diversity of drainage-area characteristics and location, availability of streamflow measurements or estimates, and transferability of results to the ground-water flow model. Willow Creek, which drains an area of 90.4 mi^2 in the northwest part of the study area, is a gaged stream with a long period of record. The Skedaddle Creek basin (83.4 mi^2) and Cottonwood Creek basin (14.6 mi^2) contribute streamflow to the eastern part of the study area. Long-term annual streamflow of Skedaddle and Cottonwood Creeks was estimated from monthly measurements by correlation with average monthly and annual measurements on gaged streams (Rockwell, in press).

On the basis of data for 1961-80 water years, the average streamflow from the Willow Creek basin is 25,000 acre-ft/yr (table 5). Runoff simulated by the DPM (37,000 acre-ft/yr) for the same period is 1.5 times greater for approximately the same area. Streamflow simulated by the DPM for Cottonwood Creek (3,700 acre-ft/yr) is 2.3 times greater than the estimated streamflow, and streamflow simulated for Skedaddle Creek (17,000 acre-ft/yr) is 3.4 times greater than estimated streamflow.

The Willow Creek basin provides the best comparison because it has a long period of record and because relatively impermeable, granitic bedrock along the southwest margin of the area impedes subsurface outflow. Consequently, most of the outflow is at the surface and is measured by the gage. Furthermore, the Willow Creek stream gage is in a canyon upstream from areas where substantial infiltration losses are likely. Streamflow estimates for the Skedaddle and Cottonwood Creek basins are less accurate because they are based on limited data, the drainage areas are underlain by fractured volcanic rocks that are permeable in places, and the potential exists for upstream infiltration and subsurface outflow.

TABLE 5.--*Simulated runoff and measured or estimated runoff from Willow, Skedaddle, and Cottonwood Creek drainage areas, 1961-80*

Stream	Simulated with Deep Percolation Model: Runoff (acre-feet per year)	Simulated with Deep Percolation Model: Drainage area (number of square-mile cells)	Measured or estimated: Runoff (acre-feet per year)	Measured or estimated: Drainage area (square miles)	Ratio of simulated runoff to measured or estimated runoff
Willow Creek	37,000	87	[a]25,000	90.4	1.5
Skedaddle Creek	17,000	81	5,000	83.4	3.4
Cottonwood Creek	3,700	13	1,600	14.6	2.3

[a] Measured.

In summary, the DPM is a useful tool in estimating recharge, but it could underestimate recharge in areas where precipitation is not the sole source of recharge, in places where materials beneath the soil zone are permeable, and in places where rainfall is temporarily stored in an existing snowpack. To improve recharge estimates, data on daily streamflow and application of irrigation water should be included in the DPM. Because daily streamflow and irrigation rates were not available for this study, a separate, independent estimate of surface-water infiltration was made.

Infiltration of Streamflow

Total mean annual streamflow in the study area, summarized in table 6, is an estimated 230,000 acre-ft (Rockwell, in press), on the basis of (1) continuous measurements at 3 streams, (2) monthly measurements during 1988 at 18 small streams that were adjusted to long-term average by comparison to gaging station records, and (3) relations between drainage area and long-term average streamflow (that is, average flow per unit drainage area) for 10 intermittent streams and for unmeasured basins in the rest of the study area. Streamflow was estimated at mountain-front locations for all drainages areas except for two streams: the Susan River is measured at the stream gage at Susanville and the flow of Long Valley Creek is estimated from measurements at the study-area boundary south of Doyle. The estimated streamflow total is 21 percent higher than the amount estimated by Clements (1988, p. 8) for surface-water and subsurface inflow to the valley floor. The difference may be a result of different methods used in the two studies to estimate flow from ungaged areas.

Some streamflow evaporates or is transpired by vegetation along stream channels, some of it flows into Honey Lake, and some of it percolates to become ground-water recharge. To determine the relation of streamflow to ground-water recharge for streams in different settings, seepage measurements were made on eight reaches of five streams in the basin (Fort Sage, Long Valley, Mill, Piute, and Gold Run Creeks) in December 1987, when evapotranspiration was minimal (Rockwell, in press). Seepage rates are estimated from near-simultaneous measurements of streamflow at different places along a channel to determine whether streamflow is increasing or decreasing in a downstream direction. The five streams are shown on plate 1. These measurements were insufficient to define specific relations, but they can be used to verify infiltration, and some generalizations can be extrapolated from them. In upland draws in the northwest part of the basin (Piute and Gold Run Creeks) and in the Sierra Nevada (Mill Creek), streams may gain or lose water in different reaches depending on slope, bed materials, and stage. Larger streams (Long Valley Creek) may gain water after they reach the valley floor. However, in the most arid parts of the basin (Fort Sage Creek), nearly all flow infiltrates through permeable deposits and fractured rock to the ground-water system. In areas of fractured volcanic rock north and east of the Fort Sage Mountains, streamflow seldom reaches the valley floor; for example, the flow of Skedaddle Creek decreased at a rate of about 0.8 ft^3/s per mile of reach (Gerald L. Rockwell, U.S. Geological Survey, oral communication, 1988).

Along the Sierra Nevada, most seepage is through alluvial fan and nearshore deposits. The downstream extent of these deposits, measured from the contact with bedrock to the contact with less permeable lake deposits, averages about 1 mile. If the seepage rate is assumed to be 0.34 $(ft^3/s)/mi$, which is the average rate for four losing reaches reported by Rockwell (in press), then annual ground-water recharge from 12 perennial streams draining the Sierra Nevada into Honey Lake Valley would be about 3,000 acre-ft. This rate probably is a minimum because the measurements were made during 1988, an unusually dry water year, when little streamflow was available to infiltrate. Seepage from 13 reaches on 6 streams in similar settings in the Goose Lake basin, about 120 miles north of Honey Lake Valley on the California-Oregon border, averaged 0.5 $(ft^3/s)/mi$ (Morgan, 1988, p. 26), which is also lower than the rate observed on Skedaddle Creek. By using the 0.5 $(ft^3/s)/mi$ rate, annual recharge from the 12 Sierra Nevada streams in

Honey Lake Valley would be about 4,300 acre-ft; by using the Skedaddle Creek rate of 0.8 (ft^3/s)/mi, it would be about 7,000 acre-ft. Infiltration from more than 20 additional streams that drain the Sierra Nevada is difficult to estimate because the streams are intermittent; on the basis of streamflow-to-drainage area relations (estimated at 0.54 (ft^3/s)/mi^2 by Rockwell [in press]), however, it would total about 8,000 acre-ft during dry years and could be considerably greater during normal years. On the basis of seepage observations, average annual recharge along the Sierra Nevada from Willow Ranch Creek to Gold Run probably is more than 12,000 acre-ft.

In the north and east parts of the study area, vegetation in stream channels is sparse, irrigation diversions are uncommon, and, in dry to normal years, nearly all streamflow infiltrates. From Spencer Creek to Fort Sage Creek, annual streamflow is about 13,000 acre-ft (table 6), hence annual recharge also is about 13,000 acre-ft.

TABLE 6.--*Distribution of mean annual streamflow*

[Based on data from Rockwell (in press)]

Part of basin (plate 1)	Drainage area[1] (square miles)	Streamflow[2] (acre-feet)
North and east (Spencer Creek near Herlong to Fort Sage Creek near Flanigan [ground-water flow-model area])	168	13,000
Southeast (Dry Valley Creek near Doyle to Willow Ranch Creek near Doyle, excluding Long Valley Creek)	135	8,400
Long Valley Creek near Doyle	266	17,000
South and west (Willow Ranch Creek near Doyle to Willow Creek near Susanville, excluding Susan River)	282	90,000
Susan River at Susanville (gaged)	184	69,000
Northwest (between Willow Creek and Spencer Creek)	565	31,000
Total area, excluding valley floor	1,600	230,000

[1] Rounded to three significant figures.

[2] Rounded to two significant figures.

Almost all irrigation diversions are from streamflow in the southeast to northwest parts of the basin, principally from the Susan River, Gold Run Creek, and Long Valley Creek. Annual streamflow from these areas totals about 220,000 acre-ft (table 6). About 54,000 acre-ft is diverted for irrigation, as discussed in the following section of this report, leaving about 170,000 acre-ft of streamflow to recharge ground water or flow to Honey Lake. The difference between 170,000 acre-ft and the amount of streamflow that reaches the lake is approximately equal to ground-water recharge from these areas, assuming that evapotranspiration from streams and stream channel vegetation is minor.

Some streamflow from the Susan River and Long Valley Creek reaches Honey Lake during periods of snowmelt, occasional large storms, and as irrigation return flow during the growing season. Streams from the Diamond Mountains of the Sierra Nevada also discharge into the lake during parts of the year. Much of the water in Honey Lake is from these sources, although a small amount probably is from ground-water discharge into the lake by natural seepage through the lake bottom and by discharge of geothermal water. In addition, approximately 39,000 acre-ft is from precipitation directly onto the lake surface each year.

On the basis of bathymetric measurements of Honey Lake (Rockwell, in press) and water-surface altitudes for 1984-88 (figure 15), and assuming that the contribution to the lake from ground-water seepage is negligible, annual streamflow into the lake was estimated using the following relation:

$$I = \Delta V - P + E,$$

where I = annual lake inflow from streams,
ΔV = change in lake volume during the inflow period,
P = precipitation onto the lake surface during the inflow period, and
E = evaporation from the lake surface during the inflow period.

The lake volume was determined from bathymetric data (Rockwell, in press). Monthly precipitation was estimated as 0.7 times the recorded precipitation at Susanville Airport, on the basis of the ratio of mean annual precipitation over the lake (10 inches) to mean annual precipitation at Susanville Airport (14 inches; see figure 5). Monthly precipitation data are from the National Weather Service. On the basis of monthly pan-evaporation measurements at the Fleming Wildlife Area (Robert Anton, U.S. Soil Conservation Service, written communication, 1988), lake-surface evapotranspiration was estimated using a pan-evaporation coefficient of 0.72 (Farnsworth and others, 1982, map 4). Long-term annual streamflow into the lake was estimated using the relation of annual streamflow for each year during water years 1984-88 to long-term annual streamflow for the period of record for the Susan River (1900-88 water years) and Willow Creek (1951-88 water years). Results indicate that the long-term annual streamflow into the lake is about 130,000 acre-ft (table 7). The difference in streamflow, about 40,000 acre-ft (170,000 acre-ft remaining after irrigation diversions, minus 130,000 acre-ft that reaches the lake), is approximately the amount of streamflow that recharges ground water from the northwest, west, and southwest parts of the basin.

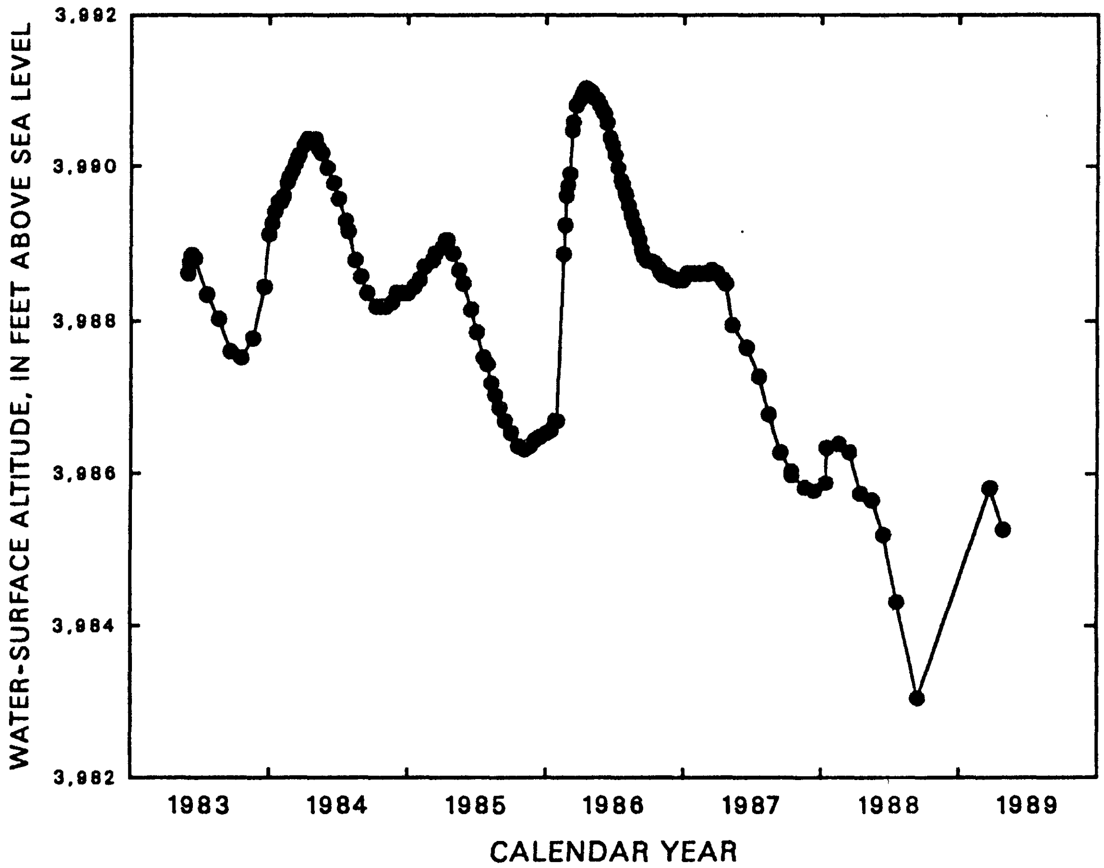


FIGURE 15.--Water-surface altitude of Honey Lake, June 1983 through April 1989 (data from California Department of Fish and Game, Fleming Unit, Wendel, Calif., and U.S. Geological Survey, Sacramento, Calif.).

The Honey Lake water budget was also used to estimate the average lake volume and lake-surface altitude. The estimated long-term streamflow into the lake (130,000 acre-ft/yr) is assumed to be balanced by an equivalent amount of net surface evaporation. This balance occurs when the lake-surface area is about 47,000 acres. This area corresponds to a lake volume of about 120,000 acre-ft and a lake-surface altitude of about 3,983 feet above sea level.

TABLE 7.--*Estimated stream inflow to Honey Lake, water years 1984-88, and calculated long-term average stream inflow*

Water year (1)	Estimated stream inflow to lake (acre-feet) (2)	Streamflow measured at gage, as percentage of long-term annual average at gage[1]			Calculated long-term average stream inflow to lake (acre-feet per year, rounded), on basis of comparison between annual estimates (column 2) and data for Susan River and Willow Creek (column 5) (6)
		Susan River (3)	Willow Creek (4)	Average of columns 3 and 4 (5)	
1984	192,000	120	130	125	150,000
1985	86,000	40	81	60	140,000
1986	274,000	149	155	152	180,000
1987	22,000	28	64	46	50,000
1988	52,000	15	60	38	140,000
Mean of calculated values for long-term stream inflow to lake (rounded)					130,000

[1] Based on U.S. Geological Survey data for Susan River, 1900-88 water years, and for Willow Creek, 1951-88 water years.

Infiltration of Irrigation Water

Additional sources of ground-water recharge are infiltration of irrigation water from surface-water and ground-water sources, and a minor amount of seepage from waste-water disposal systems. The estimated average volume of surface water that is diverted from the Susan River, Long Valley Creek, and other streams within the study area for irrigation each year is 54,000 acre-ft. This is computed from a range of about 41,000 acre-ft to about 67,000 acre-ft. The smaller amount is based on a 1985 water-use estimate of 46,000 acre-ft by William E. Templin (U.S. Geological Survey, written communication, 1988), minus about 5,000 acre-ft of water used in Long Valley upstream from the study area. The larger amount is based on an estimate of 72,000 acre-ft by Walters Engineering (1986, table 2), minus about 5,000 acre-ft for Long Valley use. Return flows reported in the western United States, including both conveyance loss (seepage from canals) and deep percolation of water applied to fields in excess of crop needs, range from 3 to 86 percent of the water diverted for irrigation; the average return flow has been estimated as 25 to 33 percent (Lauritzen and Terrell, 1967, p. 1105). On the basis of these estimates, assuming a conservative average irrigation return of 25 percent in Honey Lake Valley, about 14,000 acre-ft of water annually infiltrates to the ground-water flow system from surface-water irrigation. The conveyance loss reported for Honey Lake Valley for 1985 was 840 acre-ft (William E. Templin, U.S. Geological Survey, written communication, 1988); the rest of the estimated irrigation return is from deep percolation of applied water. Total study-area streamflow (about 230,000 acre-ft, minus irrigation diversions, 54,000 acre-ft/yr), leaves about 180,000 acre-ft of surface water available for evapotranspiration and ground-water recharge in an average year.

Annual withdrawals of ground water for irrigation are about 43,000 acre-ft (based on Clements, 1988, p. 8). Assuming that 25 percent of the total withdrawal infiltrates, the same average infiltration used for estimating surface-water irrigation return, recharge from this source would be about 11,000 acre-ft. Because annual withdrawals of ground water for all other uses in the study area total only about 10,000 acre-ft (table 8), return flow from these other sources is a negligible component of recharge.

TABLE 8.--*Estimated annual ground-water use in study area*[1]

[Acre-feet, rounded to two significant figures]

Water-use category	California	Nevada	Total
Public supply (all uses)	3,900	0	3,900
Commercial (self-supplied)	240	0	240
Domestic (self-supplied)	640	4	640
Industrial:			
self-supplied	1,200	0	1,200
public-supplied	22	0	22
Geothermal	3,400	0	3,400
Livestock	280	10	290
Irrigation	37,000	5,900	43,000
Total withdrawal	47,000	5,900	53,000

[1] Data from Clements (1988), California Department of Water Resources, and U.S. Geological Survey.

Comparison with Maxey-Eakin Estimate of Potential Recharge

A method of estimating potential ground-water recharge as a percent of precipitation was developed for east-central Nevada (Maxey and Eakin, 1949; Eakin and others, 1951). It is an empirical relation between average annual precipitation within a basin and incremental recharge to ground water, based on zones of precipitation. The method assumes that the percentage of precipitation that ultimately contributes to recharge is about 25 percent where the average annual precipitation is greater than 20 inches, about 15 percent in the 15- to 20-inch precipitation zone, 7 percent in the 12- to 15-inch zone, 3 percent in the 8- to 12-inch zone, and nil where the average annual precipitation is less than 8 inches.

By using this method and the precipitation zones shown in figure 5, average annual recharge generated within the study area is estimated to be about 95,000 acre-ft/yr, or about 9 percent of the total volume of precipitation in zones where precipitation exceeds 8 inches. The Maxey-Eakin method was developed for closed basins in eastern Nevada; the precipitation-recharge relation may be different in less arid basins such as Honey Lake Valley. An evaluation of the relation by methods other than Maxey-Eakin in 174 basins in the Great Basin province in Nevada indicates that 3 to 10 percent of precipitation exceeding 8 inches becomes ground-water recharge in these basins (James R. Harrill, U.S. Geological Survey, written communication, 1989). In Honey Lake Valley, potential recharge would be close to or possibly greater than 10 percent of the precipitation exceeding 8 inches because of a wetter climate in the Sierra Nevada and western part of the basin.

To compare the Maxey-Eakin estimate of potential recharge with the total ground-water recharge estimated by other methods for this study, an adjustment was needed to account for potential recharge that originates within the Honey Lake Valley drainage area but outside the study area. About 86,000 acre-ft of surface-water enters the study area in the Susan River and Long Valley Creek in an average year (table 6). Based on the assumption that 30 percent of the water infiltrates (the average rate estimated for the basin), then the Maxey-Eakin estimate should be increased by about 26,000 acre-ft to a total of about 120,000 acre-ft.

A comparison of the Maxey-Eakin estimate of potential recharge with independent estimates of recharge from precipitation and surface-water infiltration (table 9) shows that the two estimates differ significantly for small basins but are close for the overall study area. The Maxey-Eakin technique is not intended for use on individual drainages within a basin; estimates for Willow, Skedaddle, and Cottonwood Creeks are included only for comparative purposes. For the entire study area, the estimates were approximately equal.

TABLE 9.--*Recharge values calculated from infiltration and Maxey-Eakin estimates*

[Acre-feet per year, rounded to two significant figures]

	Infiltration estimates				
Area	From precip-itation	From stream-flow and surface-water irrigation	Total	Maxey-Eakin estimates	Ratio of infiltration estimate to Maxey-Eakin estimate
Willow Creek	3,600	13,000	17,000	8,300	2.0
Skedaddle Creek	1,000	5,000	6,000	2,220	2.7
Cottonwood Creek	650	1,600	2,200	1,000	2.2
Flow-model area	4,200	13,000	17,000	11,000	1.5
Study area	55,000	70,000	120,000	[a]120,000	1.0

[a] Adjusted for infiltration of water transported into the study area by Susan River and Long Valley Creek.

Subsurface Inflow

Estimates of recharge were based on the assumption that ground-water and surface-water (topographic) divides coincide and that all ground-water recharge originates as precipitation over the study area or as streamflow that enters the area in the Susan River and Long Valley Creek. However, several factors have led to speculation that ground water also enters the study area from adjacent basins. These factors include (1) the large flow from springs before ground-water withdrawals began at Fish Springs Ranch in the southeast part of the basin, (2) an observed deficiency of discharge by evapotranspiration in Warm Springs (Palomino) and Dry Valleys to the southeast, (3) an apparent imbalance between recharge and discharge in the eastern part of Honey Lake Valley, and (4) the existence of geothermal water and major faults in the basin. Some of these factors were discussed by Rush and Glancy (1967, p. 42) and R.W. Beck and Associates (1987, p. II-3) and were examined for the present study.

Although ground-water levels beneath the floor of Warm Springs Valley are higher than those in Honey Lake Valley (Bedinger and others, 1984, sheet 1; Reed and others, 1984), and deep faults associated with the Walker Lane structure pass through both valleys, the confirmation of interbasin flow requires additional data. To test the possibility of inflow from the southeast, a pair of wells (wells 14 and 15) was installed in Section 33, Township 26, Range 18, near the Warm Springs Fault. Well locations are shown on plate 1 and well data are in table 10. One well of the pair is 400 feet deep; the other is 290 feet deep. During a 1-year period of measurement, water levels in the deeper well were consistently 0.2-0.6 foot lower than those in the shallower well. This indicates that the vertical component of ground-water flow at this location is slightly downward. An upward gradient would suggest possible subsurface inflow, but a lateral or downward gradient at this site is inconclusive.

TABLE 10.--*Data for wells referred to in this report*

[--, data not available]

Well No. (plate 1)	U.S. Geological Survey site identification[1]	Local site number[2]	Land-surface altitude (feet above sea level)	Open interval (feet below land surface)	
				Top	Bottom
		California wells			
1	400728120005901	26N 17E 11A01	4,000	444	456
2	400728120005902	26N 17E 11A02	4,000	90	100
3	400832120004701	26N 17E 01D01	4,005	--	[a]158
4	400918120011601	27N 17E 35B01	4,010	--	[a]57
5	401057120071001	27N 16E 24G01	4,018	--	[a]188
6	401223120070701	27N 16E 12J01	4,009	--	[a]139
7	401236120085901	27N 16E 11E01	4,000	--	[a]400
8	401416120033101	27N 17E 03H01	4,010	--	[a]20
9	401604120081601	28N 16E 23J01	4,004	--	[a]230
10	401745120251101	28N 14E 08J01	4,000	142	502
11	401822120261701	28N 14E 07A01	4,020	40	535
12	402350120291501	29N 13E 02L01	4,080	--	[a]56
13	402614120265701	30N 14E 19P01	4,180	--	--

TABLE 10.--*Data for wells referred to in this report*--Continued

Well No. (plate 1)	U.S. Geological Survey site identification[1]	Local site number[2]	Land-surface altitude (feet above sea level)	Open interval (feet below land surface)	
				Top	Bottom
		Nevada wells			
14	400435119583201	97 N26 E18 33CBCA1	4,145	388	400
15	400435119583202	97 N26 E18 33CBCA2	4,145	273	290
16	400507119560001	97 N26 E18 35BABA1	4,176	--	[a]1,340
17	400509119530401	97 N26 E19 30DDDD1	4,025	95	492
18	400532119545501	97 N26 E18 25CABA1	3,979	36	246
19	400533119554801	97 N26 E18 26ACCC1	3,995	160	440
20	400555119524101	97 N26 E19 29BABC1	4,012	60	400
21	400557119554401	97 N26 E18 26ABBD1	3,980	13	18
22	400600119562301	97 N26 E18 23CCCC1	3,988	269	599
23	400701119565201	97 N26 E18 15DCBD1	3,979	--	[a]488
24	400759119504001	97 N26 E19 10CBCC1	4,012	63	255
25	400849119485301	97 N26 E19 02DCA 1	4,172	224	240
26	400858119552501	97 N26 E18 02CDBA1	3,986	0	235
27	400903119571501	97 N26 E18 03CABB1	3,988	168	188
28	400928119540301	97 N26 E19 06BBBB1	3,991	40	184
29	401002119530101	97 N27 E19 31CCCC1	3,991	90	208
30	401121119545101	97 N27 E18 24CDDB1	3,995	40	200
31	401138119472301	97 N27 E19 24ADDD1	4,010	168	180
32	401144119494801	97 N27 E19 22ADCA1	4,001	70	199
33	401208119542301	97 N27 E18 13DDBD1	4,063	--	[a]145
34	401216119491001	97 N27 E19 14CACA1	4,005	--	--
35	401235119491601	97 N27 E19 14BDCB	4,031	117	127
36	401410119505701	97 N27 E19 04ACCC1	4,254	--	[a]390
37	401422119474801	97 N27 E19 01ACCC1	4,045	--	--
38	401424119565601	97 N27 E18 03ABAC1	4,110	--	[a]175
39	401528119470501	97 N28 E20 31BACD1	4,178	317	330
40	401105119450301	81 N27 E20 28BBCA1	3,996	--	--

[1] Sites are identified by the standard Geological Survey identification, which is a unique number based on the grid system of latitude and longitude. The number consists of 15 digits: The first 6 denote the degrees, minutes, and seconds of latitude; the next 7 denote degrees, minutes, and seconds of longitude; and the last 2 digits (assigned sequentially) identify the sites within a 1-second grid. For example, site 400435119583201 refers to 40°04'35" latitude and 119°58'32" longitude, and it is the first site recorded in that 1-second grid. If a more precise latitude and longitude subsequently is determined, the initial site-identification number is retained.

[2] Local well numbers are assigned on the basis of the grid system for subdivision of public lands referenced to the Mt. Diablo base line and meridian. For California wells, the first four characters indicate the township, the next four characters indicate the range, the next two characters indicate the section, the letter following the section number indicates a 40-acre subdivision of the section, and the last two digits are assigned sequentially to wells within each 40-acre tract. (For more information on California local well numbers, see Lamb and others [1988, p. 3].) For Nevada wells, each local well number consists of four units: The first unit is the hydrographic area number (Rush, 1968), the second unit is the township, and the third unit is the range. The fourth unit consists of the section number, followed by letters designating the quarter section, quarter-quarter section, and so on, and a number indicating the sequence in which the site was recorded. (For more information on Nevada local well numbers, see Pupacko and others [1988, p. 12].)

[a] Reported depth of well.

A comparison of chemical analyses of water samples from springs and wells in the Winnemucca Ranch area of Warm Springs Valley with samples from the Fish Springs Ranch area in the southeast part of the Honey Lake basin indicate that water-quality characteristics in the two areas are similar. Analytical data were from Washoe County, U.S. Public Health Service, and U.S. Geological Survey records, and from samples collected for this study. Two samples from Fish Springs Ranch wells located near the playa contain a higher proportion of chloride and sulfate ions than the others, a composition that is typical of water derived from lake sediments and concentrated by evaporation. Composition of water from the other wells is similar to that of local surface water, indicating that ground water in both the Fish Springs Ranch and Winnemucca Ranch areas is derived from local precipitation. Different compositions would indicate that the flow systems probably are not connected; similar compositions indicate that either a hydraulic connection or a similar source of recharge exists.

Isotope concentrations can be used as another indicator of ground-water flow paths. The stable isotopes oxygen-18 and deuterium are present as part of the water molecule and can be used as natural tracers of ground water. Concentrations of these isotopes expressed as delta oxygen-18 (the ratio of oxygen-18 to oxygen-16) and delta deuterium (the ratio of deuterium to hydrogen), in the water can indicate source areas and mixing patterns of different waters. Data for samples collected for this study from three streams and three wells in the southeastern part of Honey Lake Valley, and other analyses for the study area reported by Juncal and Bohm (1987, p. 605) and Harding Lawson Associates (1989a, p. 22), show the isotopic composition of the waters to be similar, indicating that these waters probably are from the same or similar sources (James M. Thomas, U.S. Geological Survey, oral communication, 1990). The isotope data also indicate that the waters have undergone some evaporation.

Geothermal water is evidence of deep circulation, but not necessarily of interbasin flow. Faults may impede flow in some places and provide a conduit for flow in other places. Recent reports on geothermal resources of the Wendel and Amedee areas conclude that the geothermal water originates as precipitation in the Diamond Mountains of the Sierra Nevada within the basin, circulates through faulted and fractured bedrock at depths as great as 7,000 feet, and rises along faults (Juncal and Bohm, 1987, p. 605; Harding Lawson Associates, 1989b, p. 10-11). Regional inflow is not considered a source of geothermal water in Honey Lake Valley; however, interbasin flow along other faults in the basin may be possible.

The collective evidence described above indicates that regional inflow is possible, but the evidence is insufficient to confirm or quantify its contribution to ground-water recharge in the basin.

Ground-Water Movement and Storage

Ground water moves from the upland recharge areas to the valley lowlands. It moves toward Honey Lake from volcanic uplands in the northwest, from the Sierra Nevada in the southwest, and from Long Valley through alluvial deposits in the south. Water-level altitudes in wells are used to estimate flow directions. They indicate flow from upland areas toward the valley floor, a slight flow from west to east across the State line, and a local depression in the water table at the playa near Fish Springs Ranch. Water levels also indicate flow from west to east across the east boundary of the study area toward Smoke Creek Desert and Pyramid Lake Valley. Water-level altitudes in the eastern part of the study area are shown in figure 19 in the section "Simulation of ground-water flow." Generalized directions of ground-water flow, on the basis of ground-water levels, are shown in figure 16.

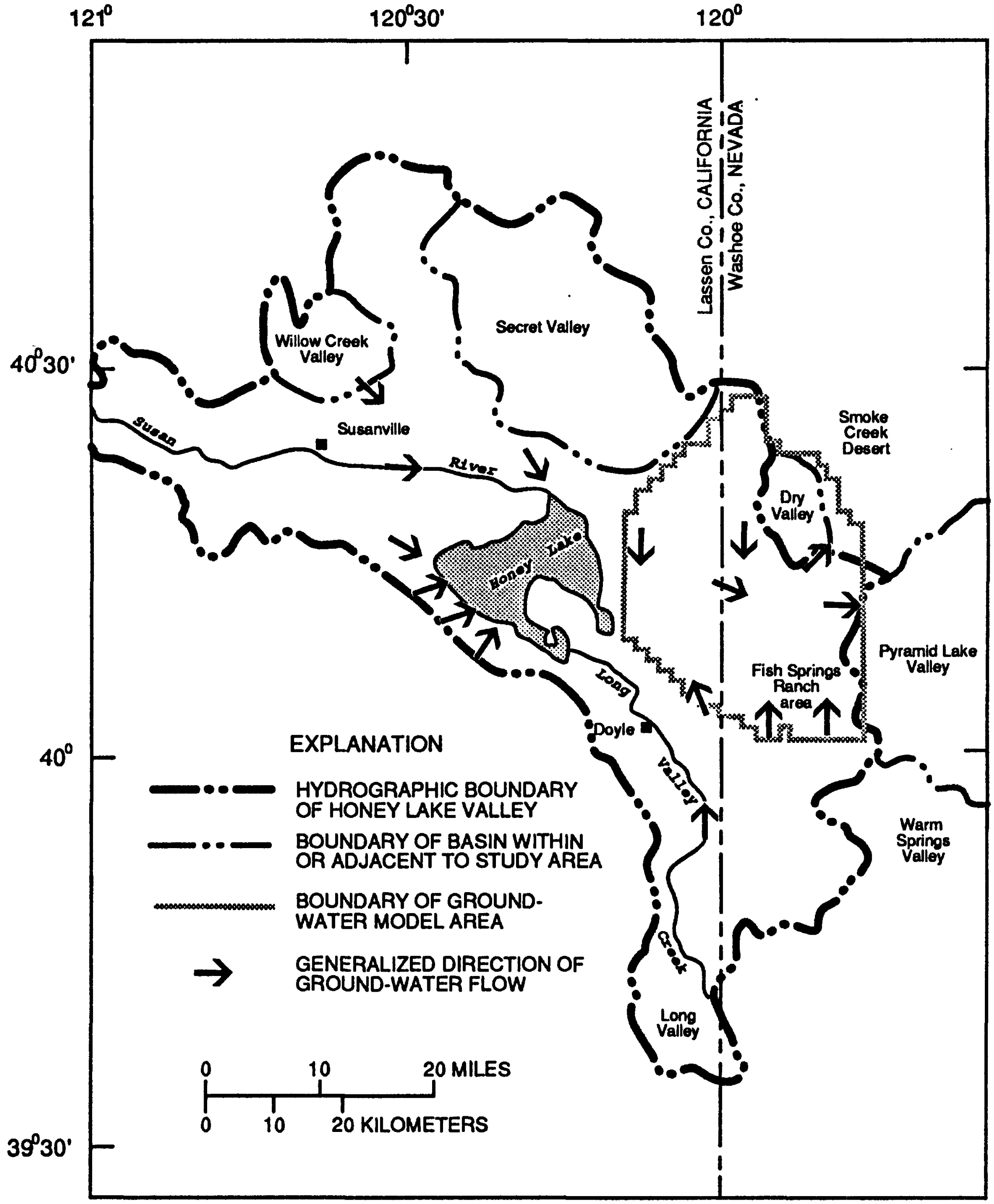


FIGURE 16.--Generalized directions of ground-water flow.

In general, water flows through coarse sand of the alluvial fans and nearshore deposits at the rate of about 1 ft/d, and through lakebed clays at the rate of about 0.02 ft/yr (Heath, 1983, p. 25). Flow through fractured rock can be even faster than through coarse sediments, depending on the size and continuity of the openings in the rock. Nevertheless, hundreds of years may be required for some water to move through an entire ground-water flow system from recharge areas to ultimate discharge.

The total volume of water stored in the upper 100 feet of saturated basin-fill deposits and volcanic-rock aquifers in the study area is an estimated 10 million acre-ft. This is based on the product of the specific yield, area, and thickness of each unit. Not all of this water is economically recoverable or of acceptable quality for practical use.

Changes in ground-water levels in wells indicate changes in storage caused by increases or decreases in recharge, natural discharge, or withdrawals. The California Department of Water Resources and the U.S. Soil Conservation Service have measured water levels in a network of irrigation wells in Honey Lake Valley, before and after the irrigation season, to monitor long-term trends. The network began with 4 wells in the fall of 1972, and included about 50 wells by 1988. The network area was expanded in 1988 to include wells in Willow Creek, Secret, and Long Valleys, which are tributary to the floor of Honey Lake Valley (see figure 1). Hydrographs of spring water levels in selected wells in the study area (figure 17) indicate a general increase in water levels during the wet period 1981-83 (wells 11, 12, and 13) and water-level declines during the dry periods 1974-77 (well 12) and 1987-89 (wells 10, 12, and 13). The wet and dry years are shown in figure 7.

In the eastern part of the basin, water levels were measured weekly in 7 wells and monthly in 17 wells from March 1987 to April 1989. Water levels were measured for shorter periods in several additional wells, including observation wells drilled during 1988 for this study. Hydrographs of water levels in figure 18 show small seasonal effects superimposed on slightly declining water levels in basin-fill deposits (wells 5, 22, and 28), and larger seasonal variations superimposed on greater water-level declines in wells in volcanic rocks (wells 16, 17, and 18). Wells 17 and 18 are pumped for irrigation; well 16 is not pumped but water levels have responded to pumping at other wells more than half a mile away. Well 22 is completed in basin-fill deposits and is close to the wells that pump from volcanic-rock aquifers. Despite this, the water level in well 22 does not exhibit the pronounced seasonal fluctuations shown in wells 16, 17, and 18 (figure 18). Instead, the available water-level measurements indicate a slight but relatively constant decline. This might indicate that pumping from the volcanic-rock aquifers may be inducing a gradual drainage of water stored in adjacent basin-fill aquifers. Most of the general decline in the vicinity of Fish Springs Ranch probably results from ground-water withdrawals. However, the monitoring started in 1987 at the beginning of a dry period after several years of above-average precipitation, and part of the decline also may result from variations in recharge associated with climatic fluctuations. Elsewhere in the basin, where the California Department of Water Resources has been monitoring water levels since 1972, water levels in some wells declined during the 1976-81 drought period, subsequently recovered, and have declined slightly since 1987 (figure 17, wells 12 and 13). Several more years of data are needed to accurately estimate annual rates of decline and the magnitude of seasonal fluctuations in the eastern part of the basin.

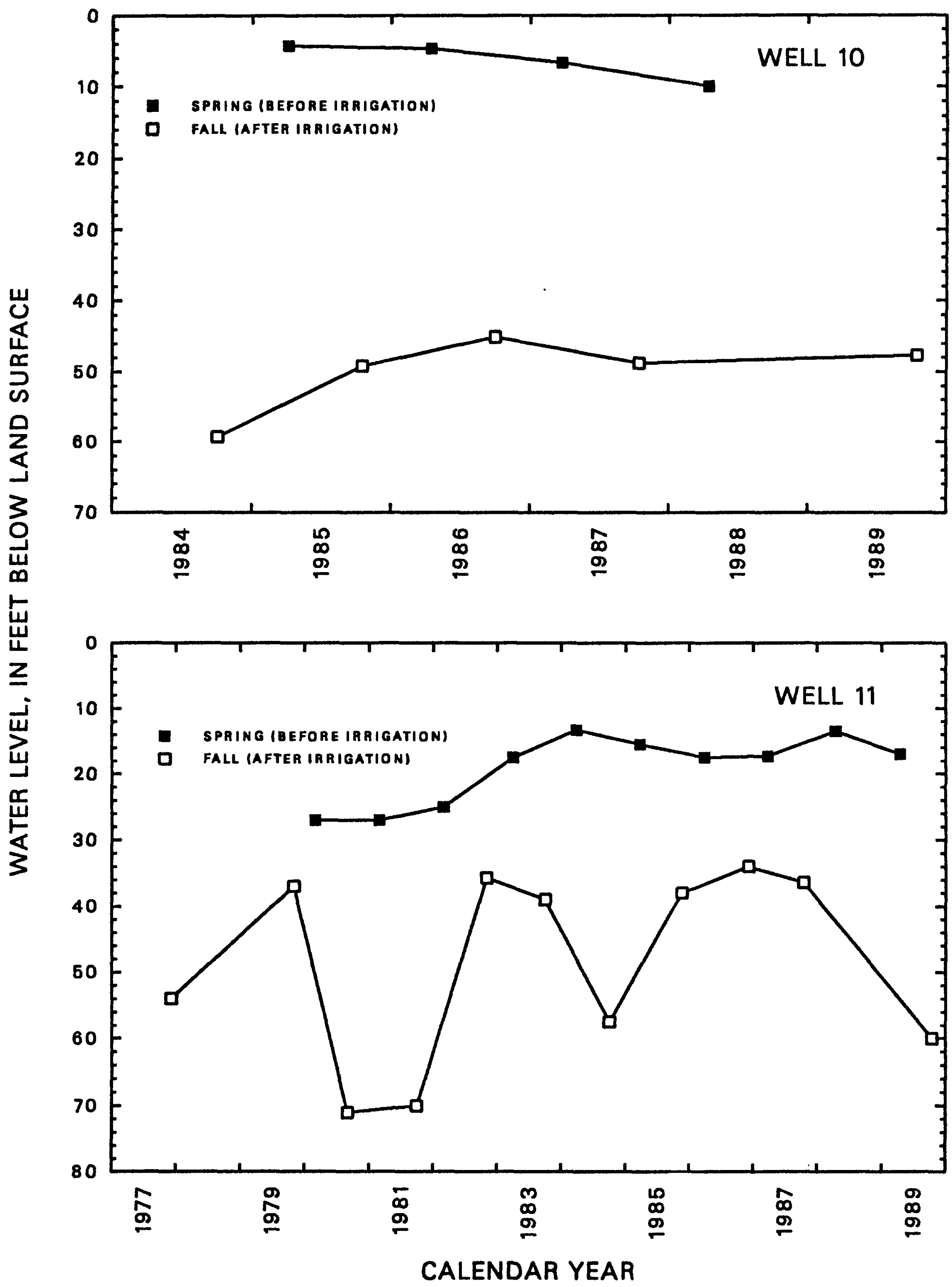


FIGURE 17.--Water levels in four selected wells before and after irrigation season (well locations are shown on plate 1; well data are listed in table 10).

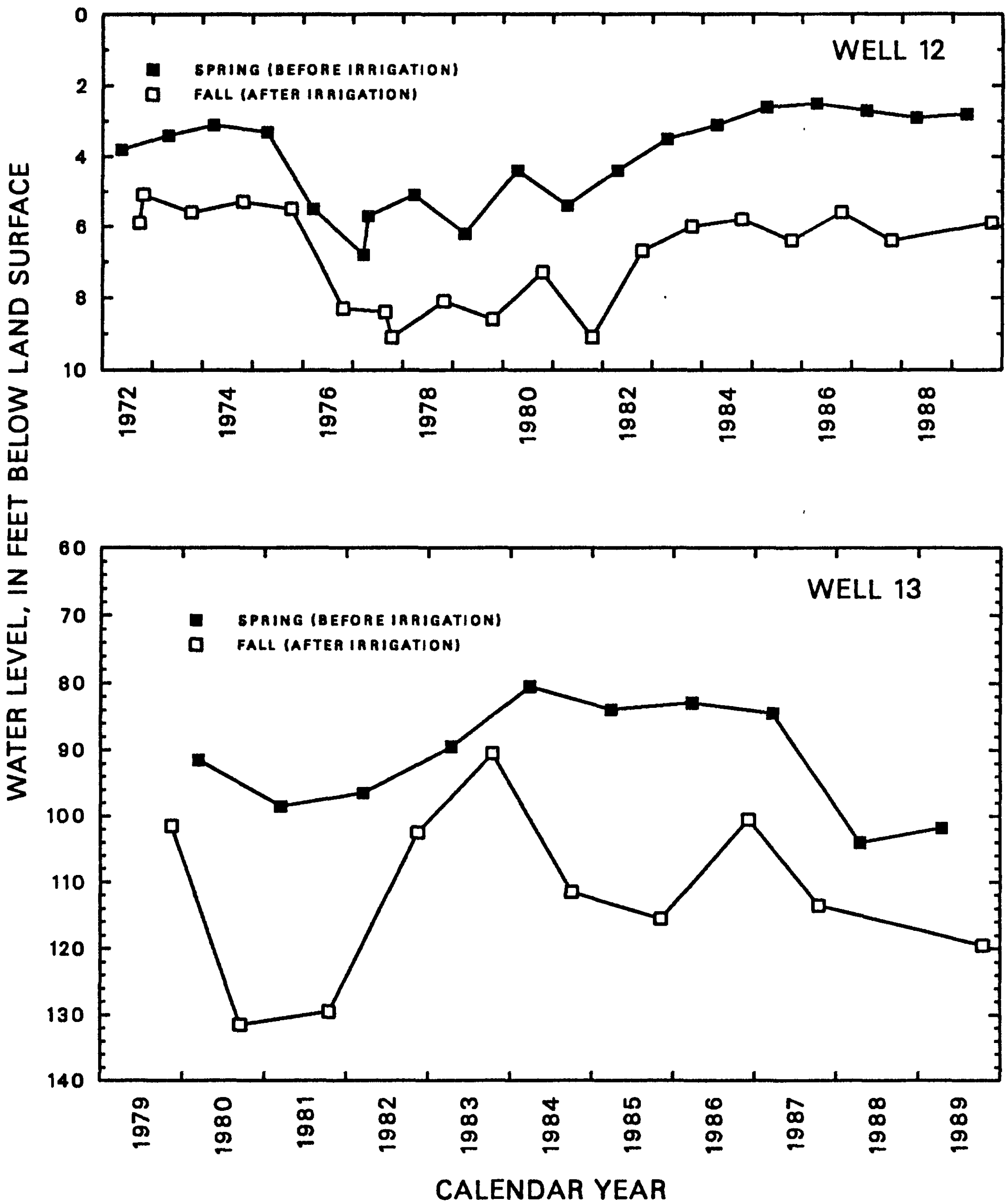


FIGURE 17.--Continued.

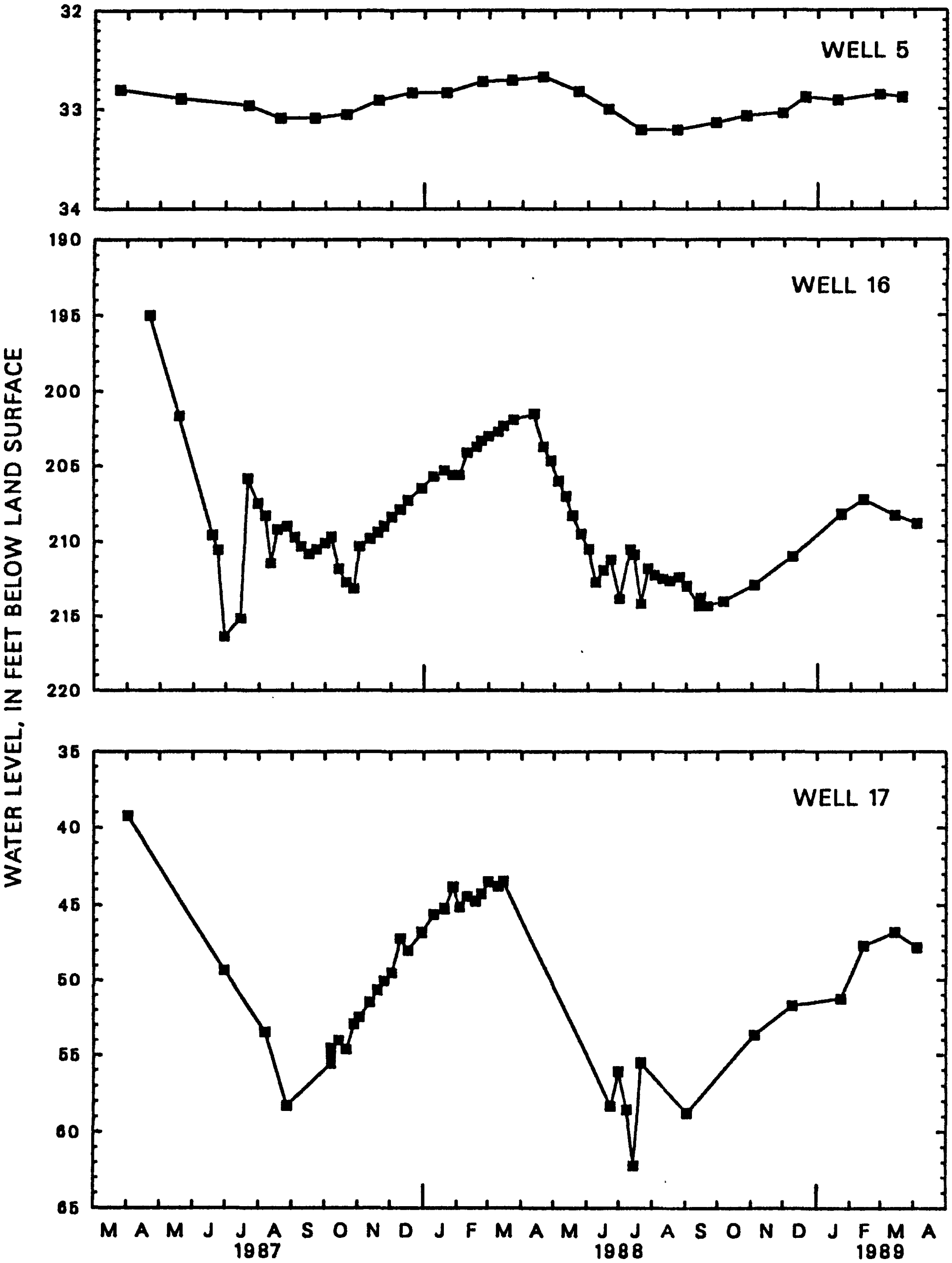


FIGURE 18.--Water levels in six wells in eastern Honey Lake Valley, 1987-89 (well locations are shown on plate 1; well data are listed in table 10).

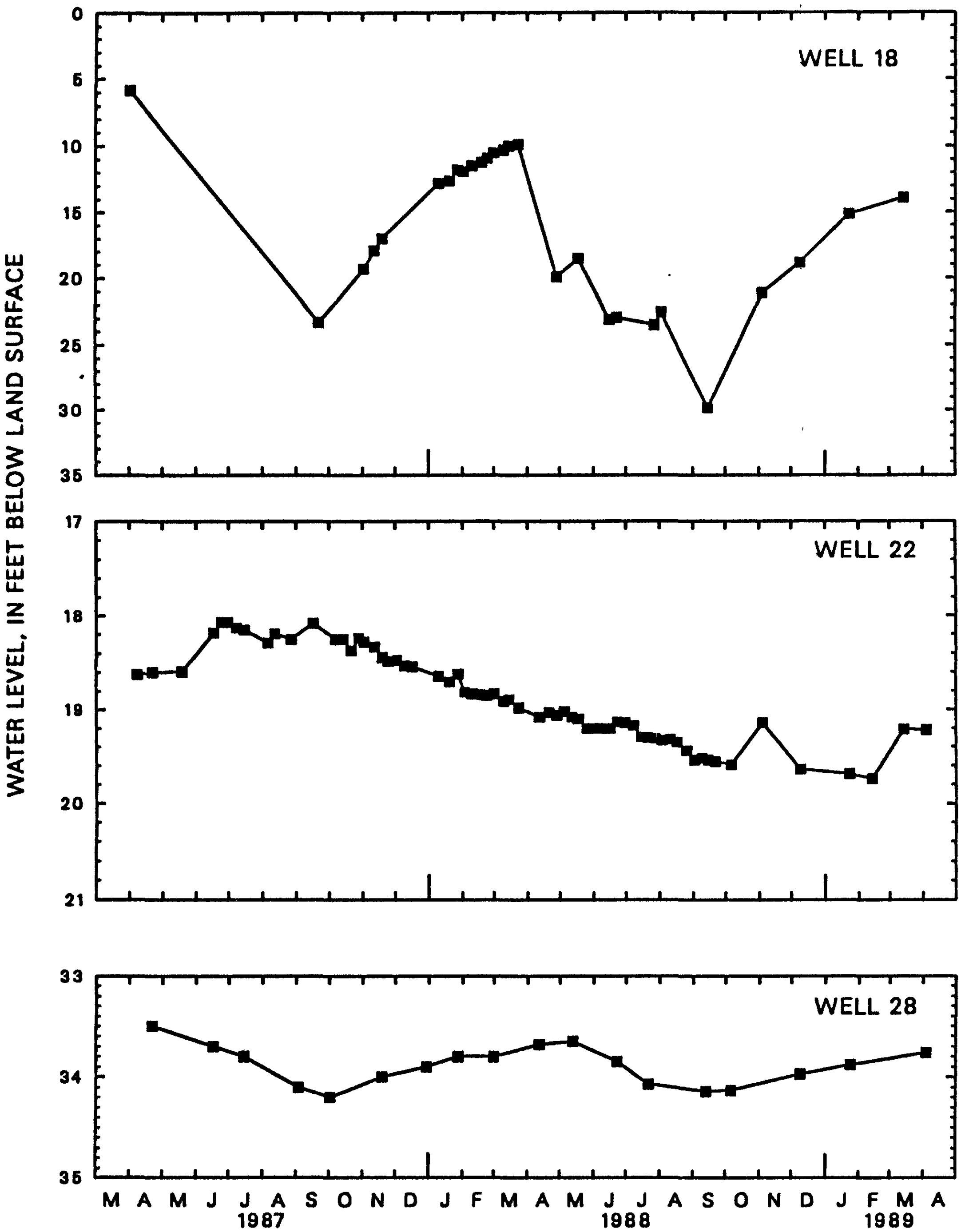


FIGURE 18.--Continued

Discharge

Under natural conditions, ground water discharges from the basin by (1) evaporation from soils and transpiration by plants, (2) seepage to and evaporation from Honey Lake, and (3) subsurface outflow. Ground water that is discharged by springs either is consumed by evapotranspiration or infiltrates back into the ground. No surface water flows from the basin.

Under 1988 conditions of development, water pumped from wells is a major component of discharge from the basin. Annually, about 5,900 acre-ft is withdrawn from the study area in Nevada and about 47,000 acre-ft is withdrawn in California (table 8). About 75 percent of the water pumped from irrigation wells evapotranspires from cultivated fields. Most of the geothermal ground water pumped at Amedee and Wendel is discharged to Honey Lake and eventually evaporates.

Evapotranspiration from Ground Water

Most precipitation evaporates (or sublimates) from the land surface or is transpired from soil moisture by shallow-rooted plants. On the basis of results of this study, about 89 percent of the total precipitation and stream inflow to the study area eventually is discharged by evapotranspiration from the land surface, streams, and surface-water irrigation, or from Honey lake. The remaining 11 percent evapotranspires from ground water or may discharge from the basin by subsurface outflow. Evapotranspiration rates are related to daily and seasonal cycles; they respond to changes in air temperature, solar radiation, wind speed, and soil moisture.

In a natural (undeveloped) system, ground water is discharged to the atmosphere by two mechanisms: (1) direct evaporation from the water table through surface sediments and (2) transpiration by phreatophytes, plants that extend their roots to the water table to obtain water. Direct evaporation occurs only where the water table is less than a few feet below land surface; the rate averages 0.1 to 0.2 ft/yr from bare soil on the basis of the estimated vertical hydraulic conductivity of sediments and the depth to water. Transpiration rates depend on the type and density of phreatophytes, climatic conditions, depth to water, and quality of ground water (Robinson 1958, p. 16). The most common phreatophytes in Honey Lake Valley are greasewood, rabbitbrush, and saltgrass. Ranges of evapotranspiration rates are estimated from results of research in other areas where these plants are common (Lee, 1912; White, 1932; Young and Blaney, 1942; Houston, 1950; Robinson, 1970). Estimates of annual ground-water evapotranspiration rates range from 0.2 foot for sparse greasewood to 2 feet for wet meadows. These rates include only evaporation and transpiration from the saturated zone. Evapotranspiration of water from the unsaturated zone (soil moisture) was discussed in the section "Infiltration of precipitation." Two sets of evapotranspiration measurements were made in each of two locations during the summer of 1988 as part of this study. One setting was a stand of mixed phreatophytes, mainly greasewood, near Fish Springs Ranch; the other was an area of crested wheat grass at the Fleming Wildlife Refuge. Evapotranspiration at these sites was from ground water, as no precipitation had fallen for several weeks preceding the test periods (July 29-August 4 and September 10-15) and soil moisture therefore was depleted. Results indicated slightly greater evapotranspiration from the native grass area (William D. Nichols, U.S. Geological Survey, written communication, 1988). Average rates during the first period were 0.08 in/d for the Fish Springs Ranch site and 0.09 in/d for the Fleming site. Average rates during the second period were 0.01 in/d for the Fish Springs Ranch site and 0.09 in/d for the Fleming site.

Evapotranspiration from ground water in the study area was estimated by determining phreatophyte distribution, primarily from Landsat imagery of September 2, 1980, and applying estimated phreatophyte evapotranspiration rates (table 11). The Landsat results were divided into seven Level-I classes (modified from the classification of Anderson and others [1976, table 2]). The rangeland class was reclassified to identify potential phreatophyte zones, which were field checked. About 77,000 acres were classified as probable phreatophytic areas, 35,000 acres as sparsely vegetated or barren (playa) areas, and 52,000 acres as native grass and cropland, including wetlands (table 11; J. LaRue Smith, U.S. Geological Survey, written communication, 1988), for a total of about 160,000 acres (plate 4). Native grass and cropland areas were grouped together because their characteristics overlap; the acreage for this group includes irrigated areas. However, areas irrigated with surface water do not contribute to ground-water evapotranspiration and areas irrigated with ground water use water that is included in the water budget as withdrawal from wells. To determine the areas from which ground water is directly discharged by evapotranspiration, irrigated areas were estimated separately and subtracted from the group.

TABLE 11.--*Estimated evapotranspiration of ground water in the study area*

Vegetation group	Area, rounded[a] (acres)	Annual evapotranspiration: Range[b] (feet)	Rate used (feet)	Acre-feet, rounded
Mixed phreatophytes				
western part of study area	51,000	0.2-0.5	0.4	20,000
eastern part (table 15)	26,000	0.2-0.4	.3	7,800
Bare playa and sparse phreatophytes	35,000	0.1-0.2	.1	3,500
Native grass and cropland[c]	[d]52,000	1.5-2.0	1.8	[e]94,000
Total	[d]160,000	--	--	[e]130,000
Total from native vegetation	140,000	--	--	85,000

[a] Based on analysis of Landsat data for September 1980 and limited field checking by J. LaRue Smith (U.S. Geological Survey, written communication, 1988). Distribution of phreatophytic vegetation is shown on plate 4.

[b] Rates are based on data from work in Honey Lake Valley by California Department of Fish and Game (written communication, 1988), William D. Nichols (U.S. Geological Survey, written communication, 1988), and work in other areas by Lee (1912), White (1932), Young and Blaney (1942), Houston (1950), Robinson (1970), and Patrick A. Glancy and James R. Harrill (U.S. Geological Survey, oral communication, 1990).

[c] Native-grass areas include some wetlands.

[d] Includes cropland irrigated with surface water and ground water in California.

[e] Includes evapotranspiration of some irrigation water in California.

Estimates of irrigated areas in the California part of Honey Lake Valley, excluding Long Valley south of the study-area boundary, range from 22,000 acres for 1985 (William E. Templin, U.S. Geological Survey, written communication, 1988) to 39,000 acres for 1986 (Clements, 1988, p. 6). The larger estimate probably is more accurate because it is based on a detailed land-use inventory. However, on the basis of the location and distribution of cropland under natural (nonirrigated) conditions, a significant part of the irrigated area would support native grasses and other phreatophytes. Therefore, the smaller estimate of irrigated cropland was used for these calculations. Assuming that irrigation in the California part of the basin was about the same in 1980 (the date of the Landsat imagery used for this study) as in 1985, and that irrigation in the Nevada part of the basin was negligible during 1980, the native-grass areas would total about 30,000 acres (52,000 minus 22,000). On the basis of estimated rates of evapotranspiration, annual ground-water evapotranspiration from these native-grass areas is about 54,000 acre-ft.

Evapotranspiration directly from ground water is likely to occur from about 140,000 acres in the basin. Annual evapotranspiration from ground water in these areas is estimated at 85,000 acre-ft (table 11).

Seepage to Honey Lake and to Streams

Observations of the lake bottom during 1976-77 and other dry periods indicate that a few seeps and springs discharge ground-water to the lake through the lakebed (Robert Anton, U.S. Soil Conservation Service, oral communication, 1987). However, the volume of water derived from these sources is small compared to the inflow of surface water during average years.

Measurements of streamflow along reaches of five streams in the basin in December 1987 (Rockwell, in press) indicate that some reaches gain water at some times. The gains typically result from discharge by springs into the stream channel. Some of this water seeps back into the ground through stream-bottom sediments downstream from the spring and re-enters the ground-water flow system. The rest is included in the water budget as evapotranspiration.

Subsurface Outflow

Ground-water levels in Smoke Creek Desert to the northeast and the water level in Pyramid Lake to the east are lower than those in Honey Lake Valley (Bedinger and others, 1984, sheet 1). To determine whether a hydrologic connection exists between Honey Lake basin and these two basins, observation wells were installed just west of Sand Pass and Astor Pass (well 35), in Sand Pass (well 39), and in Astor Pass (well 31; plate 1) during this study. Monthly water-level measurements in these wells and in existing stock wells in Sand Pass (well 37) and the eastern end of Astor Pass (well 40) were compared to determine the probable direction of ground-water flow through these passes. Water-level altitudes in wells 35 (3,962 feet), 37 (3,948 feet), and 39 (3,928 feet) through Sand Pass, and water-level altitudes in wells 35 (3,965 feet), 31 (3,950 feet), and 40 (3,901 feet) through Astor Pass, indicate that the topographic divides across these passes do not correspond to ground-water divides and the hydraulic gradient is from Honey Lake Valley toward the northeast and east. Discharge from springs and flowing wells in the southwest part of Smoke Creek Desert appears to be large with respect to the probable source area. The wells and springs may be discharging ground-water inflow from Honey Lake Valley. Stable isotopes in water samples from the flowing wells in Smoke Creek Desert and from wells and springs in Honey Lake Valley indicate that the two groups have the same source or a similar source (James M. Thomas, U.S. Geological Survey, oral communication, 1988).

Water Discharged from Wells

Estimates of the average volume of ground water withdrawn for irrigation each year range from about 20,000 acre-ft to about 50,000 acre-ft. The smaller volume is based on a 1985 water-use estimate of 14,000 acre-ft in the California part of the basin (William E. Templin, U.S. Geological Survey, written communication, 1988) plus about 5,900 acre-ft determined from flow-meter readings and irrigated acreage in the Nevada part of the basin. The larger volume is reported by Walters Engineering (1986, table 2B). Another estimate was derived using data from Clements (1988, p. 8) who reported that evapotranspiration from irrigation by ground water in 1986 was 32,000 acre-ft on the basis of a detailed computation of crop use. By using Clements' estimate and assuming that evapotranspiration is 75 percent of the irrigation water withdrawn, then 1986 withdrawals were about 43,000 acre-ft and about 25 percent of that total (11,000 acre-ft) infiltrated back to the ground-water flow system. The infiltration estimate is based on average values for return flows for the western United States (Lauritzen and Terrell, 1967, p. 1105).

The amount of ground water withdrawn annually for domestic, industrial, power-generation, and stock use (table 8), estimated as 10,000 acre-ft, is small in comparison to the amount for irrigation use, and is assumed to be a negligible part of the basin-wide ground-water budget.

GROUND-WATER BUDGET

Ground-water recharge and discharge in the study area are summarized in table 12. Each component of total recharge and total discharge is estimated independently, and is subject to at least some uncertainty. The indicated imbalance between rounded totals for recharge and discharge may be partly a result of (1) possible ground-water flow into and out of the study area, (2) the uncertainty of the component estimates, and (3) the net effect of rounding. The difference between total recharge and total discharge is considered negligible.

Ground-water recharge from infiltration of precipitation, snowmelt, and soil moisture accounts for about 40 percent of the total recharge. Infiltration of streamflow, primarily through streambeds on alluvial fans, accounts for about 40 percent, and infiltration of surface-water and ground-water irrigation flow, primarily on lower fans and the valley floor, accounts for about 20 percent of the ground-water recharge.

Ground-water evapotranspiration from soil and native vegetation accounts for more than 60 percent of total discharge. Withdrawals from wells for irrigation and other uses, and possibly subsurface outflow, accounts for the remainder of the discharge.

Recharge estimates are based in part on results of the Deep Percolation Model (DPM). The DPM may under estimate recharge because it calculates a water budget for the soil zone, and does not consider the effects of fractured and jointed volcanic rocks that may increase infiltration rates in upland areas where soil is thin. This may be why the DPM seems to overestimate runoff from the Willow, Skedaddle, and Cottonwood Creek drainages (table 5), all of which are in volcanic terrane.

The ground-water budget is based partly on a water budget for Honey Lake. The annual change in volume of Honey Lake is assumed be the difference between inflow and outflow. The major components of inflow are precipitation and streamflow. Outflow is by evaporation from the lake surface. The lake budget is based on the assumption that ground-water discharge to the lake and recharge from the lake are minor.

The most uncertain components of the ground-water budget are subsurface inflow and outflow. Although water-level measurements and the prevalence of fractured volcanic rocks in the east and southeast part of the basin indicate that ground-water outflow from Honey Lake Valley is likely, and ground-water inflow is possible, the volumes and rates are difficult to determine. The ground-water flow model, discussed in the following section of this report, was used to help quantify these components of the budget.

SIMULATION OF GROUND-WATER FLOW

A mathematical model to simulate ground-water flow within an aquifer system consists of a set of differential equations that represents ground-water flow (Wang and Anderson, 1982). Ground-water flow is simulated by simultaneously solving the differential equations with a computer. A mathematical model is useful for evaluating and refining the understanding of an aquifer system and also for predicting aquifer responses to various applied stresses. However, a model is only a simplification of the actual system being simulated; it cannot totally duplicate the actual system because it must be based on average conditions where data are available, or on estimated conditions where sufficient data are not available, and on several simplifying assumptions.

The basic mathematical model applied in this study uses a finite-difference solution technique and is commonly referred to as the "USGS modular model." The theoretical development, numerical-solution techniques, computer code, and data requirements of the model are described by McDonald and Harbaugh (1988).

TABLE 12.--*Ground-water budget for the Honey Lake Valley study area*

[Acre-feet per year, rounded to two significant figures]

Budget component	Estimated quantity
RECHARGE:	
Infiltration of precipitation	55,000
Infiltration of streamflow:	
from north and east	13,000
from south, west, and northwest	40,000
Irrigation return:	
from surface-water irrigation	14,000
from ground-water irrigation	11,000
Subsurface inflow[1]	unknown
Total recharge	130,000
DISCHARGE:	
Evapotranspiration from soil and native vegetation	85,000
Withdrawals by wells	53,000
Subsurface outflow[2]	unknown
Total discharge[3]	140,000

[1]Results of a ground-water flow model of the eastern part of the study area indicate that annual recharge may be 5,000 acre-feet greater than estimated, possibly as a result of subsurface inflow.

[2]Results of the ground-water flow model indicate about 7,000 acre-feet of subsurface outflow across the eastern boundary of the model area.

[3]The imbalance between total recharge and total discharge may be partly a result of: possible subsurface flow into and out of the study area; the uncertainty of component estimates; and the net effect of rounding.

The area modeled for this study is about 452 mi^2. It includes that part of Honey Lake Valley east of Honey Lake, excluding the Long Valley Creek drainage area (figure 1), and includes Dry Valley northeast of the basin and a small part of Smoke Creek Desert. This area was selected for flow modeling because proposed ground-water development in the vicinity of Fish Springs Ranch has created the need for a better understanding of that part of the aquifer system. The eastern part of Honey Lake Valley (called "Calneva subbasin") also was identified by the California Department of Water Resources as a part of the basin needing further study (Pearson, 1987, p. 8).

To numerically define the aquifer system, it was necessary to determine the boundary conditions for the system, identify the aquifer properties within the modeled area, and estimate the rates and distribution of recharge and discharge in the aquifer system. The accuracy of the model depends on the accuracy of these estimates.

Model Calibration

The model was calibrated to the steady-state conditions that were generally assumed to be represented by conditions within the flow-model area during the spring of 1988. Steady-state conditions describe a system in equilibrium; inflows and outflows are equal and the volume of water in storage does not change. Steady-state water levels depend on the quantities of recharge to and discharge from the ground-water system, the hydraulic conductivity of the aquifer materials, and the leakance between layers. The storage component of the system was not modeled because storage does not change under steady-state conditions.

The calibration of a ground-water-flow model requires the trial and error process of adjusting initial estimates of aquifer properties and stresses (within acceptable ranges) to obtain the best match between simulated and measured water levels. The initial values and boundary conditions are adjusted within ranges on the basis of the limits of known geologic and hydrologic properties of the basin and the degree of confidence in the original data estimates. The model is considered to be calibrated when the simulated water levels are within an acceptable range (in this case plus or minus about 5 feet) and the simulated stresses match independent estimates. The root-mean-square deviation (Hoxie, 1977, p. 27), which gives an indication of the difference between two sets of values, was used to determine the closeness of the match between measured and simulated hydraulic heads. The root-mean-square deviation (or error) is calculated using the equation:

$$RMSD = \sqrt{\frac{\Sigma (M - C)^2}{N}}$$

where RMSD = root-mean-square deviation, in feet,
M = measured water level, in feet,
C = hydraulic head computed by model, in feet, and
N = number of water-level measurements.

During 1988, most of the ground-water development in the flow-model area, except in the vicinity of Fish Springs Ranch, was limited to a few scattered, low-yielding domestic and stock wells. Total withdrawals from these wells were estimated to be about 50 acre-ft/yr by Rush and Glancy (1967, table 18); withdrawals probably have not changed much since then, as the population in the Nevada part of the basin in 1988 was only about 25 people. Domestic and stock-well withdrawals would have little effect on the ground-water system. However, in the Fish Springs Ranch area, several high-yield irrigation wells have been put into production since 1980; the ground-water system in this area may not have reached a new equilibrium. Water-level declines observed during 1987-89 (figure 18) may be the result of withdrawals from the irrigation wells, the result of less than normal recharge during that dry period (figure 7), or both.

Figure 19 shows the locations of, and water-level altitudes for, wells that were used for comparison with simulated heads to calibrate the flow model. These 31 wells were selected for use in calibration based on their location, depth, and accessibility for water-level measurements during 1988. Accurate land-surface and water-level altitudes were determined at these sites by surveying.

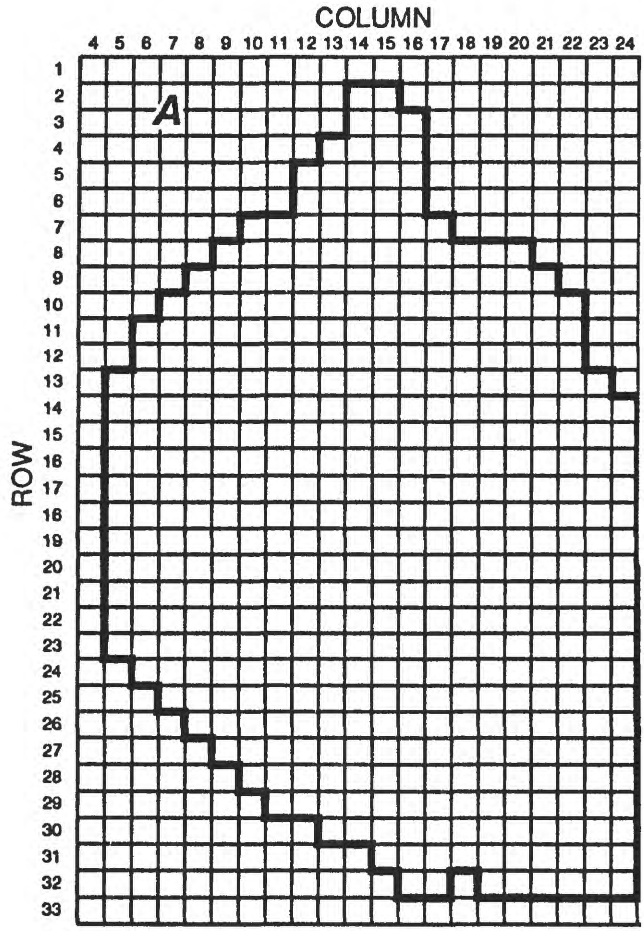
General Features of the Flow Model

For a finite-difference model, the aquifer system is divided into horizontal layers; a rectangular grid divides the layers into rows and columns. Each cell in the grid represents a three-dimensional block of the aquifer; the center point of the block is called a node. The grid is overlain on maps that show the areal distribution of ground-water levels, aquifer properties, and stresses for each layer. The average value of each property (or stress) within a cell is determined from the map and entered into the model to represent the value of that property for the entire cell. The process is repeated until a value for each property has been assigned to every cell in the modeled area.

Model Grid and Layers

The finite-difference grid designed for the model used in this study is regularly spaced and contains 24 columns by 36 rows; each grid cell is 1 mile on a side (figure 19). Columns 1-3 and rows 34-36 consist of inactive cells and are not shown on figure 19. The grid is oriented north-south to be parallel to the Nevada-California State line and to coincide with part of the Deep Percolation Model (DPM) grid used for recharge estimates. The grid network is used for each of four horizontal layers in the model. This four-layer model consists of 1,602 active cells.

Layer 1, the upper layer, extends from the water table down to an altitude of 3,700 feet above sea level (figure 20). Thickness ranges from about 260 feet at the center of the basin to about 350 feet in the Skedaddle Mountains and at the southern boundary. Layer 1 contains most of the wells in the modeled area. Layer 2 extends downward from an altitude of 3,700 feet to 3,000 feet. A few deep wells pump water from this layer. Layer 3 extends downward from an altitude of 3,000 feet to 1,500 feet, and layer 4 extends from an altitude of 1,500 feet to the granitic bedrock. The maximum thickness of layer 4 is approximately 2,000 feet. No wells withdraw water from layers 3 and 4. Cross-sections of the modeled area show the layers from south to north (figure 20, B-B' and C-C') and from west to east (figure 20, D-D').



EXPLANATION FOR PART *A*

☐ MODEL CELL -- Area, 1 square mile

▬ MODEL-AREA BOUNDARY

EXPLANATION FOR PART *B*

☐ MODEL CELL -- Area, 1 square mile

LAYER 1 -- Shallowest layer; shown only where layer 1 extends beyond layer 2

LAYER 2 -- Shown only where it extends beyond layer 3

LAYER 3 -- Shown only where it extends beyond layer 4

LAYER 4 -- Deepest layer; shows entire areal extent of layer 4

GENERAL-HEAD BOUNDARY -- All layers

NOTE: Each layer is surrounded by a no-flow boundary except where general-head boundary is indicated.

● 3961 OBSERVATION WELL, LAYER 1 -- Number indicates measured water-level altitude, 1988, in feet above sea level

○ 3986 OBSERVATION WELL, LAYER 2 -- Number indicates measured water-level altitude, 1988, in feet above sea level

FIGURE 19.--Grid, boundaries, and observation wells used in the ground-water flow model, eastern Honey Lake Valley and adjacent areas.

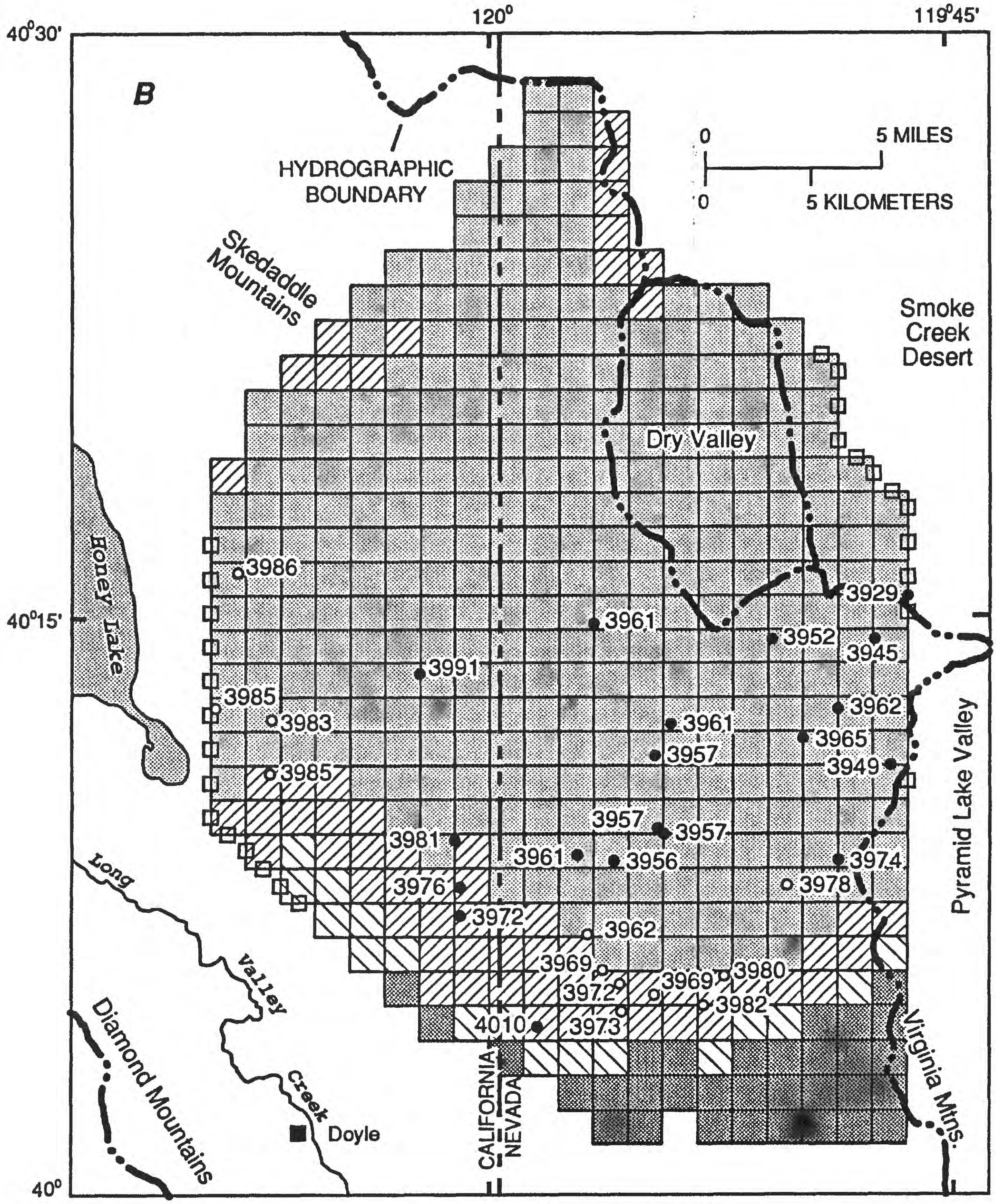


FIGURE 19.--Continued.

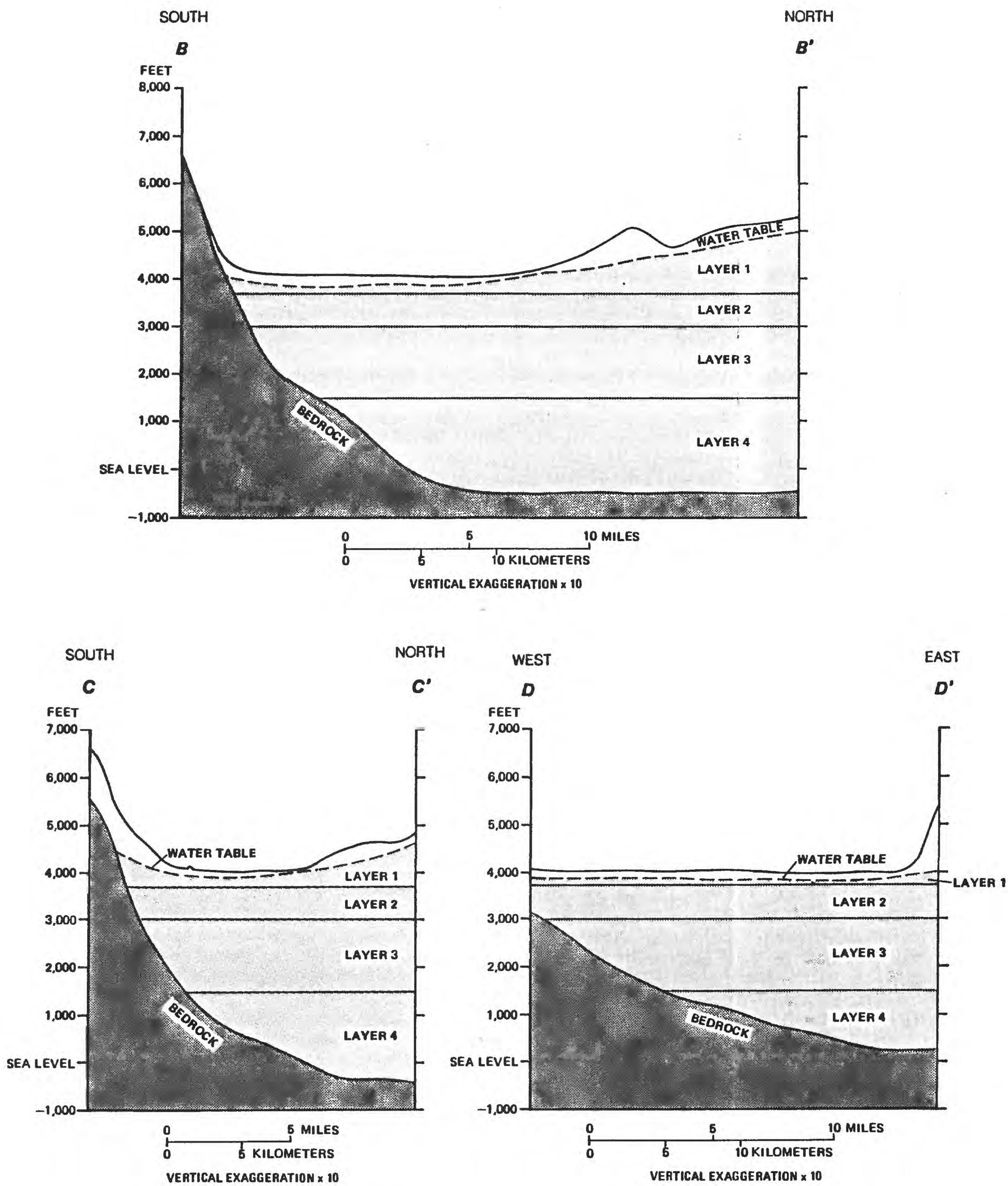


FIGURE 20.--Generalized hydrogeologic sections showing configuration of model layers in eastern Honey Lake Valley (lines of section are shown on plate 2).

Model Boundaries

The top boundary of the model is the water table. It is simulated as a free-surface boundary that is allowed to move vertically in response to imbalances between inflow and outflow. The bottom boundary of the model is the contact between the bottom of the aquifer materials and granitic bedrock, which is assumed to be impermeable.

The lateral model boundaries generally coincide with topographic divides or with contacts between aquifer materials and impermeable bedrock, except along the western and southwestern edges of the flow-model area. At the western edge, a north-south general-head boundary was placed about 1 mile east of Honey Lake; at the southwestern edge, the general-head boundary is along a diagonal line from Turtle Mountain, at the northern tip of the Fort Sage Mountains, to Honey Lake (figure 19 and plate 1). These boundaries were chosen because water levels in monitoring wells 5, 6, 7, and 9 (plate 1 and table 10) are stable and indicate the area is beyond the effects of existing development. The lateral boundaries are the same in all four layers except in the southeast and the north, where deeper layers are less extensive because depth to bedrock decreases. All lateral boundaries were simulated as either no-flow or general-head boundaries, as shown in figure 19.

No-flow boundaries were used in the northern part of the model to simulate the Skedaddle Creek drainage divide; along the east side, south of Astor Pass, to simulate the drainage divide between Honey Lake Valley and Pyramid Lake; in the south, to simulate the drainage divide between Honey Lake Valley and Dry Valley; and along part of the southwest boundary, to represent the northeast side of the granitic rocks of the Fort Sage Mountains.

General-head boundaries were used to simulate (1) the western model boundary east of Honey Lake and part of the southwest boundary along the inferred extension of the Warm Springs fault, (2) the northeastern boundary between Honey Lake Valley and Smoke Creek Desert, and (3) part of the eastern boundary (at Astor Pass) between Honey Lake Valley and Pyramid Lake Valley (figure 19 and plate 1). A general-head boundary simulates connection to an aquifer outside the model area and indicates a water source or sink, supplying water to or receiving water from adjacent model blocks at a rate proportional to the difference in hydraulic head between the outside source or sink and the model block (McDonald and Harbaugh, 1988, p. 11-1). The flow also depends on the conductance of the materials between the external source or sink and the boundary cell in the model. Conductance is defined as the horizontal hydraulic conductivity of the cell, times the vertical cross-sectional area of the cell, divided by the distance to the source or sink. The initial values of conductance for general-head boundary cells were adjusted during calibration of the model. Characteristics of general-head boundary cells are shown in table 13.

Springs observed in Honey Lake during periods of low lake levels, numerous springs and flowing wells in the southwest corner of Smoke Creek Desert east of Sand Pass, and springs near the Pyramid Lake shoreline east of Astor Pass, are indications of ground-water discharge. Heads are assumed to increase with depth in these areas.

Along the western boundary of the ground-water flow model, the distance from the edge of the model to the outside source or sink (Honey Lake) is about 1 mile. Honey Lake and the saturated sediments beneath the lake provide water to or receive water from the modeled area. The general-head altitude for this boundary was set at 3,986 feet (the approximate mean altitude of the water surface of Honey Lake during the study period) for layer 1; at 3,988 feet for layer 2; at 3,990 feet for layer 3; and at 3,992 feet for layer 4 (figure 20). Conductance values were 375 ft^2/d (layer 1), 750 ft^2/d (layer 2), 940 ft^2/d (layer 3), and 1,440 ft^2/d (layer 4) in the calibrated model. Conductance increases with depth along this boundary because the deeper layers are thicker.

TABLE 13.--*Characteristics of general-head boundary cells in the calibrated ground-water flow model*

Boundary cell(s)[1]			Minimum distance to source or sink (miles)	General-head altitude (feet above sea level)	Conductance[2] (feet squared per day)
Layer	Row(s)	Column			
			Western boundary.		
1	15-23	5	1	3,986	375
2	15-23	5	1	3,988	750
3	15-23	5	1	3,990	940
4	15-22	5	1	3,992	1,440
			Southwestern boundary		
1	24	6	3	4,050	85
2	24	6	3	4,050	200
3	24	6	3	4,050	100
1	25	7	3	4,050	85
2	25	7	3	4,050	50
			Northeastern boundary		
1	10-12	22	5	3,850	700
2	10-12	22	5	3,852	175
3	10-12	22	5	3,854	100
4	10-12	22	5	3,856	100
1	13	23	5	3,850	700
2	13	23	5	3,852	175
3	13	23	5	8,854	100
4	13	23	5	3,856	100
1	14-17	24	5	3,850	700
2	14-17	24	5	3,852	175
3	14-17	24	5	3,854	100
4	14-17	24	5	3,856	100
			Eastern boundary at Astor Pass		
1	22	24	5	3,792	700
2	22	24	5	3,794	175
3	22	24	5	3,796	100
4	22	24	5	3,798	100

[1] Figure 19.

[2] Horizontal hydraulic conductivity of the cell, multiplied by the vertical cross-sectional area of the cell, and divided by the distance to the source or sink.

Along the southwestern boundary between Long Valley and the modeled area, the general-head altitude was set at 4,050 feet on the basis of the measured water level in a well located near Long Valley Creek; this value is the same for each layer. The model is three layers thick at one general-head-boundary cell and two layers thick at the other general-head-boundary cell at this edge of the model because the deeper layers are less extensive (figure 19). Conductance values at the three-layer cell were 85 ft^2/d (layer 1), 200 ft^2/d (layer 2), and 100 ft^2/d (layer 3); conductances at the two-layer site were 85 ft^2/d (layer 1) and 50 ft^2/d (layer 2).

At the northeastern boundary between Honey Lake Valley and Smoke Creek Desert, the distance from the edge of the model to the center of the floor of Smoke Creek Desert is more than 5 miles. The general-head altitude was set at 3,850 feet (the approximate altitude of the floor of Smoke Creek Desert) for layer 1, and 2 feet higher for each of the subsequent lower layers. Smoke Creek Desert is a large basin; its valley floor is greater than 500 mi^2 and it receives only a small part of its ground-water recharge from Honey Lake Valley. The estimated distance from the boundary of the Honey Lake Valley flow model to the assumed constant head on the floor of Smoke Creek Desert is great enough to assume the head would remain relatively unaffected by ground-water withdrawals in Honey Lake Valley. Conductance values were 700 ft^2/d (layer 1), 175 ft^2/d (layer 2), and 100 ft^2/d (layers 3 and 4) in the calibrated model. Conductances simulated in these faulted volcanic rocks decrease with depth.

At Astor Pass, the distance from the general-head boundary at the edge of the model to Pyramid Lake is about 5 miles. The general-head altitude was set at 3,792 feet (the approximate altitude of Pyramid Lake) for layer 1, and 2 feet higher for each of the deeper layers. Conductance values in the calibrated model at this general-head boundary were the same as those for the boundary with Smoke Creek Desert.

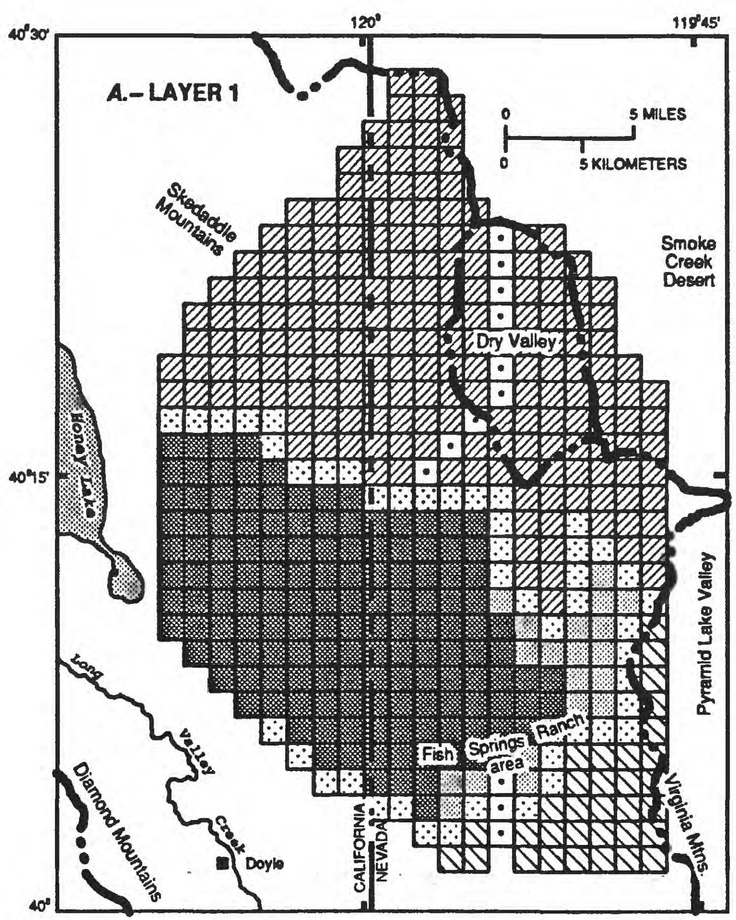
Aquifer Properties in the Flow-Model Area

Characteristics of aquifers that affect ground-water flow are estimated for each model cell. They include saturated thickness, average transmissivity or horizontal hydraulic conductivity, vertical hydraulic conductivity, and leakance between layers.

Aquifer Materials

The materials that make up an aquifer affect its ability to transmit water. The principal types of aquifers found within the flow-model area consist of unconsolidated basin-fill deposits in the low areas, and volcanic rocks that surround the low areas of the basin and, in some places, underlie or interfinger with the basin-fill deposits. The general distribution and horizontal hydraulic conductivity of aquifer materials in each layer of the flow model is shown in figure 21.

Layer 1 includes two types of basin-fill: fine-grained deposits (including the sedimentary and pyroclastic deposits of Pliocene age; table 1) composed primarily of clay, silt, and some sand, which are in the center of the basin; and coarser grained alluvial, deltaic, and near-shore pluvial deposits composed of gravel, sand, and silt, which form a belt at the edge of the valley floor. For purposes of the model, the basin-fill deposits in layers 2, 3, and 4, are assumed to consist entirely of fine-grained lakebed sediments.



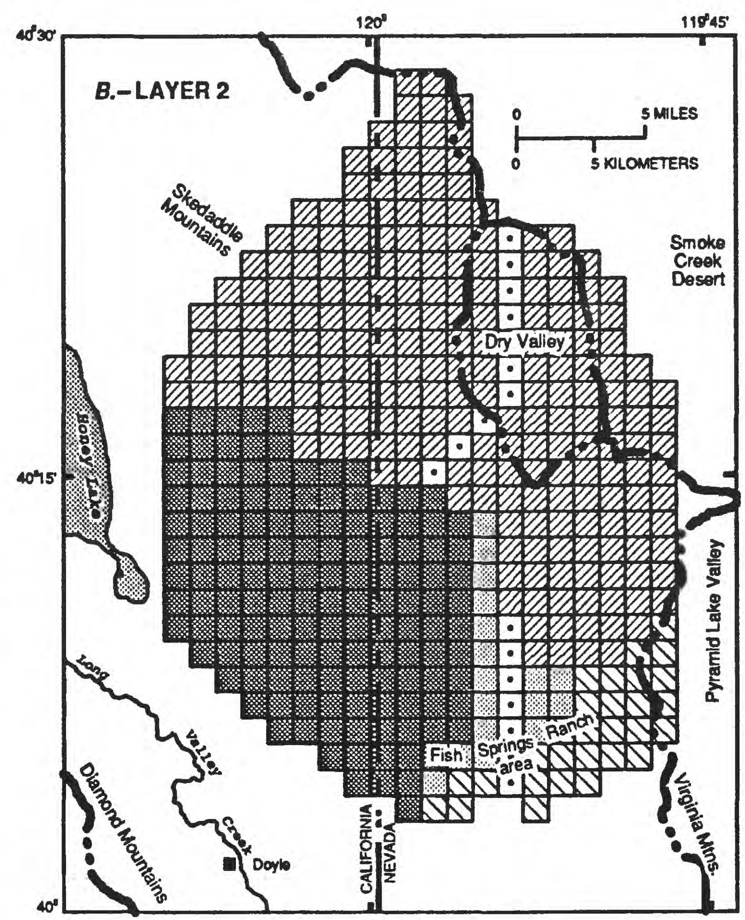
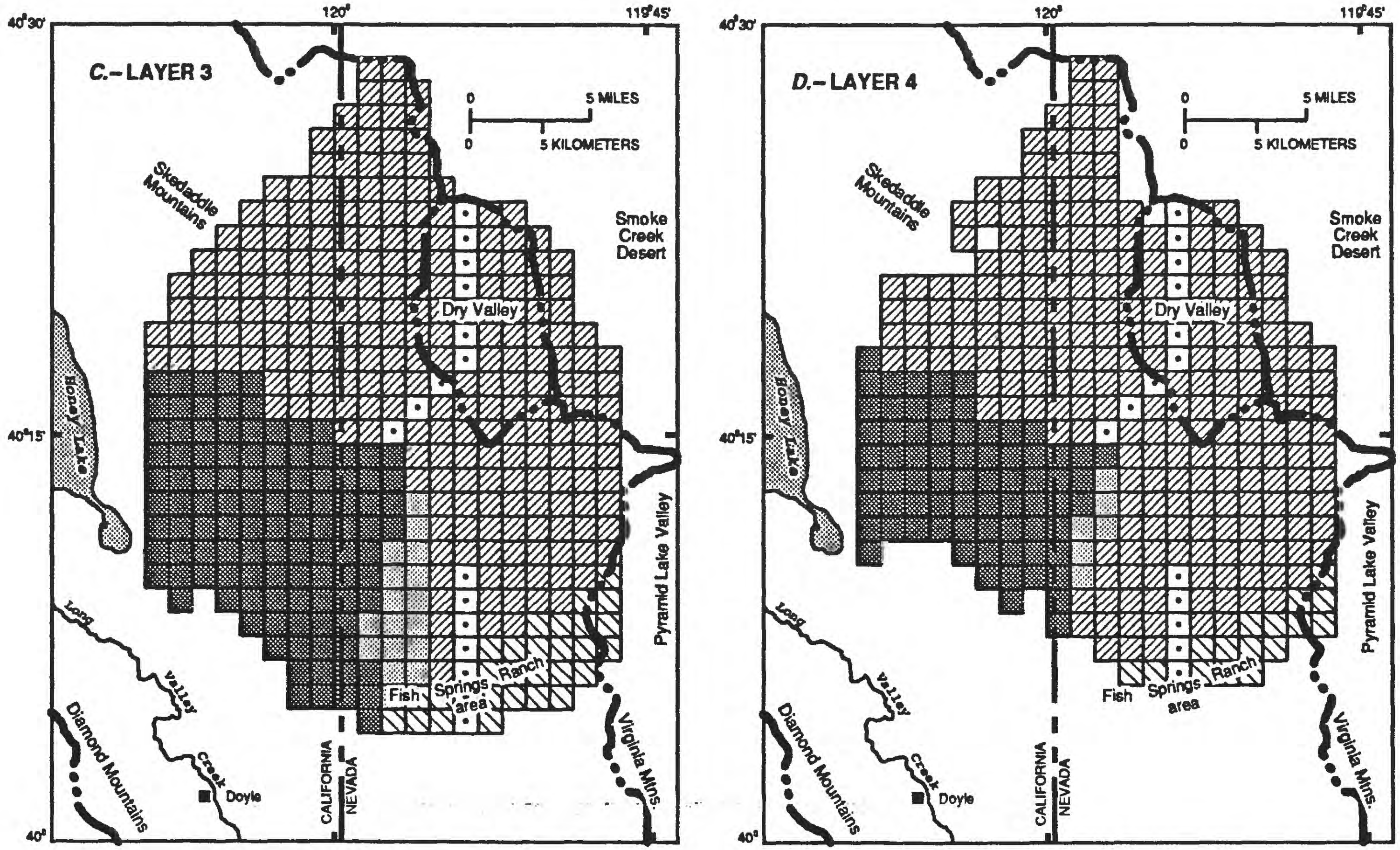


FIGURE 21.--Distribution of aquifer materials and horizontal hydraulic conductivities simulated in the ground-water model, eastern Honey Lake Valley and adjacent areas. (*A*) model layer 1; (*B*) layer 2; (*C*), layer 3; (*D*), layer 4.



EXPLANATION

AQUIFER MATERIALS AND HYDRAULIC CONDUCTIVITIES OF MODEL CELL

- SOUTHERN VOLCANIC ROCKS -- 45 feet per day in model layers 1 to 4
- NORTHERN VOLCANIC ROCKS -- 5 feet per day in model layers 1 to 4
- PERIMETER BASIN-FILL DEPOSITS -- 4 feet per day (model layer 1 only)
- CENTRAL BASIN-FILL DEPOSITS -- 1 foot per day in model layer 1; 0.75 in layer 2; 0.50 in layer 3; 0.25 in layer 4
- BASIN-FILL DEPOSITS AND VOLCANIC ROCKS -- 1 to 45 feet per day in model layer 1; 0.75 to 45 in layer 2; 0.50 to 45 in layer 3; 0.25 to 45 in layer 4
- FAULT ZONE -- 0.01 to 4 feet per day
- HYDROGRAPHIC BOUNDARY

FIGURE 21.--Continued.

Although several types of volcanic rocks are included in the flow-model area, they can be divided into two groups on the basis of age, location, and general hydraulic characteristics. The two groups are the older volcanic rocks of the Fort Sage and Virginia Mountains south of Astor Pass, including the relatively permeable Pyramid Sequence, and the younger Modoc Plateau basalts north of Astor Pass. For purposes of the model, hydraulic characteristics were assumed to differ more between the older (southern) and younger (northern) volcanic rocks than from place to place within a group. The general characteristics of each group of volcanic rocks was assumed to be uniform. The extent of the older and younger volcanic rocks is shown in figure 21. Numerous faults cut through the volcanic rocks in both the northern and southern parts of the study area. The faults are assumed to affect flow only in the volcanic rocks.

Saturated Thickness

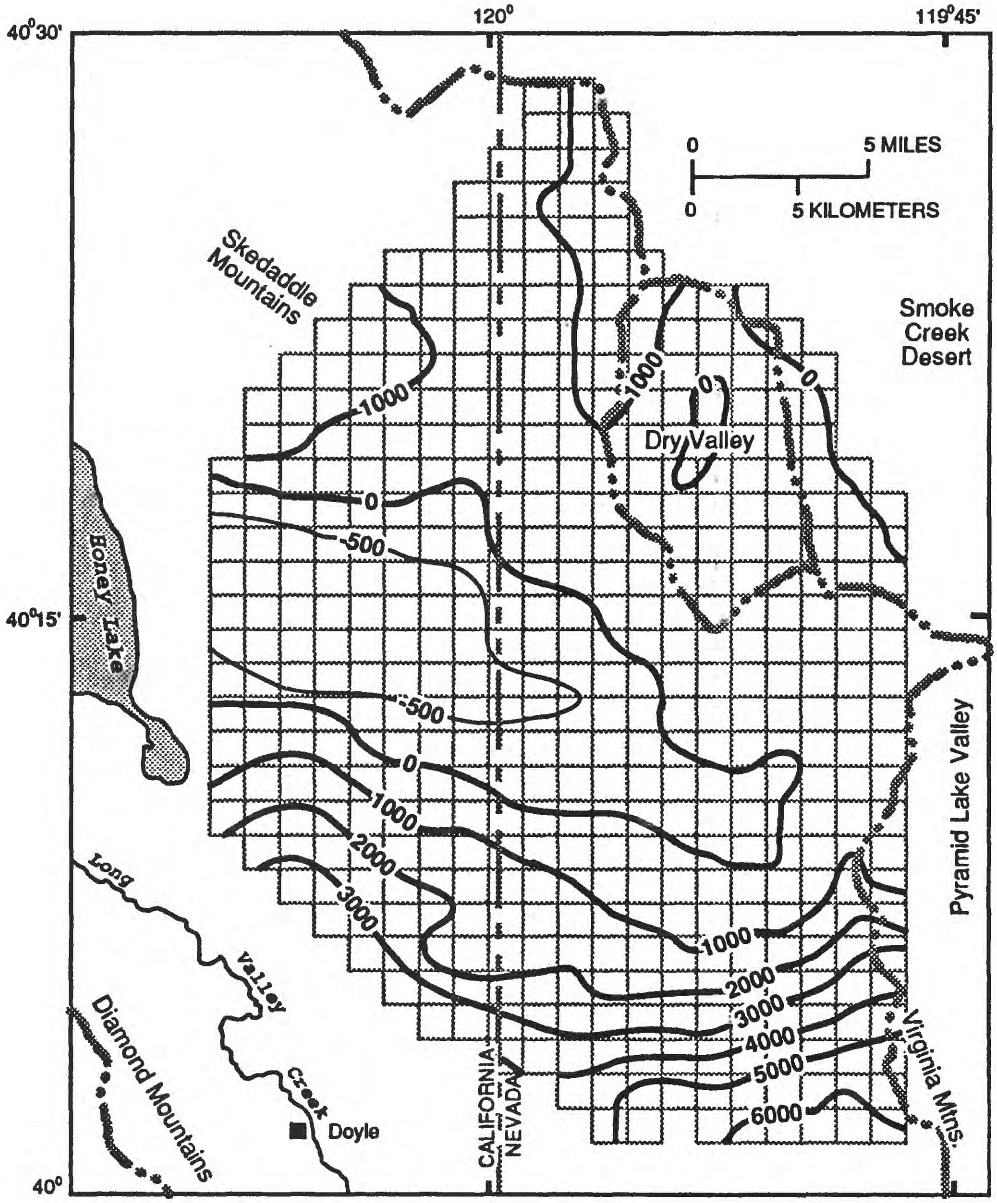
Transmissivity is computed in the model as the product of the saturated thickness of a particular model cell and the hydraulic conductivity assigned to that cell. The saturated thickness of layer 1 is calculated in the model by subtracting the altitude of the bottom of layer 1 from the altitude of the water table (figure 26A). The altitude of the bottom of layer 1 is 3,700 feet, except in the southern part of the model area where granitic bedrock is above 3,700 feet (figures 20 and 22). In this area, the bottom of layer 1 is the contact with impermeable granitic bedrock. The saturated thickness of each layer-2 and layer-3 block is calculated by subtracting the altitude of the bottom of the layer from the altitude of the bottom of the overlying layer, respectively. The bottom altitude of layer 2 is 3,000 feet and the bottom altitude of layer 3 is 1,500 feet, except where granitic bedrock is above these altitudes. In these areas, the bottom of the layer is the altitude of the contact with granitic bedrock. The saturated thickness of layer 4 is calculated by subtracting the altitude of the granitic bedrock surface from the altitude of the bottom of layer 3.

The altitude of the granitic-bedrock surface (figure 22) was estimated from geophysical surveys of the model area; the base of the flow model corresponds to this surface. During calibration of the model, the saturated thicknesses of model cells in layer 1 were recalculated with each change in altitude of the simulated water table; the saturated thickness of model cells in layers 2, 3, and 4 remained constant because the water level never declined below the tops of these layers.

Hydraulic Conductivity

Estimates of hydraulic conductivity for the basin-fill deposits, based on production tests and on specific capacities calculated from well drillers' data, range from less than 1 ft/d to greater than 10 ft/d. The higher values were generally from wells in the area of the near-shore deposits. Estimates of hydraulic conductivity for the volcanic materials range from about 2 ft/d to about 70 ft/d, on the basis of reported specific-capacity test data. Specific-capacity data from wells in the volcanic rocks were from the southern part of the basin near Fish Springs Ranch. For model cells in which basin-fill deposits overlie volcanic rocks, hydraulic conductivity was computed as a composite of the hydraulic conductivity of the basin-fill deposits and the hydraulic conductivity of the volcanic rocks, proportional to the extent and thickness of each unit within the particular cell.

For initial model simulations, uniform hydraulic conductivities were assigned to each material type in each layer. Initial values for hydraulic conductivity of basin-fill deposits for layer 1 were 1 ft/d for the fine-grained lakebed sediments and 4 ft/d for the coarser grained near-shore deposits. Hydraulic conductivity decreases with depth in deep sedimentary basins because of compaction by the overburden. For example, Durbin and others (1978, p. 76) report a 50 percent decrease in hydraulic conductivity per 1,200 feet of depth. For this study, it was assumed that the hydraulic conductivities of the basin-fill deposits in layers 2, 3, and 4 were 75, 50, and 25 percent of the hydraulic conductivities of layer 1 materials. Initially, the hydraulic conductivity of the volcanic units and the fault zones, shown in figure 21, was assumed to be 70 ft/d. The hydraulic conductivity of volcanic rocks initially was assumed to be the same for each layer.



EXPLANATION

MODEL CELL

—1000— BEDROCK CONTOUR -- Shows approximate altitude of bedrock surface. Contour interval 1,000 feet, with supplemental contour at -500 feet. Datum is sea level

HYDROGRAPHIC BOUNDARY

FIGURE 22.--Estimated altitude of bedrock surface in the ground-water model, eastern Honey Lake Valley and adjacent areas.

Initial estimates of hydraulic conductivity were changed uniformly for each material during the calibration process. Estimates for the older volcanic rocks in the south were reduced from 70 ft/d to 45 ft/d, and for the younger volcanic rocks in the north, from 70 ft/d to 5 ft/d. Initially, fault zones were assumed to have no effect on ground-water flow. However, during the calibration process, hydraulic conductivities for two fault zones within the volcanic rocks (figure 21) were reduced to 0.01 ft/d to achieve a better approximation of the measured water levels and gradients. Other faults in the modeled area may also affect flow; they were not simulated because their effects as conduits or barriers to flow are unknown. Near these faults, either water-level data are unavailable or water levels do not indicate effects. Therefore, reduced hydraulic conductivities were not required to represent these faults in the calibrated model. Initial estimates of hydraulic conductivity for the basin-fill deposits were also adjusted during the calibration process, but the original estimates produced the best match between measured and simulated heads. Figure 21 shows the distribution of hydraulic conductivity for each of the four layers at the end of the calibration process.

Leakage Between Layers

Vertical leakage of water between model layers occurs where the hydraulic head in a model block differs from the head in the block below. The rate of leakage is the effective value of vertical hydraulic conductivity between layers, multiplied by the difference in head between the two layers, divided by the length of the vertical flow path. The vertical hydraulic conductivity divided by the length of the flow path, referred to as the leakance term, is used in the model.

On a regional scale, it is not uncommon for the horizontal hydraulic conductivity to be 100 or more times greater than the vertical hydraulic conductivity (Freeze and Cherry, 1979, p. 34). During model calibration for this study, this ratio produced the most satisfactory results. On the basis of this relation, leakance values were varied proportionally in response to changes in the horizontal hydraulic conductivity of the layers during the calibration process. The resulting approximate values of leakance between layers 1 and 2 ranged from 0.00002 to 0.0001 (ft/d)/ft of saturated thickness. Leakance between layers 2 and 3 ranged from 0.000005 to 0.00005 (ft/d)/ft of saturated thickness and leakance between layers 3 and 4 ranged from 0.000002 to 0.00004 (ft/d)/ft of saturated thickness.

Simulation of Recharge and Discharge

All recharge to the flow-model area is assumed to occur as (1) infiltration of precipitation falling directly on the modeled area, (2) infiltration of surface runoff originating within the modeled area, (3) infiltration of irrigation water, and (4) ground-water inflow across the western and southwestern boundary. Discharge from the flow-model area consists of: (1) evapotranspiration, (2) ground-water outflow from the eastern boundary, and (3) ground-water withdrawals from wells.

Recharge

Infiltration of Precipitation

The Deep Percolation Model (DPM) was used to estimate ground-water recharge by the direct infiltration of precipitation in the flow-model area. The grid for the ground-water flow model overlaps the eastern part of the DPM grid so that deep percolation estimated for DPM cells can be directly used as recharge to corresponding flow-model cells. The eastern boundaries of the models are not exactly coinci- dent because the flow-model area was expanded during calibration. Recharge was extrapolated for noncoincident cells. The flow-model and DPM grids are shown in figure 8. Ground-water recharge from the direct infiltration of precipitation within the flow-model area, based on DPM results shown in figure 14, is estimated to average about 4,200 acre-ft/yr.

Figure 23A shows the areal distribution of ground-water recharge attributed to the direct infiltration of precipitation, which was used in the calibrated flow model. To calibrate the model (match simulated with observed water levels and areal distribution of evapotranspiration from phreatophyte zones), an additional 5,000 acre-ft of recharge was added in the southeast Virginia Mountains area, making the total amount in the model area 9,200 acre-ft/yr. Some possible explanations for the existence of additional recharge in the southeast part of the flow-model area are:

1. The volcanic rocks in this area are more permeable than those in the northern part of the model area and therefore precipitation infiltrates at a higher rate. The estimated hydraulic conductivity of the volcanic rocks in the southern area is 45 ft/d compared to 5 ft/d for the volcanic rocks in the northern part of the model area (figure 21), but the DPM (method used to estimate recharge from precipitation) does not consider the permeability of the underlying consolidated material and can underestimate recharge in such settings. The flow model was calibrated on the basis of this assumption.

2. The area contributing recharge to the flow-model area may be larger than the flow-model area if ground-water divides do not coincide with the topographic boundaries. Volcanic rocks in the southeast part of the study area extend beyond the topographic divides used for the DPM recharge estimates. Precipitation falling directly on volcanic rocks outside the model area may enter as subsurface inflow.

3. Water could enter the model area from basins to the south as underflow through fault zones. As discussed in the section "Subsurface Inflow," data collected for this study could not confirm or refute this possibility.

Infiltration of Streamflow

Observations made during the course of this study indicate that in the flow-model area, almost all the surface-water runoff from the mountainous areas infiltrated into the basin-fill deposits within a short distance of the mountain front. Also, vegetation is sparse along stream channels in basin-fill deposits, indicating evapotranspiration of streamflow is minimal. Therefore, for modeling purposes, all surface-water runoff from the mountains surrounding the valley was assumed to infiltrate the basin-fill deposits near the mouths of the streams (figure 23B) and evapotranspiration from streamflow was assumed to be negligible.

Rush and Glancy (1967, p. 18) estimated that about 4,000 acre-ft/yr of surface-water runoff from the mountainous areas recharges the basin-fill deposits in a study area about one-half the size of the flow-model area. Because they estimated a discharge of 11,000 acre-ft/yr from the same area, they considered their recharge estimated to be too small, either underestimated or a result of unaccounted subsurface inflow (Rush and Glancy, 1967, p. 42). Estimates made for the present study indicate that about 13,000 acre-ft/yr of surface-water runoff from the mountains (table 14) recharges the basin-fill deposits in the modeled area. These estimates of recharge by streamflow are based on the work of Rockwell (in press) as part of this study.

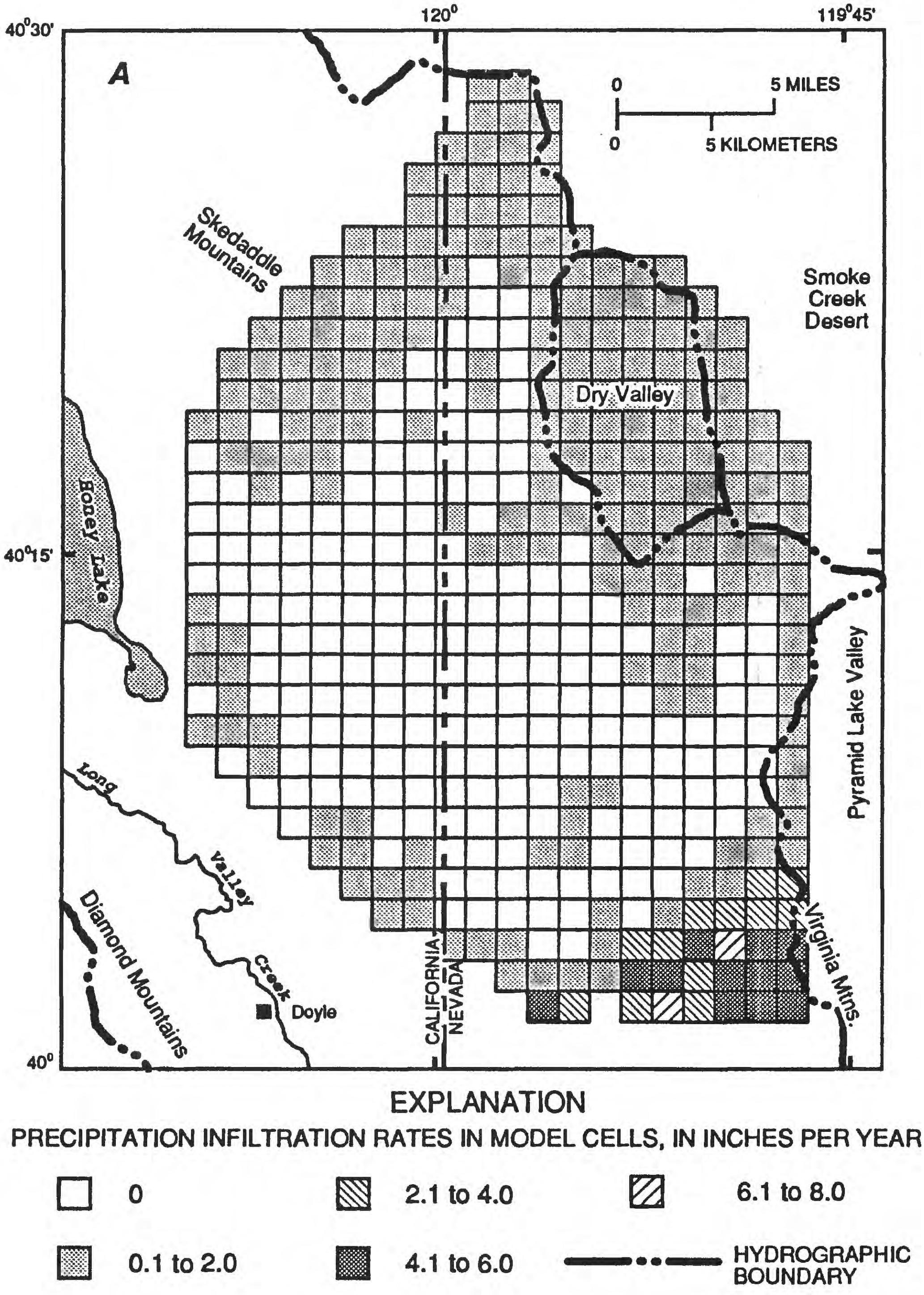
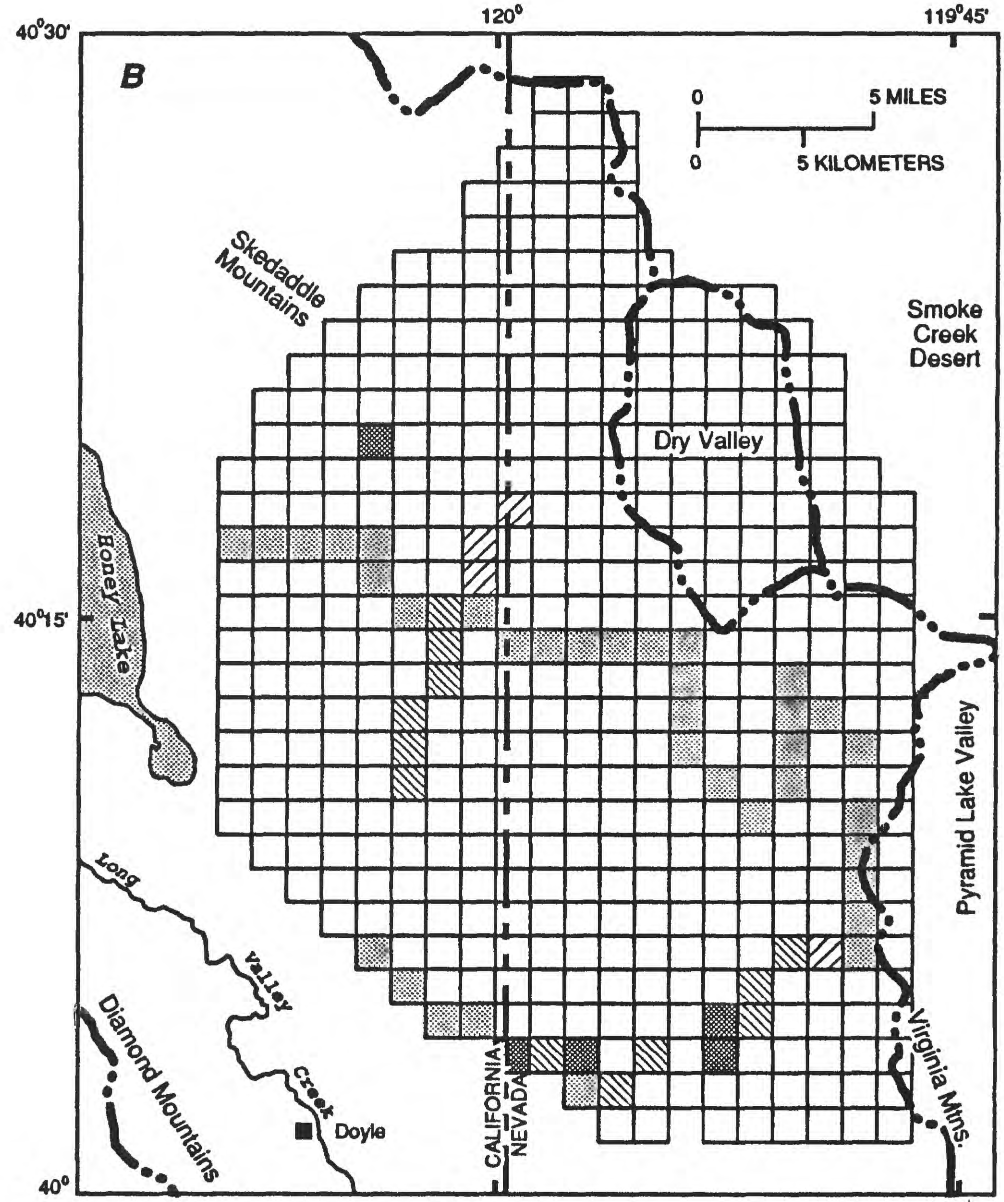


FIGURE 23.--Simulated distribution of recharge from infiltration of (*A*) precipitation and (*B*) streamflow in the ground-water model, eastern Honey Lake Valley and adjacent areas.



EXPLANATION

STREAMFLOW INFILTRATION RATES IN MODEL CELLS, IN ACRE-FEET PER YEAR

- 0
- 1 to 150
- 151 to 300
- 301 to 600
- 1,200 to 1,600
- HYDROGRAPHIC BOUNDARY

FIGURE 23.--Continued.

TABLE 14.--*Streamflow estimates for the ground-water flow model*

[Based on data from Rockwell (in press)]

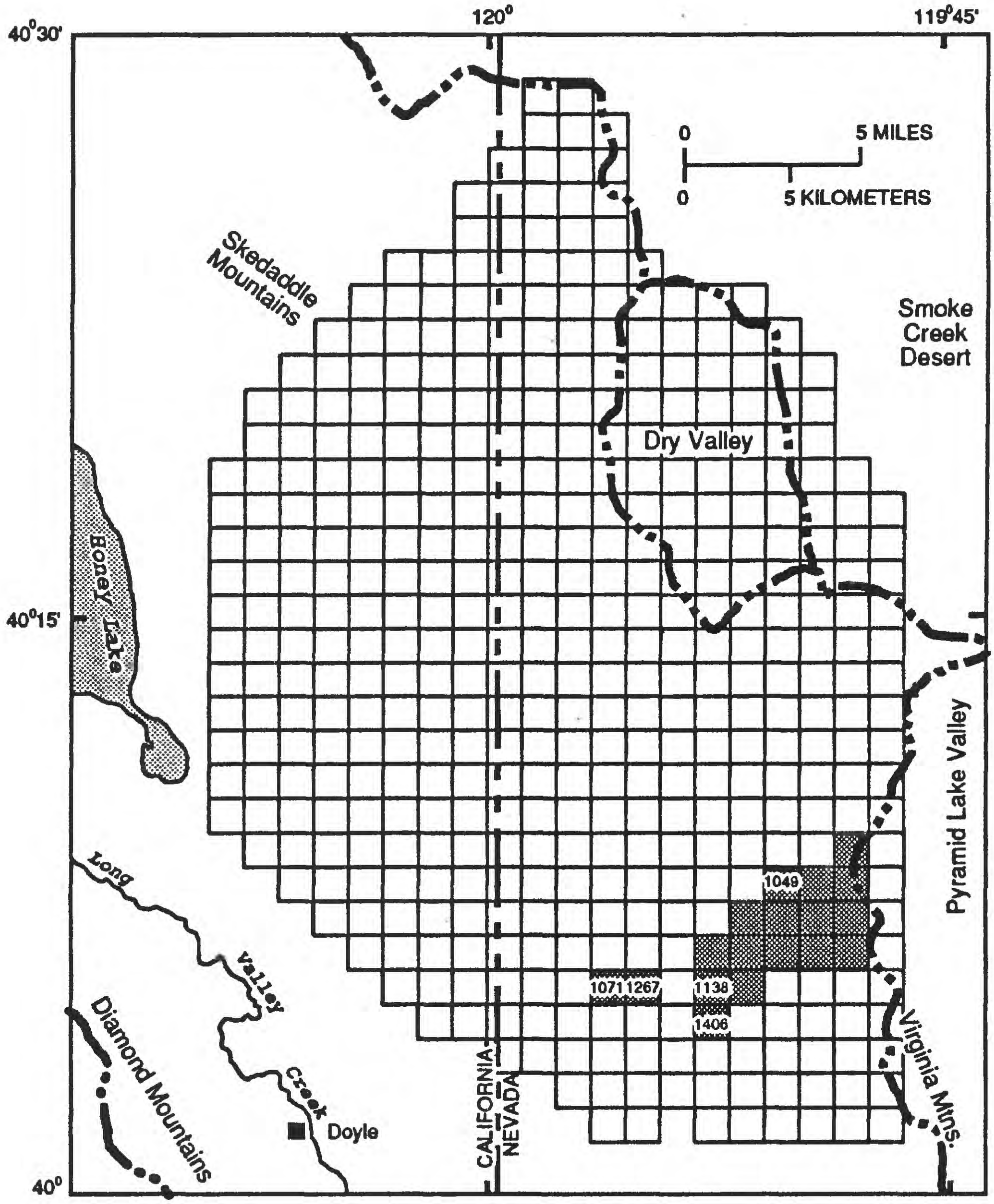
Stream or area	Drainage area (square miles)	Annual streamflow (acre-feet, rounded)
Spencer Creek	4.72	380
Skedaddle Creek	83.4	5,000
Intervening areas from Amedee Mountains to Never Sweat Hills	25	2,000
Cottonwood Creek	14.6	1,600
Gasperoni Creek	1.24	60
Rock Springs Creek	2.46	210
Milne Creek	2.28	200
Willow Springs Creek	.79	160
Fish Springs Creek	3.73	310
Antoinette Creek	.77	30
Butler Creek	.50	30
Mullen Creek	.46	110
Fort Sage Creek	1.56	440
Intervening areas in Virginia and Fort Sage Mountains	26	2,400
Total (rounded)	168	13,000

Infiltration of Irrigation Water

Of the ground-water withdrawn for irrigation and applied to fields, 25 percent is assumed to infiltrate to the water table as recharge. The total annual recharge from irrigation return flow is estimated to be about 1,500 acre-ft, on the basis of 1988 irrigation withdrawals, and was assumed to recharge the model cells in layer 1 at the location of each of the five pumping wells simulated during calibration as shown in figure 24.

Subsurface Inflow

Ground-water inflow from the west and southwest was simulated with general-head boundaries (figure 19). Initial estimates of conductance across these boundaries were based on limited information. During model calibration, the conductance was adjusted until the best match between simulated and measured heads in wells near the boundaries was obtained. These adjustments affected the amount of inflow at the boundaries.



EXPLANATION

1071 MODEL CELL IN WHICH PUMPING WAS SIMULATED DURING CALIBRATION AND FOR HYPOTHETICAL DEVELOPEMENT CONDITIONS -- Number is pumping rate, in acre-feet per year, for calibration simulation. For hypothetical developement conditions, pumping rate is 834 acre-feet per year

MODEL CELL IN WHICH PUMPING WAS SIMULATED FOR HYPOTHETICAL DEVELOPMENT CONDITIONS -- Pumping rate, 834 acre-feet per year

HYDROGRAPHIC BOUNDARY

FIGURE 24.--Simulated distribution of ground-water pumping for the model calibration and for hypothetical development conditions, eastern Honey Lake Valley and adjacent areas.

The calibrated model simulated a total of about 580 acre-ft/yr entering the system across the western and southwestern general-head boundaries in the upper two layers. This agrees with Rush and Glancy's (1967, p. 24) estimate that about 600 acre-ft/yr of underflow enters from the west between the mouth of Skedaddle Creek in the north and Turtle Mountain in the south (plate 1). The model also simulated about 590 acre-ft/yr flowing westward toward Honey Lake in the lower two layers. No water-level data are available for layers 3 and 4 to verify the simulated outflow. The pattern of inflow near the surface and outflow in deeper layers along the western boundary may be the result of errors in heads specified outside the model along the general-head boundary. However, this would not affect net flow at the boundary or the State line, or changes in net flow in different simulations. The calibrated model simulated a net flow of about 700 acre-ft/yr eastward across the State line (table 18).

Discharge

Evapotranspiration

Evapotranspiration from ground water is simulated in the model by a linear decrease with depth, decreasing from a maximum evapotranspiration rate at land surface until it reaches a specified depth below land surface at which evapotranspiration ceases (extinction depth). The simulated rates and distribution can then be compared with independent estimates.

The maximum evapotranspiration rate used in the model (based on open-water evaporation at land surface) was assumed to be 4 ft/yr, about 5 percent higher than the evaporation rate estimated for Honey Lake, to account for the drier climate in the eastern part of the basin. The rate was not corrected for evapotranspiration of precipitation and soil moisture above the water table. However, sensitivity analysis indicated that a maximum evapotranspiration rate 50 percent greater or smaller would have negligible effects on the results.

The model simulates discharge by evapotranspiration in layer-1 blocks where the water-table altitude is above the evapotranspiration extinction depth. The extinction depths of evapotranspiration were assigned to model blocks (figure 25) on the basis of the following: (1) In the playa areas, evaporation of ground water through the bare surface is assumed to cease at an average depth of 12 feet below land surface, on the basis of depth to water in a test hole augered in the playa (Rush and Glancy, 1967, p. 32, 57). The extinction depth of evapotranspiration is shallow because no phreatophyte roots are present and sediments are very fine grained. (2) In the areas where greasewood (a phreatophyte) predominates, the extinction depth is estimated to be 36 feet below land surface on the basis of observations in Honey Lake Valley and elsewhere. For example, Meinzer (1927, p. 41) reported depth to water was as much as 33 feet below land surface in greasewood areas in Big Smoky Valley, Nev. Greasewood roots were observed at a depth of 57 feet below land surface near Grand View, Idaho (Robinson, 1958, p. 65-66). In Honey Lake Valley, greasewood grows near observation wells 1 and 2 (plate 1), where the depth to water is about 20 feet, but this probably is not the maximum depth to water in areas where greasewood is dominant. (3) The other areas contain mixed phreatophytic vegetation composed mainly of greasewood, rabbitbrush, grasses, and forbs. These plants are assumed to transpire ground water from a maximum depth of 24 feet below land surface. In Honey Lake Valley, mixed phreatophytes grow near observation well 21 (plate 1), where the depth to water is about 13 feet, also probably not the maximum. Phreatophytes survive on soil moisture where the water table is too deep to reach; the model simulates no evapotranspiration in these areas. Distribution of phreatophytic vegetation is shown on plate 4.

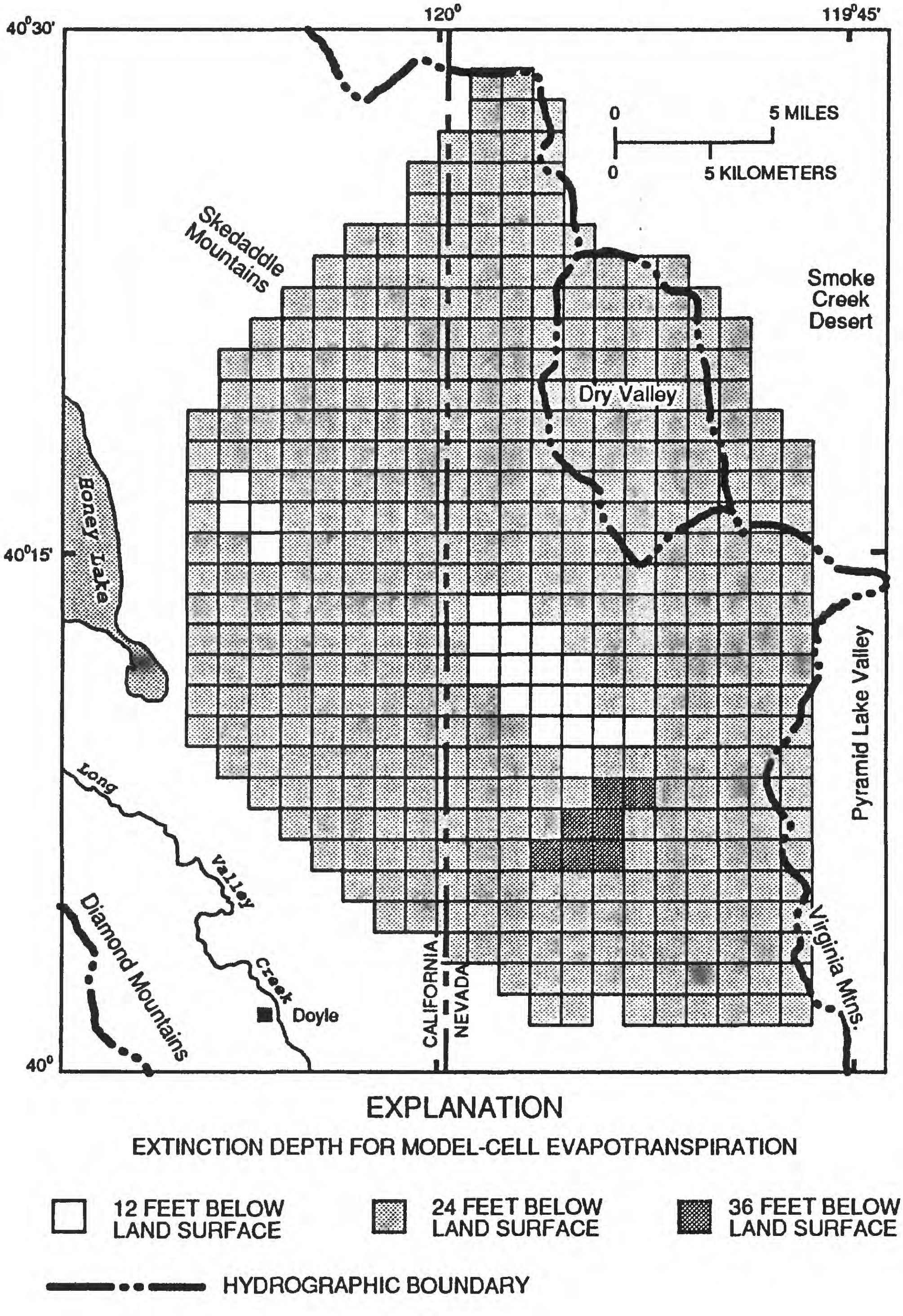


FIGURE 25.--Assumed extinction depths for evapotranspiration in the ground-water model, eastern Honey Lake Valley and adjacent areas.

For this study, for comparison with model simulations, the distribution of phreatophytic vegetation in the flow-model area was derived from a color-enhanced Landsat image (for September 2, 1980), and verified in the field. The Landsat results were divided into seven Level-I classes modified from those of Anderson and others (1976, table 2). The rangeland class was reclassified to identify potential phreatophyte zones, which were then field checked. About 26,000 acres in the flow-model area were classified as phreatophytic zones; about 31,000 acres as nearly barren or playa areas; and about 1,000 acres as native grass, including some wetlands (J. LaRue Smith, U.S. Geological Survey, written communication, 1988), for a total of about 58,000 acres where evapotranspiration from ground water is likely to occur. The distribution is shown on plate 4. Acreages in each class were multiplied by the corresponding evapotranspiration rates to compute an estimated 13,000 acre-ft of annual evapotranspiration (table 15).

At the end of the calibration process, the predevelopment flow model simulated annual evapotranspiration of approximately 15,000 acre-ft (about 15 percent greater than the 13,000 acre-ft based on Landsat imagery and estimated evapotranspiration rates), distributed over 88 cells of the model area, as shown in figure 29. The Landsat distribution was compared with the predevelopment simulation because the images are for 1980, before most of the irrigation pumpage began in the Fish Springs Ranch area. The simulated evapotranspiration rate is about 20 percent higher than that estimated by Rush and Glancy, 9,000 acre-ft for 50,000 acres (1967, table 14), an area about 10 percent smaller than that simulated by the model. The simulated area of evapotranspiration, about 56,000 acres, is about the same as the 58,000 acres estimated from Landsat images.

TABLE 15.--*Estimated evapotranspiration of ground water in the flow-model area*

Vegetation group	Area, rounded[1] (acres)	Annual evapotranspiration: Range[2] (feet)	Annual evapotranspiration: Rate used (feet)	Annual evapotranspiration: Acre-feet, rounded
Mixed phreatophytes	26,000	0.2-0.4	0.3	7,800
Bare playa and sparse phreatophytes	31,000	0.1-0.2	.1	3,100
Native grass[3]	1,000	1.5-2.0	1.8	1,800
Total	58,000	--	--	13,000

[1]Based on analysis of Landsat data for September 1980 and limited field checking by J. LaRue Smith (U.S. Geological Survey, written communication, 1988). Distribution of phreatophytic vegetation is shown on plate 4.

[2]Rates are based on data from work in Honey Lake Valley by the California Department of Fish and Game (written communication, 1988), William D. Nichols (U.S. Geological Survey, written communication, 1988), and work in other areas by Lee (1912), White (1932), Young and Blaney (1942), Robinson (1970), Houston (1950), and Patrick A. Glancy and James R. Harrill (U.S. Geological Survey, written communication, 1990).

[3]Native grass areas include some wetlands. The extent of irrigated cropland in the flow-model area was negligible in 1980.

Subsurface Outflow

Ground-water outflow eastward to Smoke Creek Desert and Pyramid Lake Valley was simulated in the model by general-head boundaries. During the calibration process, the conductances of the general-head boundaries were varied until the simulated heads for the model blocks near these boundaries matched measured heads in wells located in the blocks. This resulted in simulated outflow to Smoke Creek Desert of about 5,300 acre-ft/yr and through Astor Pass to Pyramid Lake Valley of about 1,500 acre-ft/yr.

Ground-Water Withdrawals

Ground-water withdrawals by wells from the flow-model area were simulated as discharge from five model blocks that correspond to locations of five existing irrigation wells in the Fish Springs Ranch area (figure 24). (Discharge from small-capacity wells was ignored.) Estimates of total annual withdrawals from the irrigation wells during 1988 range from about 3,500 acre-ft, on the basis of flow-meter readings (adjusted for periods when the meters were malfunctioning), to 6,500 acre-ft on the basis of irrigated acreage and estimated crop usage. An intermediate value of about 5,900 acre-ft was used for the model simulations (table 16).

TABLE 16.--*Simulated ground-water withdrawals from irrigation wells in the Fish Springs Ranch area, 1988*

Model cell (figure 19)			Well number (plate 1 and table 10)	Withdrawals (acre-feet per year)
Layer	Row	Column		
2	25	21	24	1,049
2	28	16	19	1,071
2	28	17	18	1,267
2	28	19	20	1,138
2	29	19	17	1,406
Total (rounded)				5,900

RESULTS OF FLOW-MODEL SIMULATIONS

A steady-state model that simulates the ground-water flow system under three sets of equilibrium conditions was completed for this study. First, the model was calibrated to current (1988) conditions of development. This provided a steady-state representation of the system using the 1988 distribution of pumped wells. Second, the steady-state predevelopment conditions were simulated by removing the pumping stress from the 1988 calibration. This produced a means for evaluating the validity of the assumption that the ground-water system was approximately in equilibrium during 1988. Third, the projected steady-state conditions resulting from hypothetical increased development were simulated. Each simulation resulted in a set of water levels and recharge and discharge rates for each model cell, and a ground-water budget for the entire modeled area.

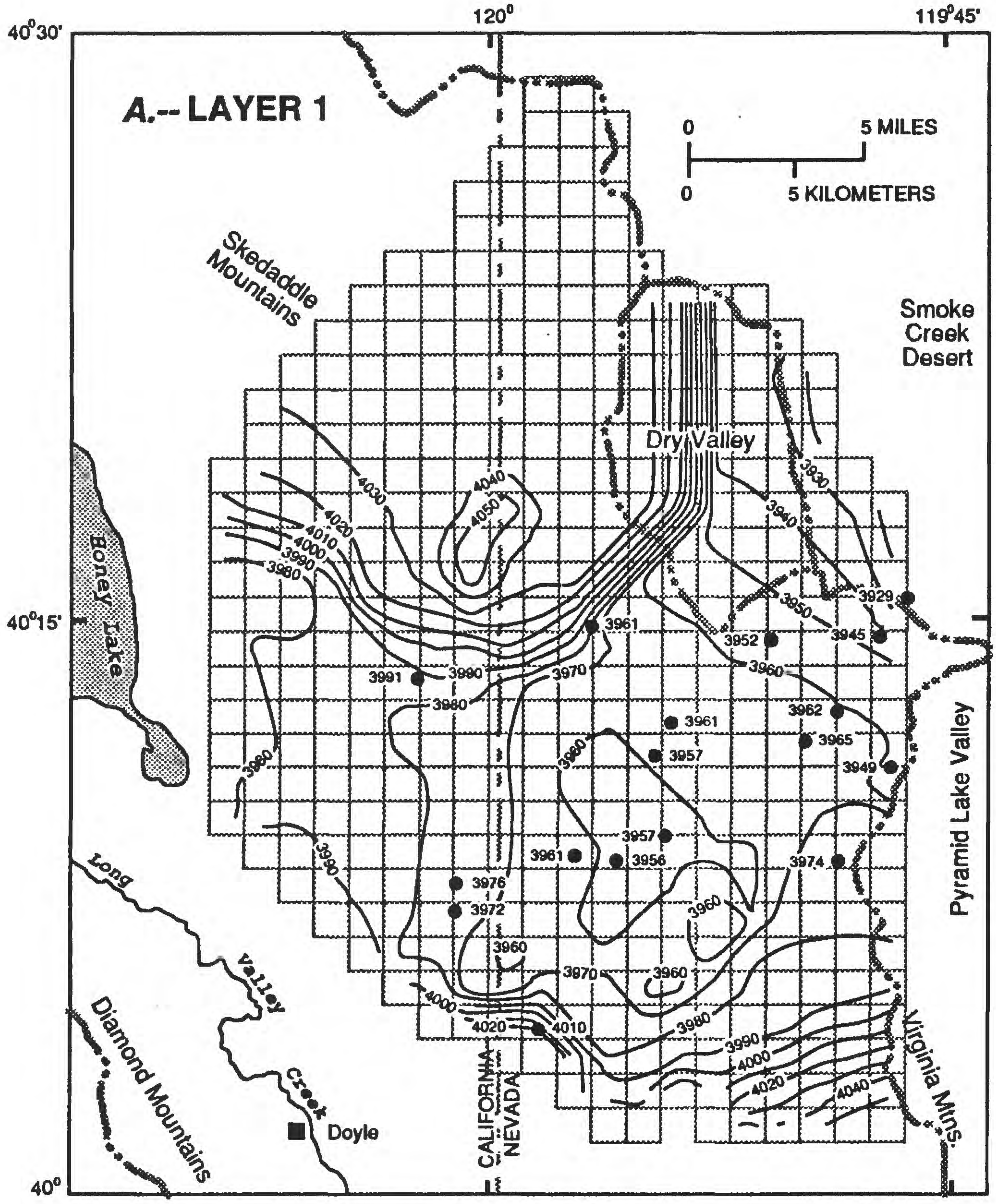
The Calibrated Model

The model was calibrated to 1988 conditions rather than to predevelopment conditions because hydrologic data for the period before development are limited. Mean annual rates of natural recharge and discharge and 1988 ground-water withdrawals and irrigation return flows were used in the model, and ground-water levels measured in 1988 were used to calibrate the model. The ground-water system was assumed to be in equilibrium.

Results of the model representing the ground-water flow system at current (1988) conditions of development are shown in figure 26A-D, as simulated water levels for each model layer at the end of the calibration process. Figure 26A and 26B also show the locations of, and measured water levels for 30 selected wells used for model calibration. The root-mean-square error between the measured water levels at the wells and the simulated water levels at corresponding model cells is about 2.7 feet, and the greatest difference between measured and simulated water levels is about 6 feet (table 17). In the north part of the valley, heads simulated in the lower layers are above land surface. This is consistent with observed flowing wells. In the western part of the model, simulated water levels in layer 1 are lowered by evapotranspiration, but the general direction of flow in the model is from west to east.

Both inflow and outflow are simulated across the west-southwest boundaries of the model area. Inflow primarily occurs in the upper layers of the model; outflow primarily occurs from the lower layers. However, water-level data to confirm the gradient are not available for depths corresponding to the lower layers of the model. In the calibrated model, inflows are within 2 percent of outflows across the western and southwestern boundaries. At the State line, simulated net flow is eastward at about 700 acre-ft/yr (table 18).

To demonstrate the effects of the simulated fault zone, a separate simulation was made assuming no reduction in hydraulic conductivity in model blocks representing the faults. Differences resulting from this simulation were negligible except for water levels in the vicinity of Fish Springs Ranch. Near the fault, simulated water levels were lower east of the fault and were higher west of the fault, resulting in a water-level gradient from west to east, which is opposite to the gradient indicated by measured water levels in wells in that area. The maximum difference in head at the State line was an increase of about 1 foot. Annual ground-water flow across the western boundary did not change. Outflow across the eastern boundary decreased by about 100 acre-ft (less than 1 percent) and discharge by evapotranspiration also increased by less than 1 percent, in comparison to the calibrated model.



EXPLANATION

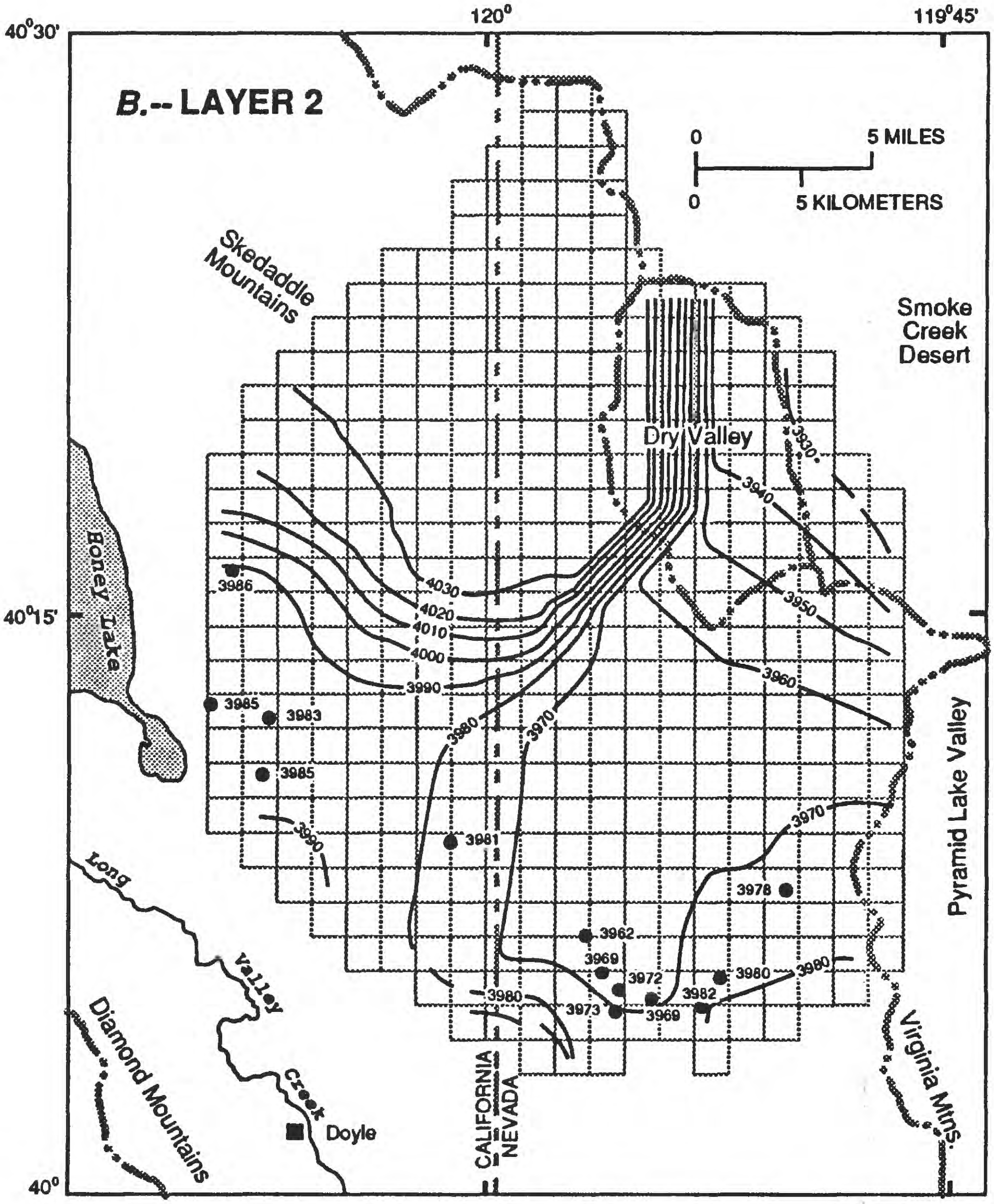
MODEL CELL

HYDROGRAPHIC BOUNDARY

4000 SIMULATED WATER-LEVEL CONTOUR -- Shows water-surface altitude, in feet above sea level. Contour interval, 10 feet

3949 ● WELL AND MEASURED WATER LEVEL, 1988 -- Water-surface altitude, in feet above sea level

FIGURE 26.--Measured ground-water levels, 1988, and simulated water-level contours at the end of model calibration, eastern Honey Lake Valley and adjacent areas. (*A*), layer 1; (*B*), layer 2; (*C*), layer 3; (*D*), layer 4.



EXPLANATION

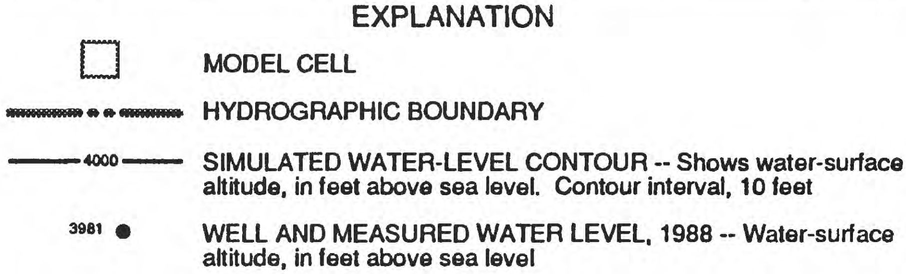
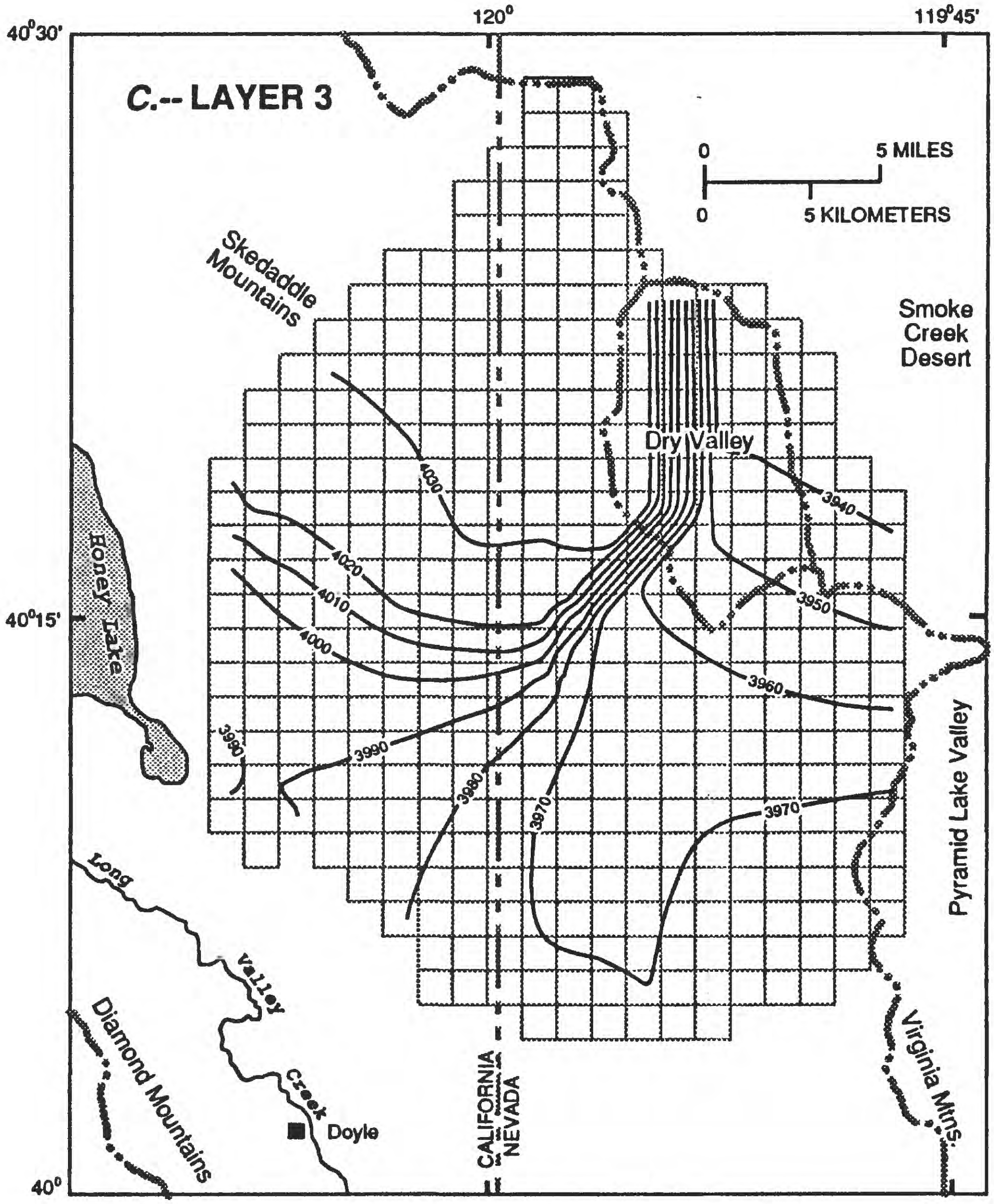


FIGURE 26.--Continued.



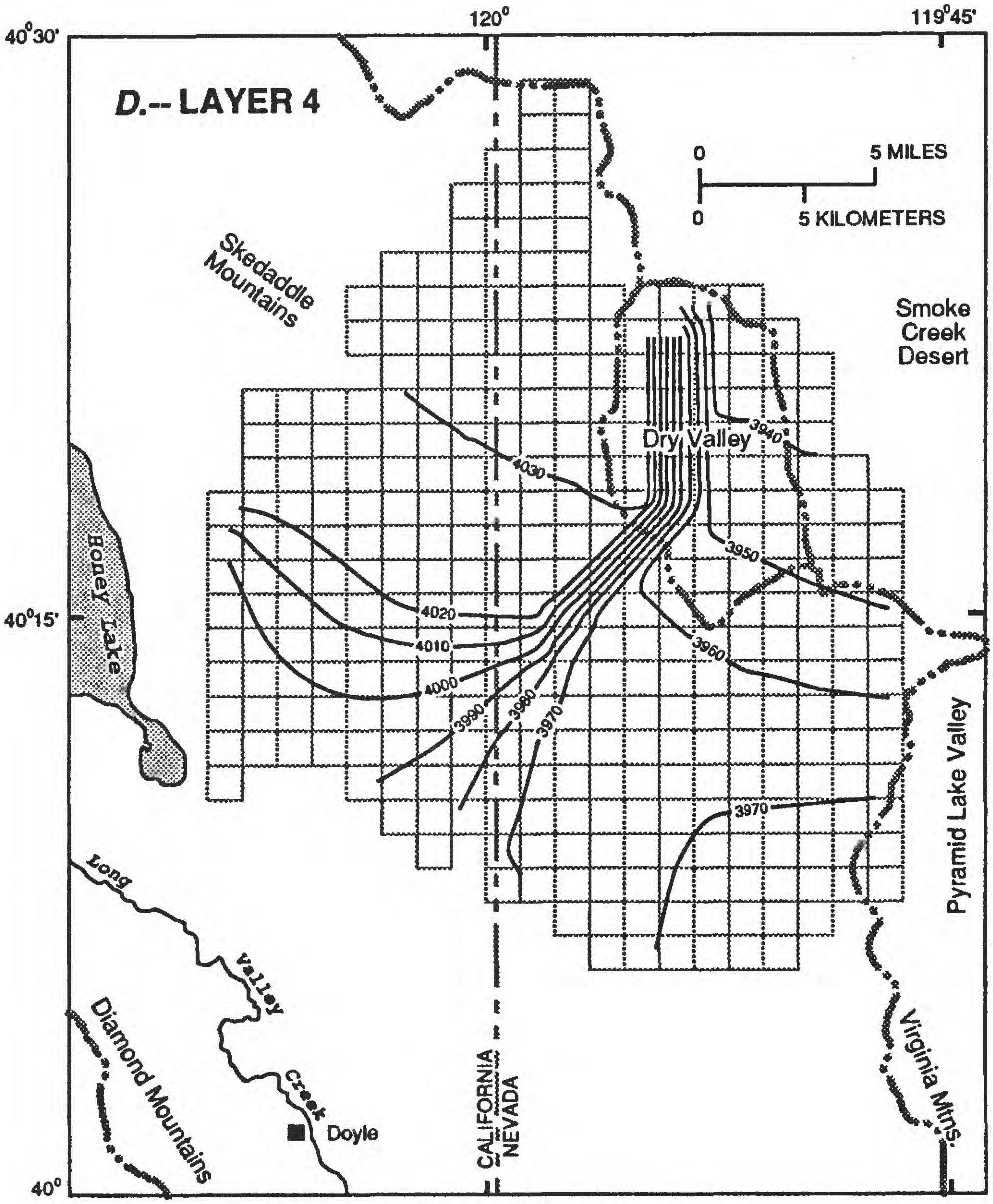
EXPLANATION

MODEL CELL

HYDROGRAPHIC BOUNDARY

SIMULATED WATER-LEVEL CONTOUR -- Shows water-surface altitude, in feet above sea level. Contour interval, 10 feet

FIGURE 26.--Continued.



EXPLANATION

MODEL CELL

HYDROGRAPHIC BOUNDARY

SIMULATED WATER-LEVEL CONTOUR -- Shows water-surface altitude, in feet above sea level. Contour interval 10 feet

FIGURE 26.--Continued.

TABLE 17.--*Measured and simulated water levels at selected observation wells, 1988*

Model cell (figure 19)			Well number (plate 1 and table 10)	Land-surface altitude (feet above sea level)	Water-level altitude (feet above sea level)		
Layer	Row	Column			Measured	Simulated	Difference[1]
1	17	16	38	4,110	3,961	3,965	+4
1	17	24	39	4,178	3,929	3,931	+2
1	18	21	36	4,254	3,952	3,956	+4
1	18	24	37	4,045	3,945	3,948	+3
1	19	10	8	4,010	3,991	3,986	-5
1	20	18	33	4,063	3,961	3,966	+5
1	20	22	34	4,005	3,962	3,963	+1
1	21	17	30	3,995	3,957	3,963	+6
1	21	21	32	4,001	3,965	3,968	+3
1	22	24	31	4,010	3,949	3,951	+2
1	23	17	28	3,991	3,957	3,956	-1
1	24	15	27	3,988	3,961	3,963	+2
1	24	16	26	3,986	3,956	3,959	+3
1	24	18	29	3,991	3,957	3,957	0
1	24	22	25	4,172	3,974	3,974	0
1	25	12	3	4,005	3,976	3,974	-2
1	26	12	2	4,000	3,972	3,972	0
1	29	14	14	4,145	4,010	4,008	-2
2	16	5	9	4,004	3,986	3,988	+2
2	20	5	7	4,000	3,985	3,985	0
2	20	6	6	4,009	3,983	3,984	+1
2	22	6	5	4,018	3,985	3,986	+1
2	24	11	4	4,010	3,981	3,979	-2
2	25	21	24	4,012	3,978	3,974	-4
2	26	15	23	3,979	3,962	3,966	+4
2	27	16	22	3,988	3,969	3,967	-2
2	28	16	19	3,995	3,972	3,968	-4
2	28	17	18	3,979	3,969	3,968	-1
2	28	19	20	4,012	3,980	3,979	-1
2	29	16	16	4,176	3,973	3,973	0
2	29	19	17	4,025	3,982	3,982	0

[1] Plus sign indicates simulated level above measured level; minus sign indicates simulated level below measured level.

TABLE 18.--*Simulated quantity and direction of net ground-water flow across the California-Nevada State line*

Model layer and altitude (feet above sea level)	Net interstate flow: Model calibration (1988 withdrawals)		Predevelopment (no withdrawals)		Proposed development (hypothetical withdrawals)	
	Acre-feet per year	Direction	Acre-feet per year	Direction	Acre-feet per year	Direction
Layer 1 (water table[1] to 3,700 feet)	350	eastward	320	eastward	600	eastward
Layer 2 (3,700 to 3,000 feet)	270	eastward	210	eastward	700	eastward
Layer 3 (3,000 to 1,500 feet)	60	eastward	40	westward	710	eastward
Layer 4 (1,500 feet to bedrock)	10	westward	40	westward	330	eastward
Net combined flow for layers 1-4 (rounded)	700	eastward	400	eastward	2,300	eastward

[1] Water table for 1988 conditions.

To ascertain the effects of the assumed altitude of the Honey Lake surface on flow across the general-head boundary at the western edge of the model, another simulation was made. The general-head (lake-surface) altitude was set at 3,983 feet (3 feet below the level in the calibrated model). As a result of this change, simulated ground-water inflow across the western boundary decreased about 140 acre-ft/yr (about 25 percent) and the maximum change in water levels was a 3-foot decline at the western edge of the model.

Model Sensitivity

Many assumptions and estimates are used in the design and construction of a ground-water flow model. To test the response of the model to a range of values for the initial estimates, a sensitivity analysis is made. This analysis indicates what changes in water levels and in components of the water budget would result from the use of different estimates of aquifer properties within assumed limits. The limits are set to encompass the range of uncertainty in the estimated values. The procedure involves uniformly changing values of hydraulic properties in the calibrated flow model. For each set of sensitivity tests, one property is varied as the others are held constant, and the magnitude and direction of resultant changes in water levels and in components of the water budget are recorded.

To evaluate the results, the root-mean-square deviation (error) between measured and simulated heads in the modeled area (1988 conditions), and the differences in simulated flow across the western and eastern boundaries were determined. Differences provide a qualitative assessment of sensitivity, but they cannot be used to verify the accuracy of a steady-state model, because solutions to steady-state models are not unique. Errors in one set of estimated values can compensate for errors in another set and produce the same results.

The root-mean-square deviations (errors) in water levels and in net ground-water flow across the western and eastern boundaries were plotted against the change factor for each aquifer property (figure 27A-C) and each recharge and discharge estimate (figure 27D-F). A change factor of 1.0, indicated by a vertical line at the center of each plot, represents the calibrated model and can be used for comparison. The greater the deviation of the flow rate or water level from its initial value at a change factor of 1.0, the greater the sensitivity of the model to an increase (change factor greater than 1) or decrease (change factor less than 1) in that aquifer property or initial estimate. The change factor was applied to the values for each property in all four layers simultaneously. It should be noted that an increase in hydraulic conductivity has the same result as an equal increase in thickness of a layer, because the model calculates transmissivity as hydraulic conductivity multiplied by thickness. Change factors for each property were varied from 0.1 to 10 times the calibrated values, where possible. However, in some cases the simulation could not be completed (equations of flow could not be solved) using extreme values. For example, when the conductance across the eastern (Smoke Creek Desert and Pyramid Lake) boundary was decreased to one-half the calibrated value (see plus (+) symbol in figure 27A), the simulation could not be completed.

The results of sensitivity analysis on simulated water levels are shown in figure 27A and 27D. Figure 27A indicates that simulated water levels are most sensitive to the conductance of the interface at the general-head boundaries along the eastern edge of the model (Smoke Creek Desert and Pyramid Lake Valley). Figure 27D indicates that simulated water levels also are sensitive to withdrawals from the irrigation wells, the evapotranspiration extinction depth, recharge from precipitation, and recharge from the infiltration of streamflow originating within the valley. Simulated water levels are least sensitive to the hydraulic conductivity (or thickness) of the basin-fill (near-shore and lake-bottom) deposits and southern volcanic rocks, leakage between model layers, maximum evapotranspiration rate, and recharge from irrigation return.

EXPLANATION, PARTS A-C

HYDRAULIC CONDUCTIVITY

- ■ ——— PERIMETER BASIN-FILL DEPOSITS
- □ — — — CENTRAL BASIN-FILL DEPOSITS
- ● ——---— SOUTHERN VOLCANIC ROCKS
- ○ ——- -— NORTHERN VOLCANIC ROCKS
- Δ ··············· FAULT ZONE
- × ———— LEAKAGE BETWEEN MODEL LAYERS
- ∇ ——·—·· CONDUCTANCE ACROSS HONEY LAKE AND LONG VALLEY GENERAL-HEAD BOUNDARIES
- \+ —·—·— CONDUCTANCE ACROSS SMOKE CREEK DESERT AND PYRAMID LAKE VALLEY GENERAL-HEAD BOUNDARIES

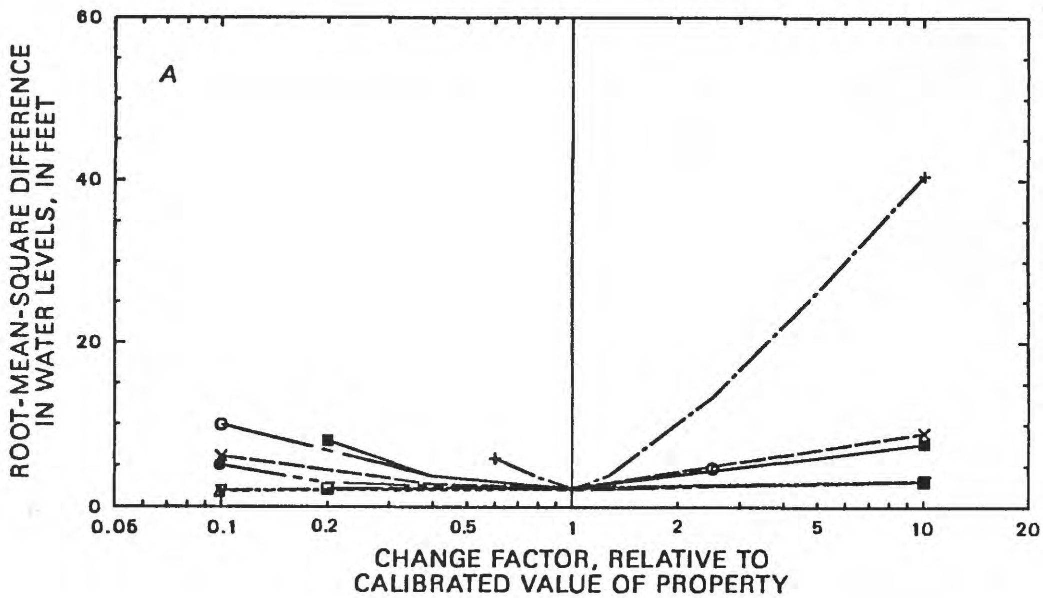


FIGURE 27.--Results of sensitivity analysis on simulated ground-water levels and simulated ground-water flow across western and eastern boundaries of the model, eastern Honey Lake Valley and adjacent areas. In graphs *B*, *C*, *E*, and *F*, negative values indicate flow out of model area, and positive values indicate flow into model area.

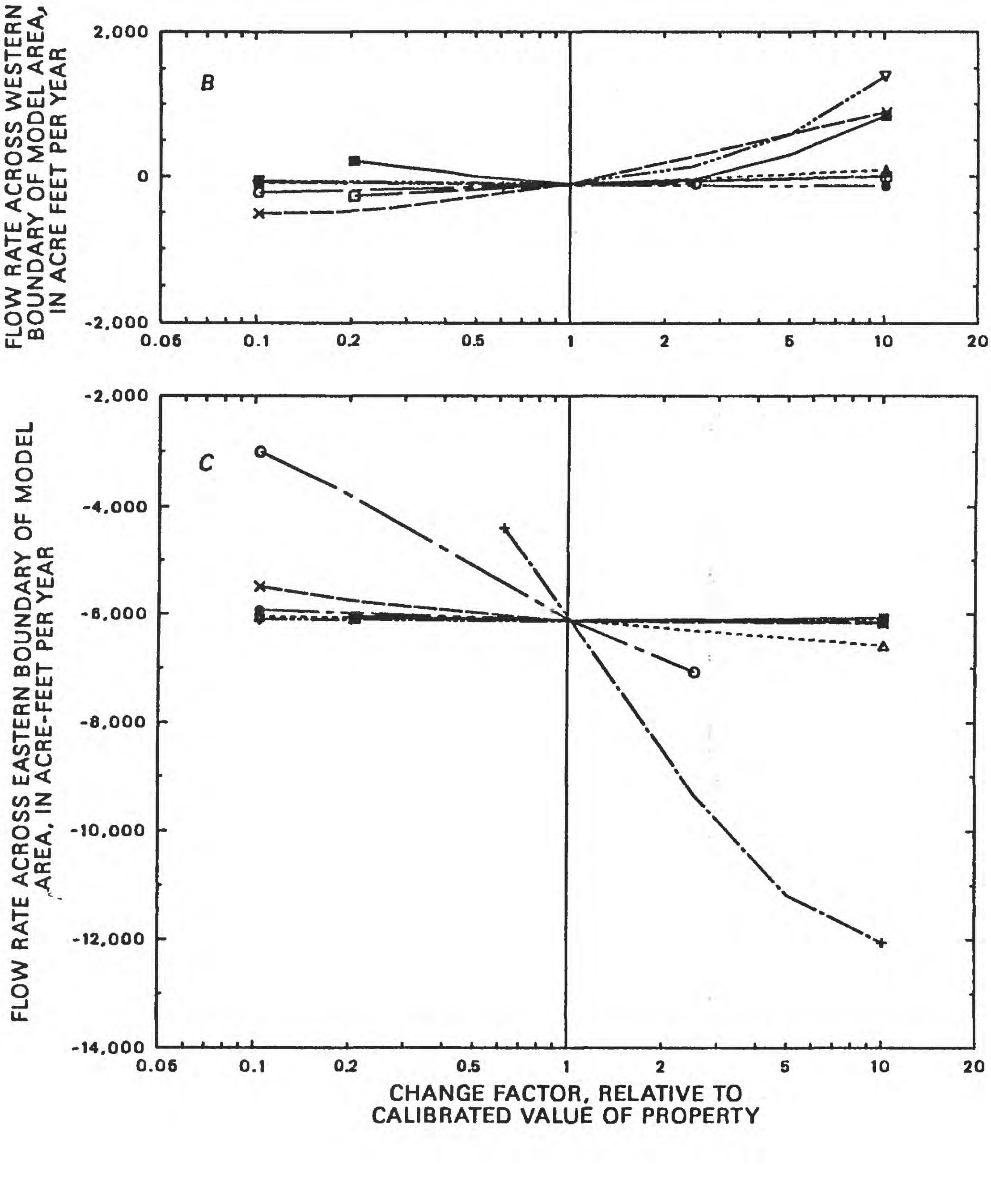


FIGURE 27.--Continued.

EXPLANATION, PARTS D–F

■ ———— MAXIMUM EVAPOTRANSPIRATION RATE

□ — — — EVAPOTRANSPIRATION EXTINCTION DEPTH

● ——---— IRRIGATION WITHDRAWALS

○ ———-— RECHARGE FROM DIRECT INFILTRATION OF PRECIPITATION

△ --------------- RECHARGE FROM INFILTRATION OF STREAMFLOW

× ——————— RECHARGE FROM INFILTRATION OF IRRIGATION WATER

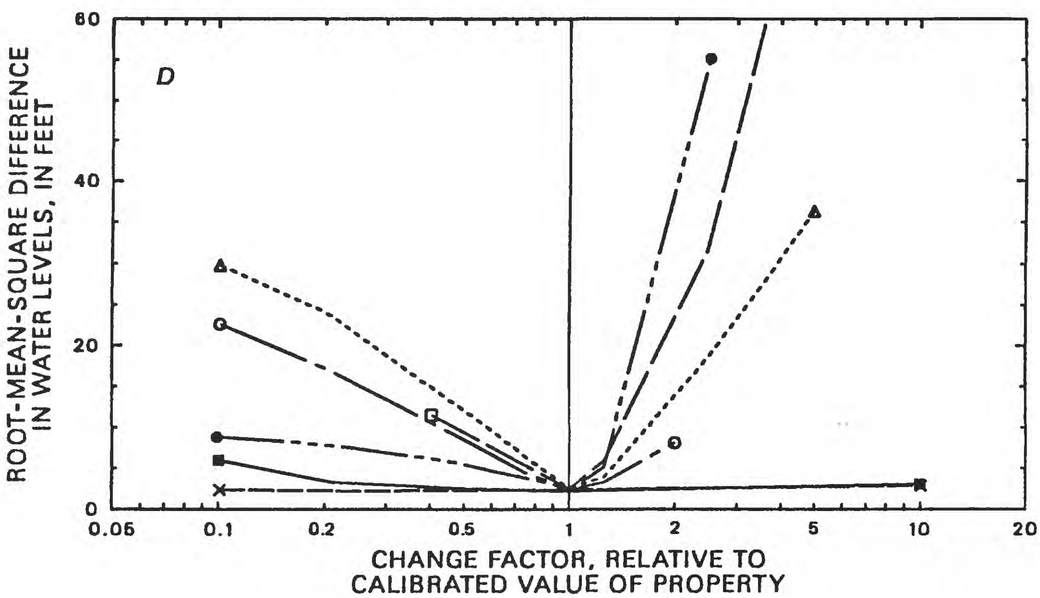


FIGURE 27.--Continued.

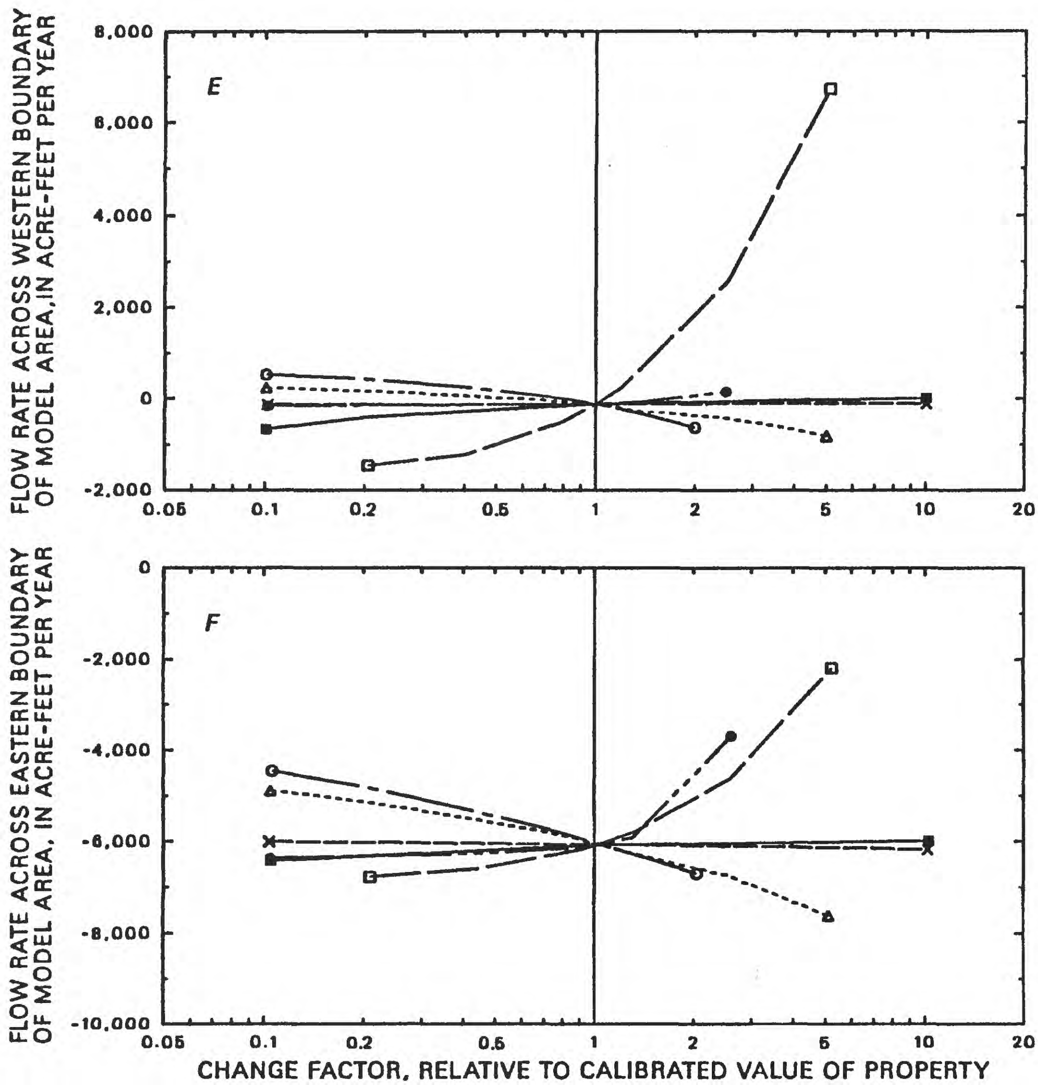


FIGURE 27.--Continued.

Flow across the western general-head boundary (figure 27B and 27E) is most sensitive to the evapotranspiration extinction depth. It is less sensitive to recharge from precipitation, recharge from the infiltration of streamflow, the conductance at the general-head boundary along the western edge of the model area, the hydraulic conductivity (or thickness) of the aquifer materials, and the leakance between model layers.

Flow across the eastern general-head boundary (figure 27C and 27F) is most sensitive to the conductance at that boundary, the hydraulic conductivity or thickness of the northern basalts, irrigation withdrawals, and the evapotranspiration extinction depth. It is least sensitive to the hydraulic conductivity or thickness of the other aquifers, the maximum evapotranspiration rate, and recharge from irrigation return.

Small errors in estimating values for properties to which the model is most sensitive can have a significant effect on the results. Other properties can be varied more than two orders of magnitude with very little effect on model results.

Limitations of the Flow Model

A digital model is useful for testing and refining a conceptual model of a ground-water flow system, developing an understanding of the system, guiding data collection, and projecting aquifer responses to changes in aquifer stresses within specified limits. However, a model is only an approximation of the actual system; it is based on simplifying assumptions and on average and estimated conditions, and cannot duplicate detailed field conditions. The accuracy with which a model can project aquifer responses is directly related to the accuracy of the input data used in the model calibration and it is inversely related to the magnitude of the proposed changes in stress.

The major limitation to using this model for projecting aquifer responses is that it is calibrated only to conditions that are assumed to be steady state. To calibrate a transient model requires a record of changes with time. Long-term historical data in the model area were insufficient to establish a match period. The steady-state model simulates an equilibrium that ultimately would be reached in response to a change in stress. It cannot predict the period required for the system to reach a new equilibrium. It cannot project intermediate head changes and it does not account for changes in storage.

The accuracy of the simulations is limited because the model uses average values of aquifer properties and average water levels in each cell, and for several cells where measurements are sparse. The model simulates general area-wide responses to stress, but should not be used where detailed site-specific projections are needed. Another factor that limits interpretation of projections on the basis of model simulations is the uncertainty of the effects of numerous faults in the area on the flow of ground water. These effects only can be determined by water-level monitoring as development proceeds to the extent that water levels are affected in the vicinity of the faults.

Predevelopment Conditions

After the model was calibrated to 1988 water levels and development conditions, predevelopment conditions were simulated by removing irrigation withdrawals and return flow and including only the natural sources of recharge and discharge. Maps of water levels for each layer resulting from this simulation (figure 28A-D) are very similar to those from the calibrated model (figure 26A-D), except in the vicinity of irrigation wells in the Fish Springs Ranch area.

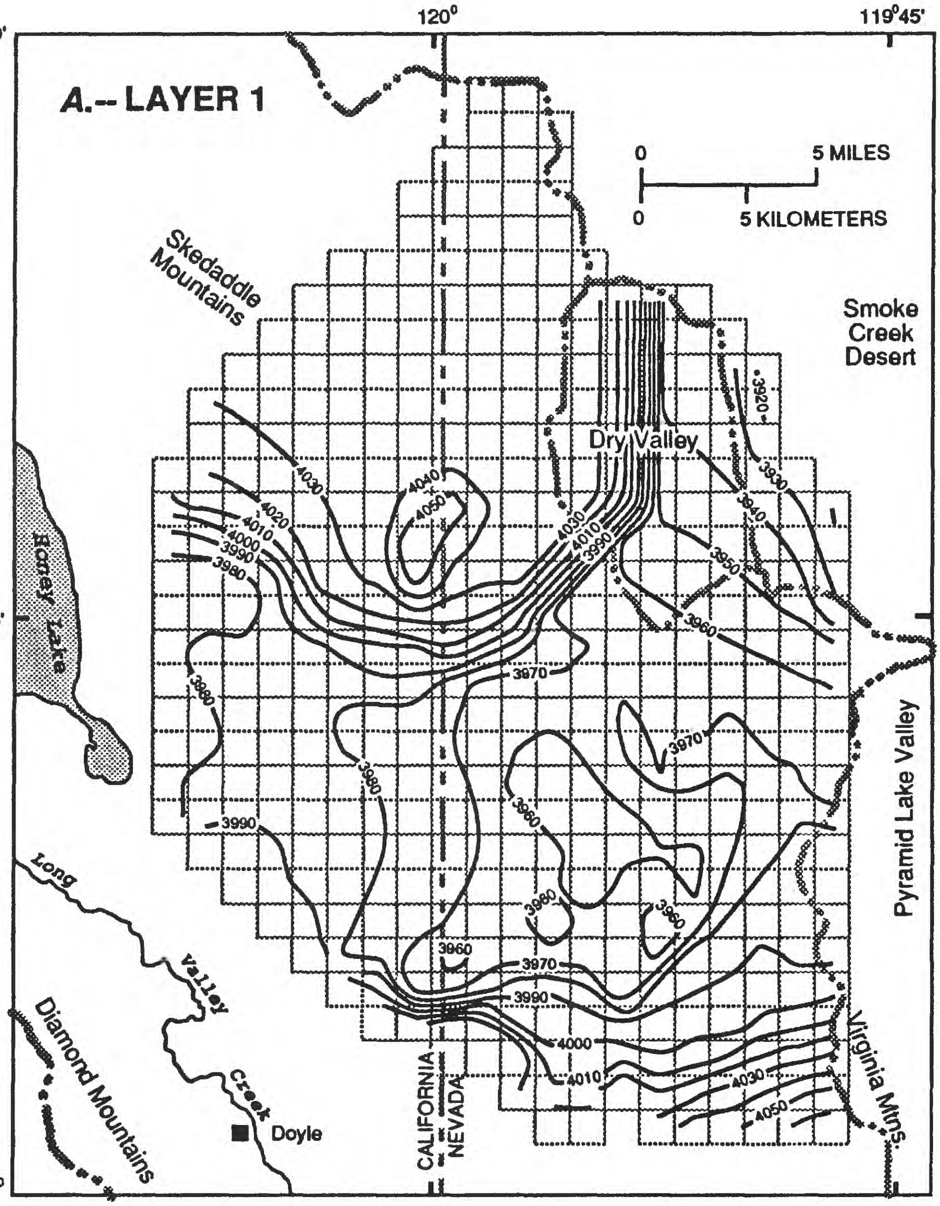
Simulated heads near the five irrigation wells (figure 24) are as much as 25 feet higher than in the stressed system; however, in the rest of the area, head differences generally are less than 5 feet. Simulated predevelopment water levels are above land surface near Fish Springs Ranch, where flowing wells ceased to flow after irrigation began. Simulated net flow eastward across the State line is about 220 acre-ft/yr less than in the calibrated model (table 18). Results of the predevelopment simulation support the assumption that current (1988) development has had only local effects on water levels and small effects on the ground-water flow system; the system is probably near equilibrium in much of the modeled area.

Potential Effects of Increased Pumping

The calibrated model was used to simulate the long-term effects of hypothetical increased withdrawals from the modeled area. Annual withdrawals were increased from about 5,900 acre-ft in the calibrated model to about 15,000 acre-ft in the model representing potential development. This rate was selected to be within the range of amounts proposed by Westpac Utilities (1989, p. IX-57 - IX-58) and The Truckee Meadows Project (1989, p. 1) for development from the aquifer system in the Fish Springs Ranch area. In these proposals, ground water from Honey Lake Valley would be transported out of the basin by pipeline to the Reno-Sparks metropolitan area as a supplemental municipal supply.

For the simulation of increased development, all withdrawals were assumed to be removed from the flow-model area; therefore, no recharge from return flow was included. Recharge from infiltration of precipitation and streamflow were used in the model at the same rates as in the calibrated model. The 15,000 acre-ft of annual pumpage was removed from 18 model blocks representing locations of 5 operating and 13 hypothetical production wells (figure 24). No attempt was made to optimize location and rates of hypothetical wells. All but two of the wells are east of the simulated fault zone in the southern part of the model area. The two wells west of the fault zone represent operating irrigation wells. Each well was assumed to pump from layers 1 and 2 at a constant rate of about 830 acre-ft/yr.

The distribution of ground-water evapotranspiration from the calibrated model and the model simulation of increased development is shown in figure 29A and 29B. No evapotranspiration is simulated in areas where the depth to water exceeds the assigned evapotranspiration extinction depth. These figures indicate how the area from which evapotranspiration occurs could be reduced as an effect of development. If the assumptions for extinction depths are correct, phreatophyte roots would not reach the water table in these areas and the phreatophytes would die.



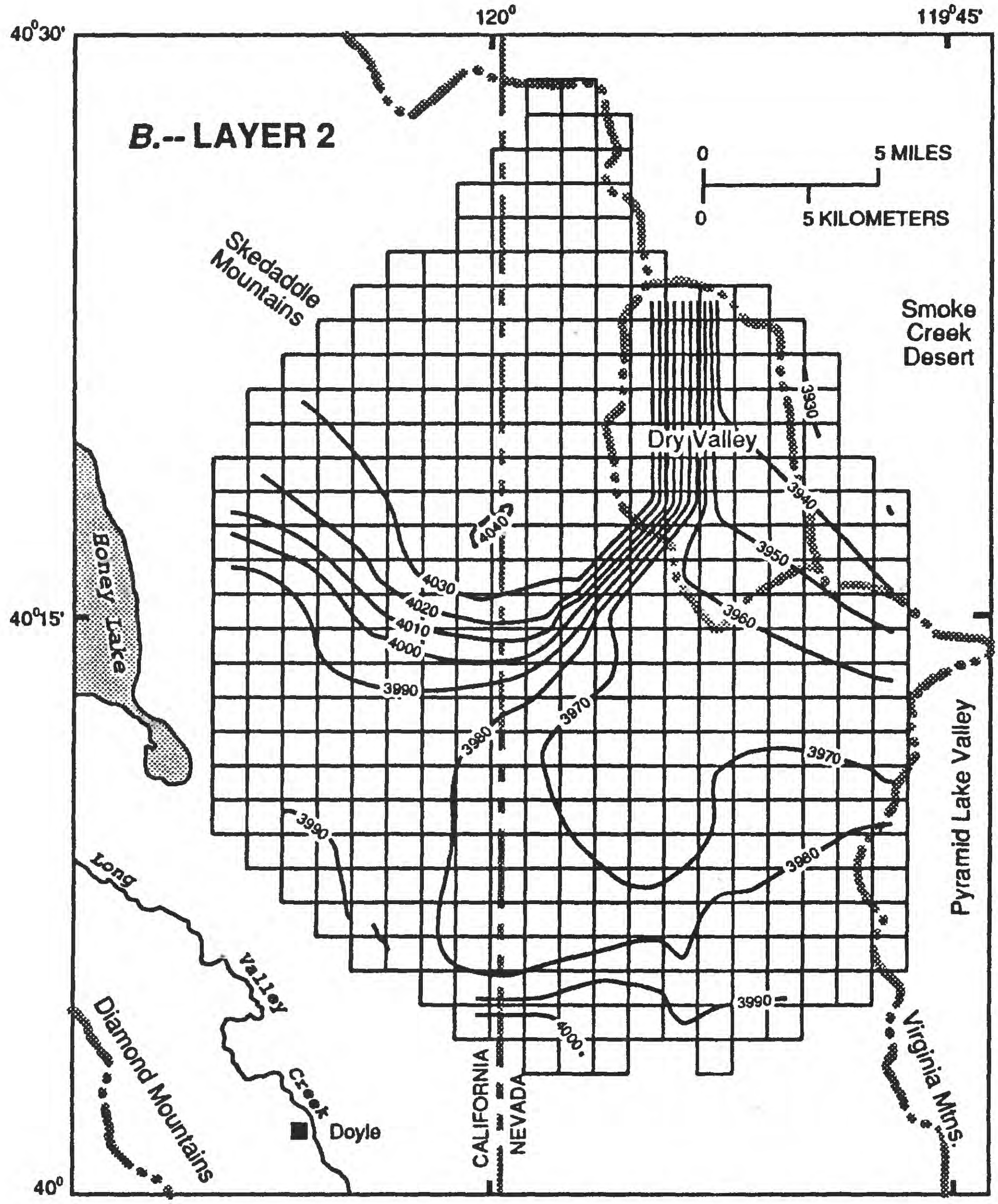
EXPLANATION

MODEL CELL

HYDROGRAPHIC BOUNDARY

SIMULATED WATER-LEVEL CONTOUR -- Shows water-surface altitude, in feet above sea level. Contour interval, 10 feet

FIGURE 28.--Simulated water-level contours for predevelopment conditions in the ground-water model, eastern Honey Lake Valley and adjacent areas: (*A*), model layer 1; (*B*), layer 2; (*C*), layer 3; (*D*), layer 4.



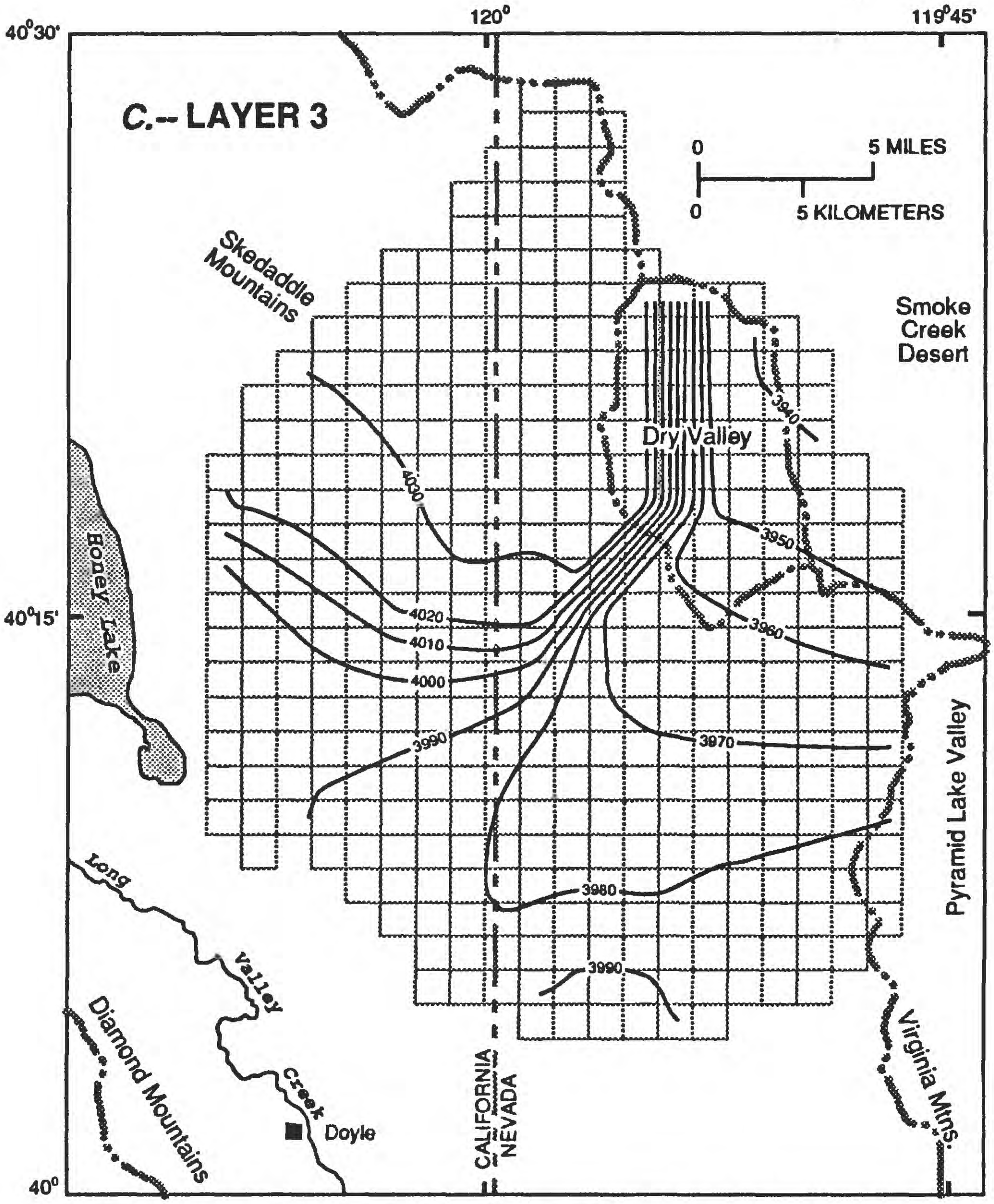
EXPLANATION

MODEL CELL

HYDROGRAPHIC BOUNDARY

4000 SIMULATED WATER-LEVEL CONTOUR -- Shows water-surface altitude, in feet above sea level. Contour interval, 10 feet

FIGURE 28.--Continued.



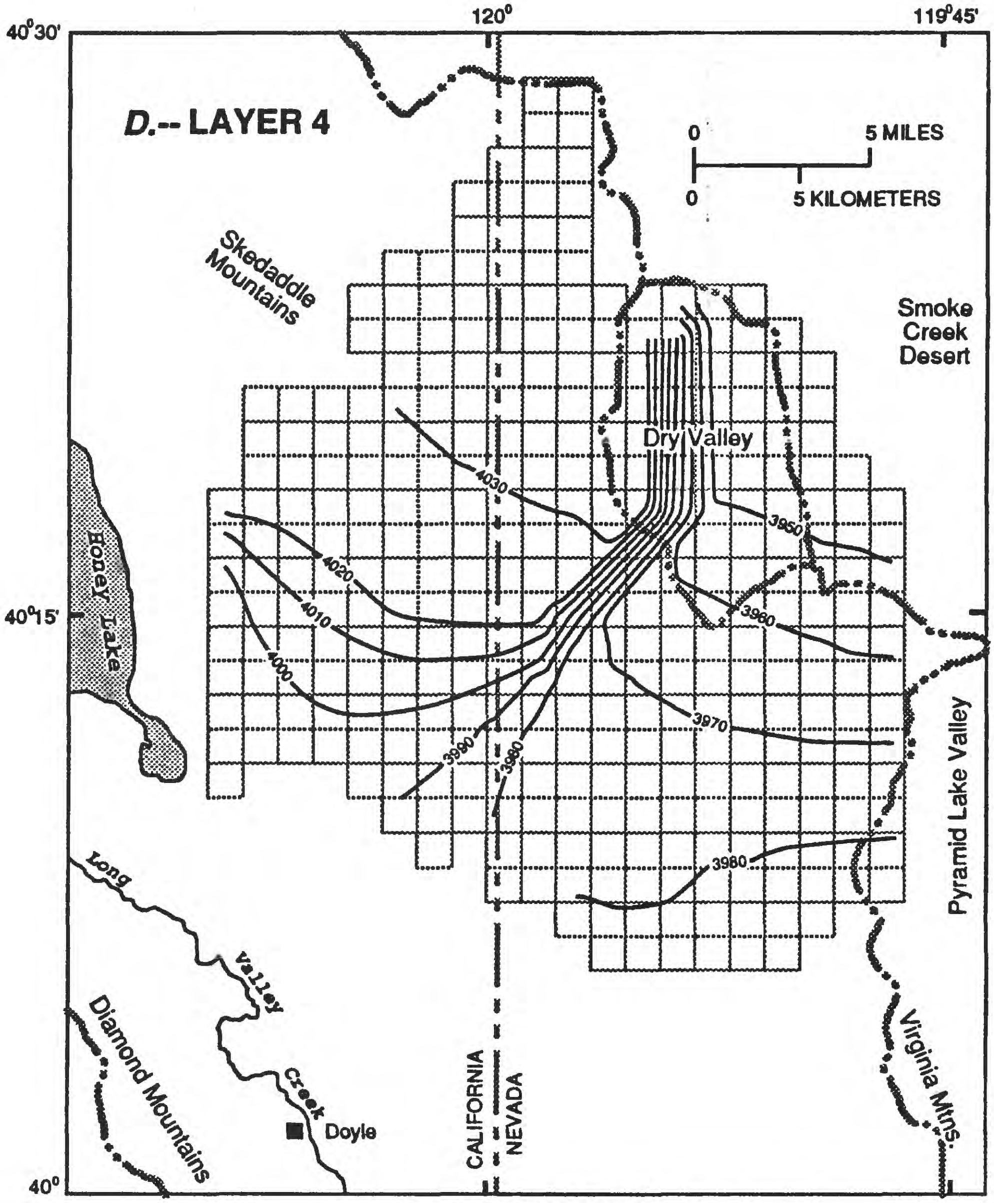
EXPLANATION

MODEL CELL

HYDROGRAPHIC BOUNDARY

4000 SIMULATED WATER-LEVEL CONTOUR -- Shows water-surface altitude, in feet above sea level. Contour interval, 10 feet

FIGURE 28.--Continued.



EXPLANATION

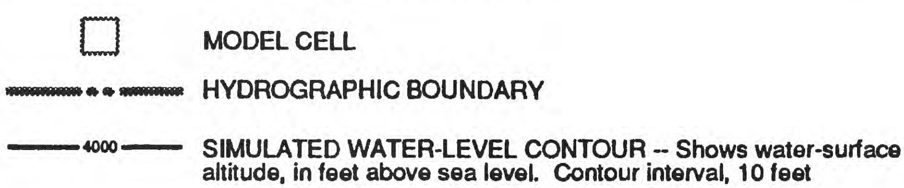


FIGURE 28.--Continued.

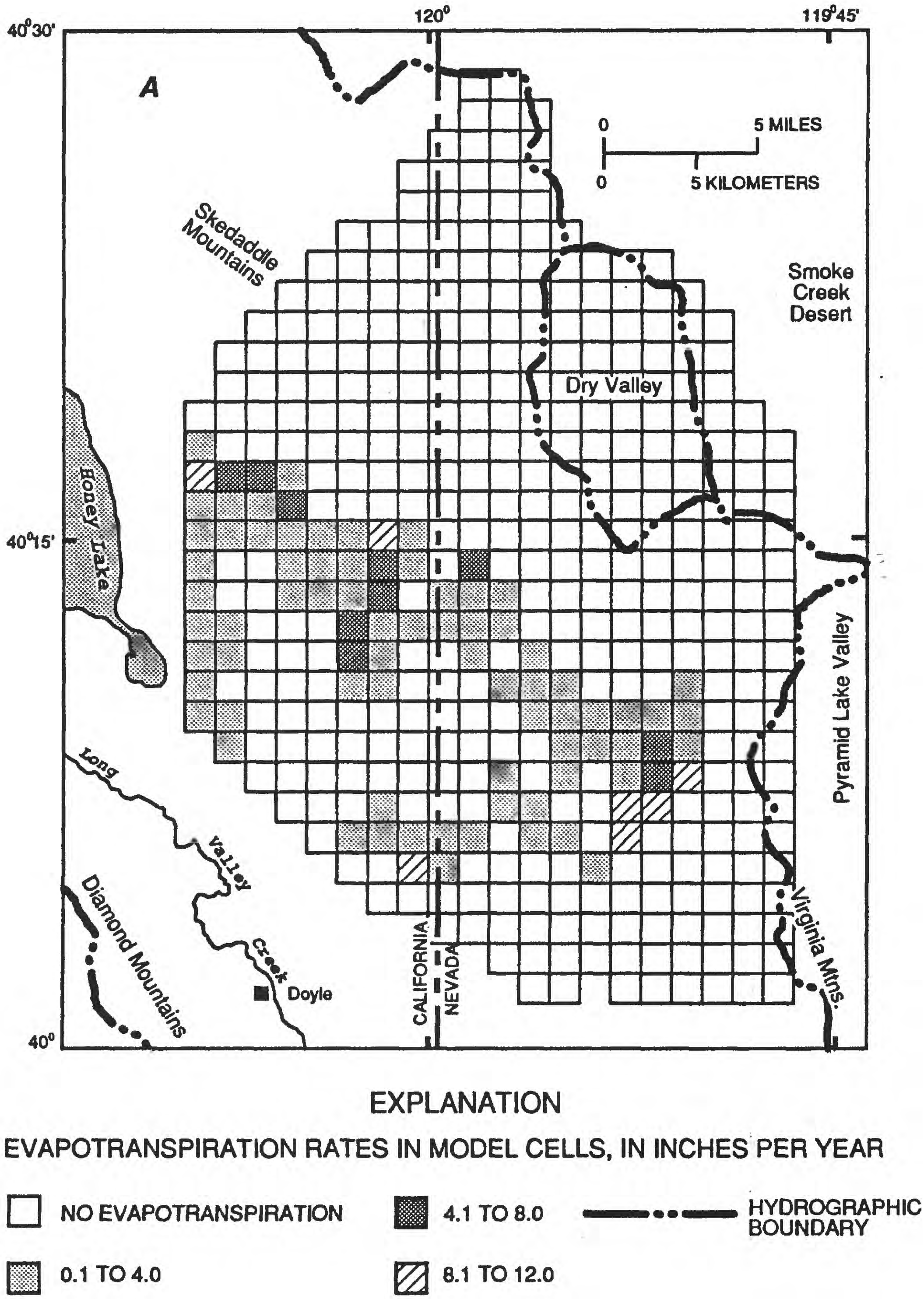
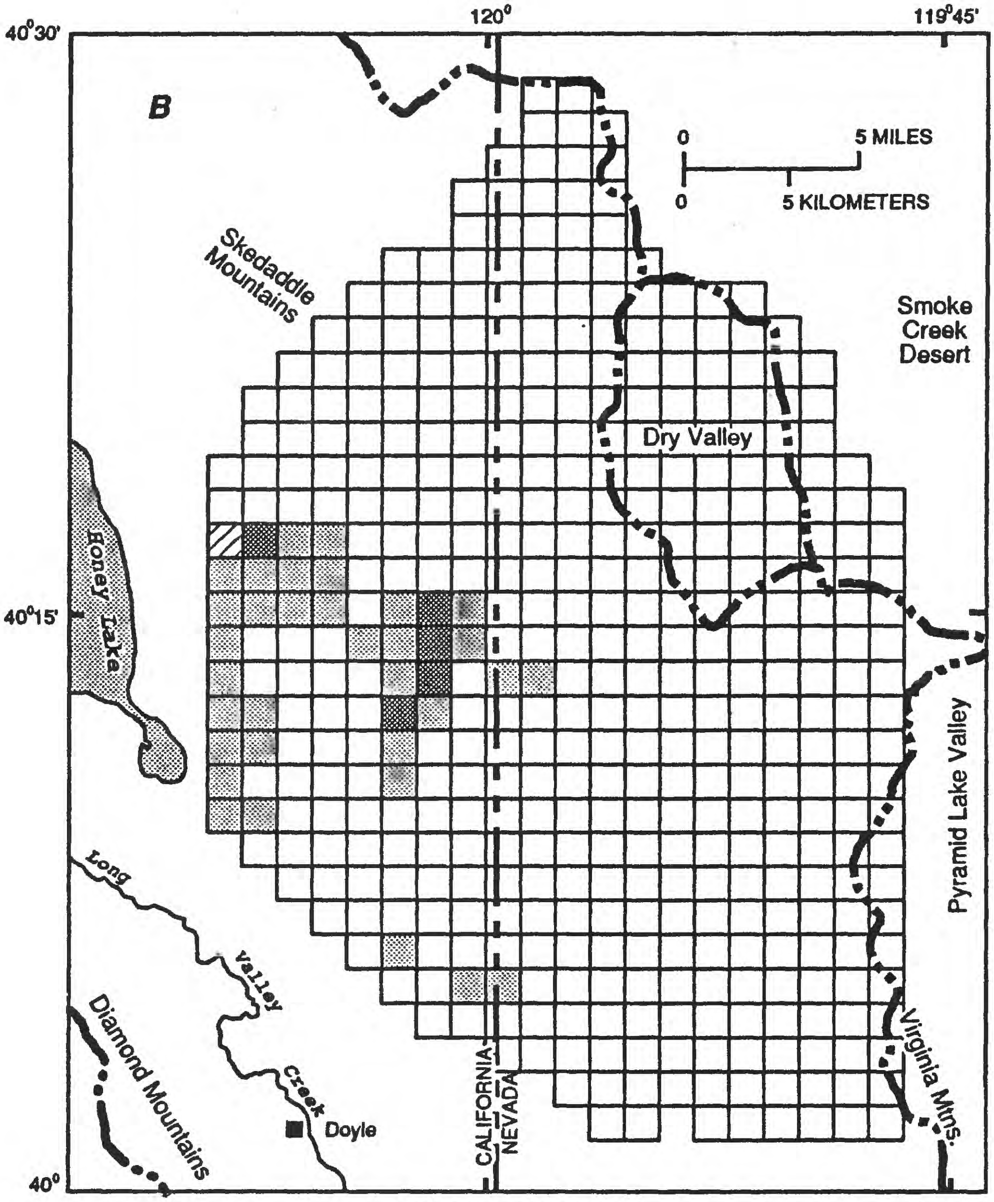


FIGURE 29.--Simulated rates of ground-water evapotranspiration (*A*) at the end of model calibration and (*B*) for hypothetical development conditions in the ground-water model, eastern Honey Lake Valley and adjacent areas.



EXPLANATION

EVAPOTRANSPIRATION RATES IN MODEL CELLS, IN INCHES PER YEAR

NO EVAPOTRANSPIRATION

0.1 TO 4.0

4.1 TO 8.0

8.1 TO 12.0

HYDROGRAPHIC BOUNDARY

FIGURE 29.--Continued.

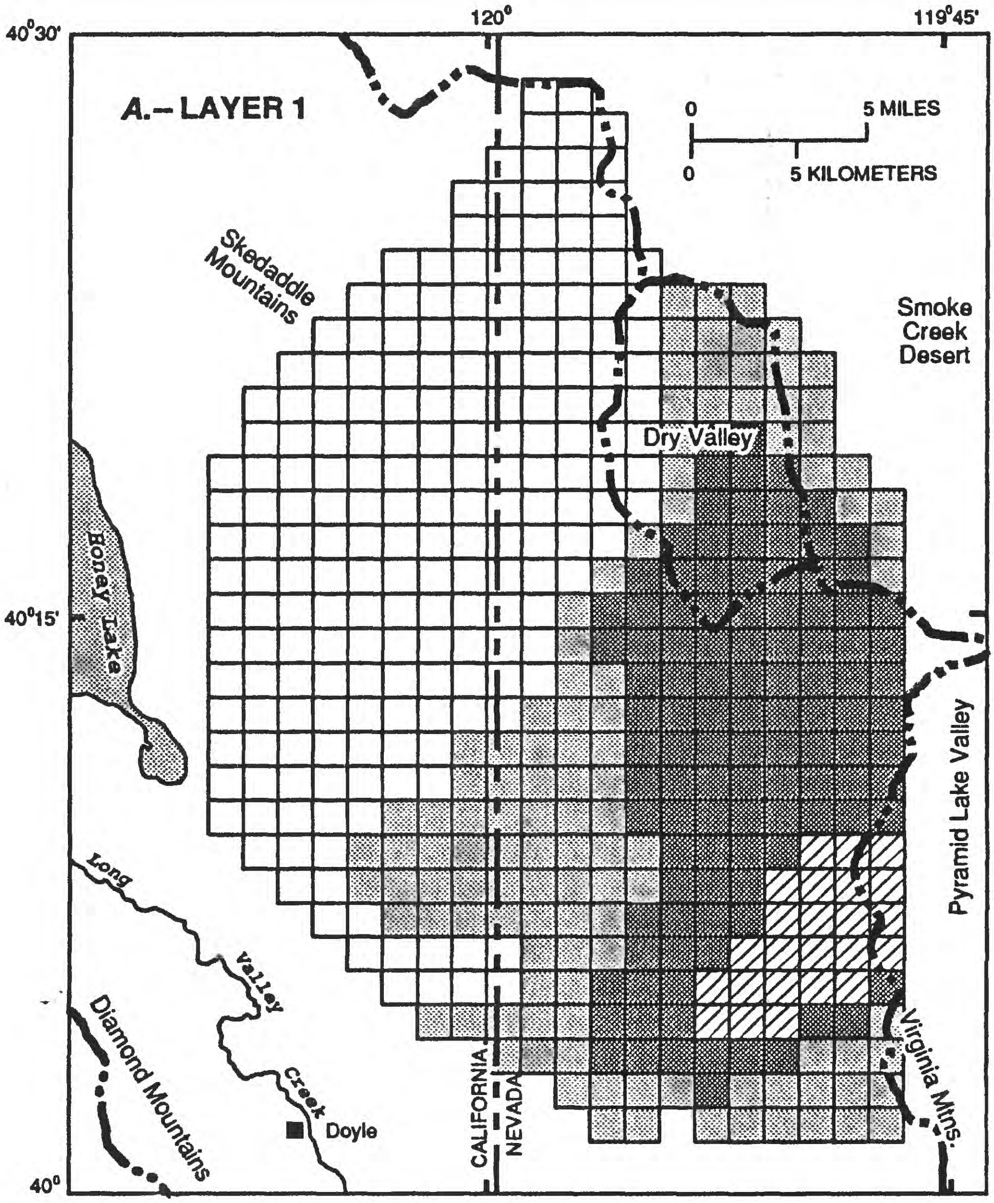
Water-level drawdowns at the end of this simulation range from less than 1 foot along the western boundary in layer 1, to about 100 feet in layers 1 and 2 near Fish Springs Ranch in the vicinity of simulated withdrawals (figure 30A-D), in comparison with current (1988) levels. The maximum simulated drawdown in the California part of the flow-model area ranges from about 10 feet in layer 1 to about 40 feet in layer 4 near Calneva Lake (plate 1). Simulated net flow eastward across the State line is about 1,600 acre-ft/yr more than in the calibrated model (table 18). Simulated effects of development are greater in the lower layers because the volcanic rocks extend farther into the valley with depth (figure 21). The volcanic rocks are more transmissive than the overlying basin-fill deposits. Therefore, the effects of ground-water withdrawals are transmitted farther into the lower layers.

Pathlines (approximations of flow lines) computed from the model results by the method of Pollock (1989) indicate the source of water to pumped wells. Pathlines show that, in the model calibrated to 1988 pumpage, the water comes from the south. Pathlines to the 18 pumped wells in the model simulating increased pumpage show that most of the water comes from the south, but some flow is from the west and northwest.

To determine the effects of simulated fault zones in the proposed-development model, a separate simulation was made. This simulation assumed no reduction in hydraulic conductivity in model blocks representing the faults. As a result of this change, annual ground-water inflow across the western boundary of the model increased about 18 acre-ft (3 percent), outflow across the eastern boundary increased about 290 acre-ft (4 percent), and evapotranspiration decreased about 260 acre-ft (2 percent). Simulated water levels increased east of the fault and declined west of the fault. The maximum change in water level at the State line was an additional decline of about 13 feet.

To determine the effects of the assumed altitude of the Honey Lake surface on flow across the general-head boundary at the western edge of the model, another proposed-development simulation was made. The general-head (lake-surface) altitude was set at 3,983 feet, 3 feet below the level in the calibrated model. As a result of this change, simulated ground-water inflow across the western boundary decreased about 160 acre-ft (25 percent) and the maximum change in water level was an additional decline of about 3 feet at the western boundary of the model.

The steady-state simulation of the ground-water flow system with hypothetical increased pumping indicates the potential long-term effects of development. This simulation was designed to be an example to indicate whether the aquifer system could support ground-water withdrawals at a proposed steady rate. These withdrawals would only be practical if ground-water recharge in fractured volcanic rocks is sufficient to maintain acceptable water levels, yields from basin-fill aquifers are adequate, and effects on water quality and vegetation are acceptable. No attempt was made to optimize the number, location, development schedule, or pumping rates of the wells, except to locate hypothetical wells east of the simulated fault zone in the Fish Springs Ranch area (figure 24).



EXPLANATION

WATER-LEVEL DRAWDOWN IN MODEL CELL

- LESS THAN 10 FEET
- 10 TO 49 FEET
- 50 TO 99 FEET
- 100 FEET OR GREATER
- HYDROGRAPHIC BOUNDARY

FIGURE 30.--Simulated water-level drawdowns for hypothetical development conditions in the ground-water model, eastern Honey Lake Valley and adjacent areas: (*A*), model layer 1; (*B*), layer 2; (*C*), layer 3; (*D*) layer 4.

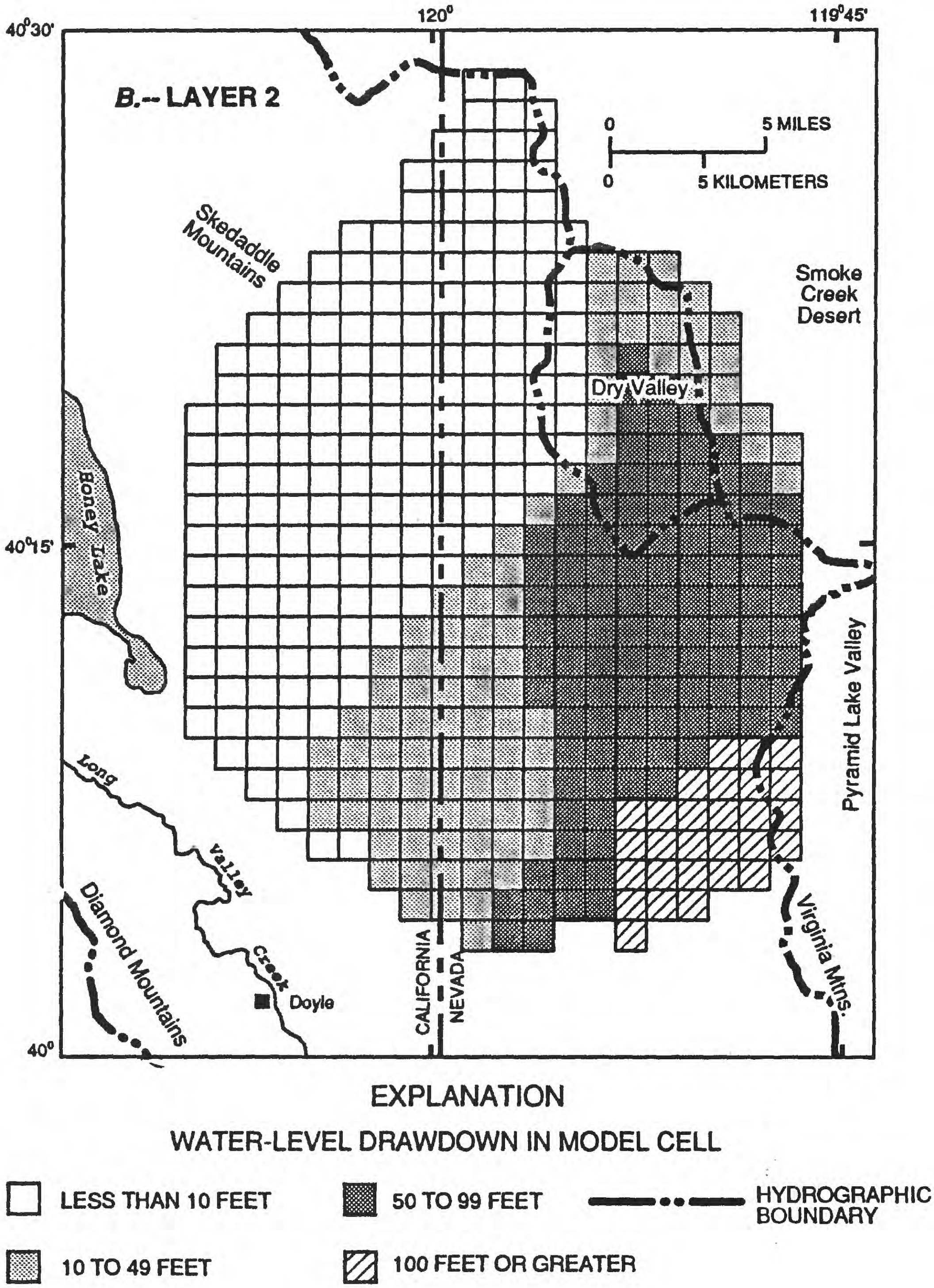
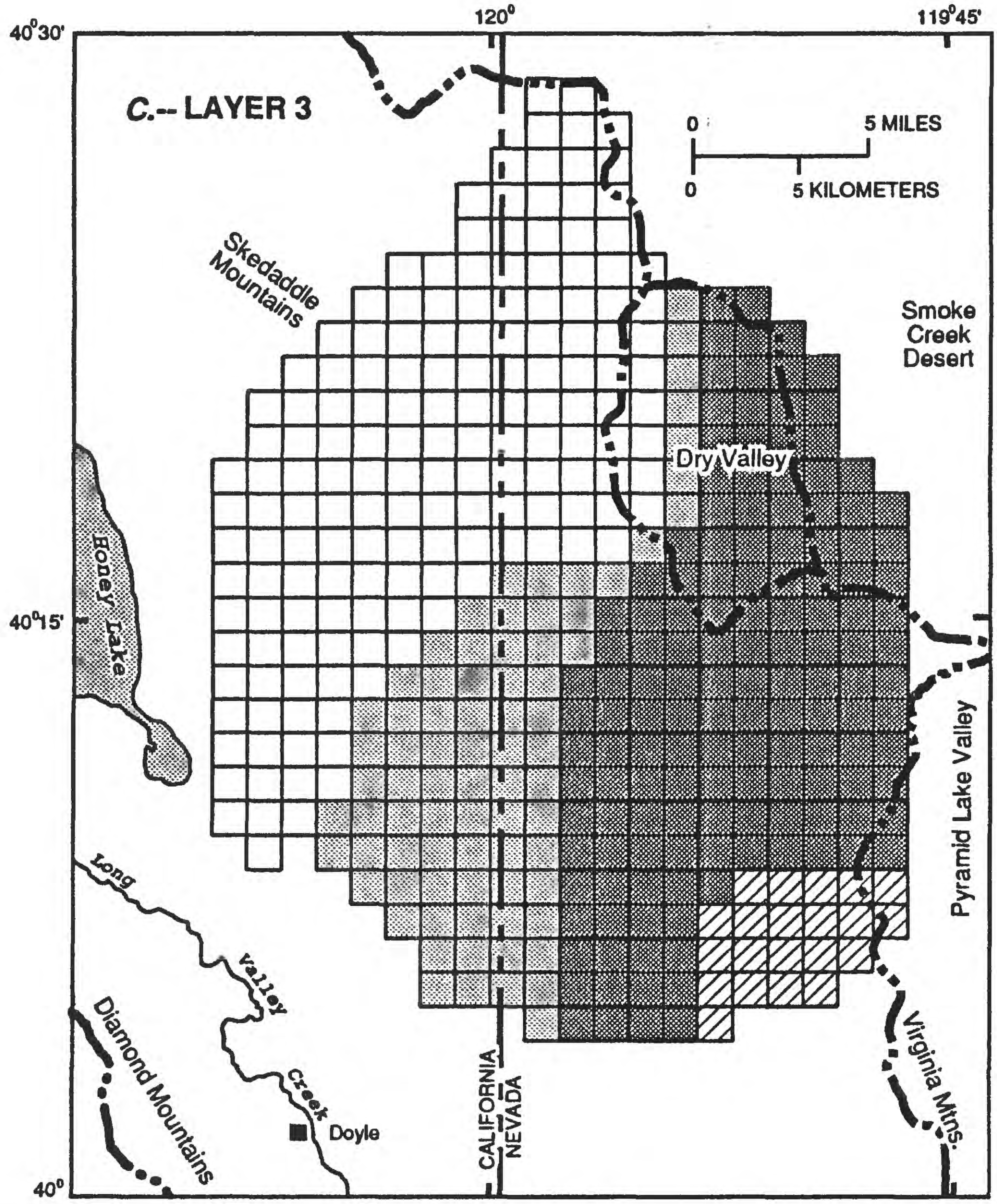


FIGURE 30.--Continued.



EXPLANATION

WATER-LEVEL DRAWDOWN IN MODEL CELL

LESS THAN 10 FEET

10 TO 49 FEET

50 TO 99 FEET

100 FEET OR GREATER

HYDROGRAPHIC BOUNDARY

FIGURE 30.--Continued.

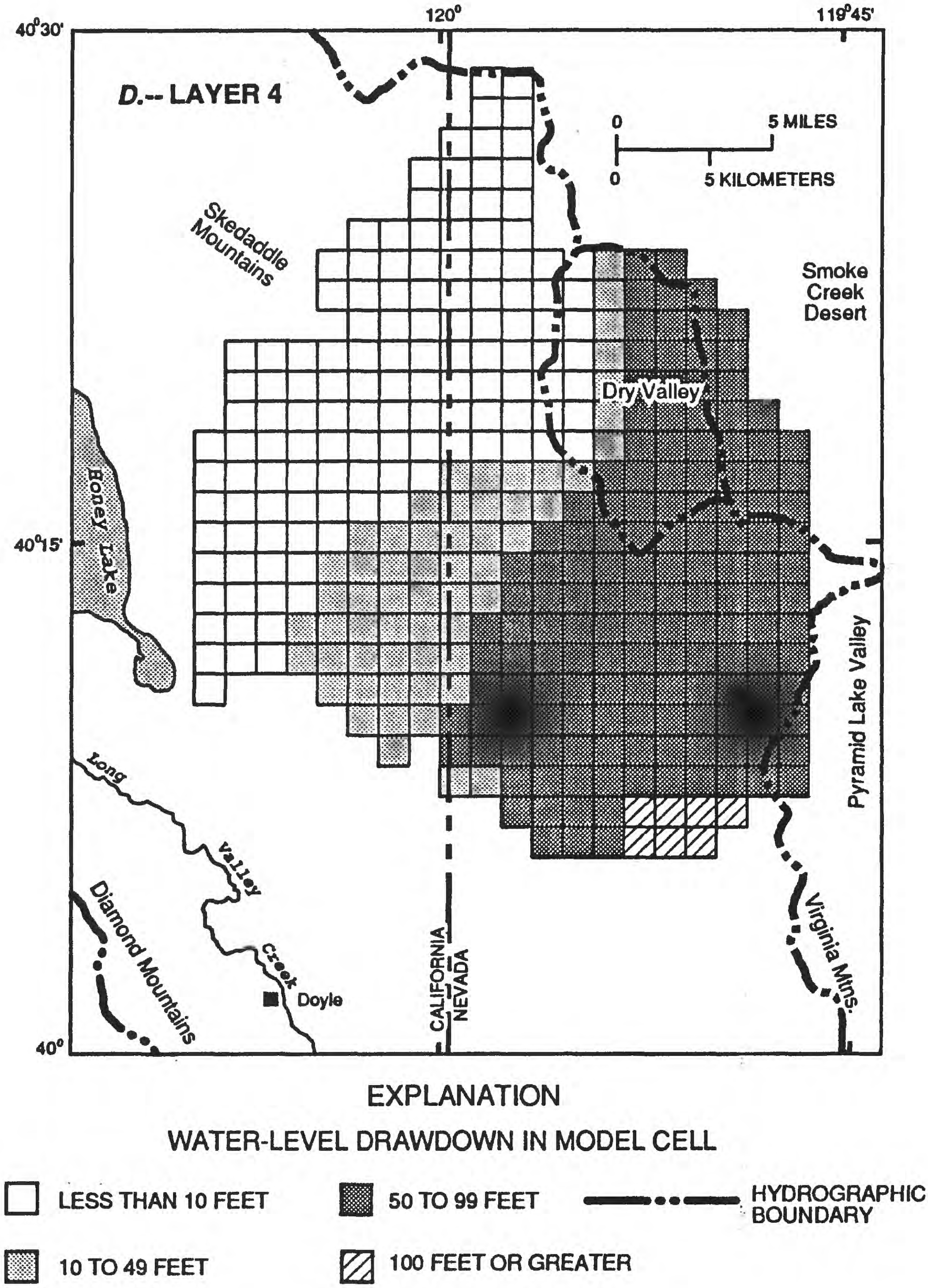


FIGURE 30.--Continued.

Simulated Ground-Water Flow Budgets

The ground-water budget simulated by the calibrated flow model indicates that approximately 24,000 acre-ft of water recharges the system annually under the simulated steady-state conditions (table 19). About 92 percent of this water is derived from precipitation on the model area. The precipitation either infiltrates directly to the water table or runs off from the mountainous areas and infiltrates basin-fill deposits near the mouths of numerous small streams. Part of the recharge modeled as infiltration of precipitation may be derived from subsurface inflow in the southeast part of the basin. The remaining recharge consists of infiltration of irrigation water applied to fields within the basin (about 6 percent of the total recharge) and inflow across the model boundaries (about 2 percent of the total).

In the calibrated flow model, evapotranspiration of ground water accounts for about 45 percent of the annual discharge from the basin. Subsurface outflow from the basin across the model boundaries accounts for another 30 percent of the annual discharge, and irrigation pumpage in the Fish Springs Ranch area accounts for the remaining 25 percent.

The ground-water budget simulated for predevelopment conditions indicates that before development about 23,000 acre-ft of water recharged the aquifer system annually (table 19). Of this total, about 97 percent originated as precipitation on the study area and infiltrated either directly or through streambeds. The remaining 3 percent entered the model area as underflow across the model boundaries from the west and southwest. Evapotranspiration accounted for 66 percent of the simulated annual discharge and 34 percent was discharged annually by subsurface outflow to the east across the model boundaries. Simulated evapotranspiration for predevelopment conditions, about 15,000 acre-ft/yr, is 15 percent greater than the rate estimated for predevelopment conditions from phreatophyte distribution based on Landsat imagery (plate 4 and table 15). At the western and southwestern boundaries, inflow was 4 percent less and outflow was 4 percent greater than in the calibrated model; inflow was within 6 percent of outflow. These results imply that 1988 withdrawals have had only a small effect on subsurface outflow across the eastern boundaries and on subsurface inflow across the western and southwestern boundaries. The results of the predevelopment simulation appear to be satisfactory and, therefore, support the assumptions and estimates on which the calibrated model is based.

The ground-water budget simulated for 15,000 acre-ft/yr of development (table 19) indicates that subsurface outflow across the eastern boundaries would decrease by about 4,000 acre-ft/yr (60 percent) compared to simulated 1988 conditions. Inflow across parts of the western-southwestern boundaries would increase by about 80 acre-ft/yr (14 percent) and outflow across other parts of this boundary would decrease by about 170 acre-ft/yr (29 percent). Net change in flow is a 230 acre-ft/yr increase across the western boundary of the flow-model area. The rate of evapotranspiration from ground water would decrease by about 6,400 acre-ft/yr (58 prcent).

The simulated ground-water budgets indicate that effects of present (1988) development have had primarily local effects on water levels and small effects on the ground-water flow system. They also indicate that the aquifer system could support ground-water withdrawals at a steady annual rate of 15,000 acre-ft, but effects on water levels, ground-water evapotranspiration by native vegetation, and subsurface outflow would be extensive.

TABLE 19.--*Simulated ground-water budgets for the flow-model area*

[Estimated quantities, in acre-feet per year, rounded to two significant figures]

Budget component	Model calibration (1988 withdrawals)	Pre-development (no withdrawals)	Proposed development (hypothetical withdrawals)
RECHARGE:			
Direct infiltration of precipitation[1]	9,200	9,200	9,200
Infiltration of runoff	13,000	13,000	13,000
Irrigation return:			
from surface-water irrigation	0	0	0
from ground-water irrigation	1,500	0	0
Ground-water inflow to model area from Honey Lake area and Long Valley Creek area (in shallow layers of model)	580	570	660
Total recharge	24,000	23,000	23,000
DISCHARGE:			
Ground-water evapotranspiration	11,000	15,000	4,600
Withdrawals from wells	5,900	0	15,000
(Number of simulated wells)	(5)	(0)	(18)
Ground-water outflow from model area westward to Honey Lake area (in deeper layers of model)	590	610	420
Ground-water outflow eastward to Smoke Creek Desert	5,300	5,500	2,000
Ground-water outflow eastward to Pyramid Lake Valley	1,500	1,500	700
Total discharge	24,000	23,000	23,000

[1]Includes 5,000 acre-feet per year that may originate outside the southeast boundary of the basin.

SUMMARY

Honey Lake Valley is a northwest-trending basin on the border between northwest Nevada and northeast California. It is at the junction of three physiographic provinces (the Sierra Nevada, the Basin and Range Province, and the Modoc Plateau) and has hydrologic characteristics of each of these settings.

Major geologic units in the area are basin-fill deposits, fractured volcanic rocks, and granitic bedrock. Unconsolidated basin-fill deposits are composed of layers of gravel, sand, silt, and clay, and include some layers of pyroclastic volcanic material that was deposited in shallow water. The center of the basin contains fine-grained lake deposits; these are surrounded by coarser grained alluvial fan and nearshore deposits. Most ground-water development has been in the coarser deposits to the south and west of Honey Lake. In the northern and eastern parts of the study area, fractured volcanic rocks are important sources of ground water. Volcanic rocks in the Virginia Mountains to the southeast are a principal aquifer near Fish Springs Ranch. Younger volcanic rocks to the north generally are less permeable, except in Willow Creek Valley, where high-yield wells have been developed. The basin is underlain by relatively impermeable granitic bedrock that crops out in the southwestern and southern parts of the study area but, as a result of faulting, is greater than 5,000 feet below land surface in the northeast.

Total ground-water recharge to the study area, estimated from direct infiltration of precipitation, streamflow, and excess irrigation from surface water, is about 120,000 acre-ft. Ground-water recharge, excluding that from infiltration of streamflow from the Susan River and Long Valley Creek, is estimated to be about 95,000 acre-ft/yr by the Maxey-Eakin method, which is based on empirical precipitation-altitude relations. Maxey-Eakin recharge, adjusted for infiltration of water from the Susan River and Long Valley Creek, is estimated as 120,000 acre-ft/yr. Results of a ground-water flow model of the eastern part of the study area indicate that both methods might underestimate ground-water recharge in the southeastern part of the basin, and that annual recharge may be as much as 5,000 acre-ft greater. Some or all of the additional recharge may be supplied by greater infiltration of precipitation than estimated or by subsurface inflow; the origin of this water was not determined. Infiltration of irrigation water withdrawn from wells provides an estimated additional 11,000 acre-ft of recharge per year, for a total recharge of more than 130,000 acre-ft/yr.

At equilibrium, ground-water discharge equals ground-water recharge and storage does not change. Of ground water discharged from the study area, an estimated 85,000 acre-ft/yr is discharged by evapotranspiration from soil and native vegetation, an estimated 53,000 acre-ft/yr is withdrawn from wells, and some water may flow out of the basin through fractured volcanic rocks to Smoke Creek Desert and Pyramid Lake Valley.

The ground-water flow model represents an area of 452 mi^2 in the eastern part of the study area and simulates the system at equilibrium, with about 5,900 acre-ft of water withdrawn annually for irrigation (equivalent to 1988 withdrawals). Mean annual recharge for the model, including irrigation return flow, is about 24,000 acre-ft. Inflow is in approximate balance with outflow across the general-head boundary at the western edge of the modeled area. Subsurface outflow of 6,800 acre-ft discharges across the general-head boundary at the eastern edge.

To simulate natural (predevelopment) conditions, a model of the same area without irrigation withdrawals or return flows was analyzed. The results indicate that before development, water levels were about 25 feet higher in the Fish Springs Ranch area, 4,000 acre-ft more water was discharged by evapotranspiration from native vegetation in the Fish Springs Ranch area, an additional 200 acre-ft of water was discharged by subsurface outflow to the east annually, and simulated net flow eastward across the State line was about 220 acre-ft/yr less, in comparison with simulated current (1988) conditions. With the exception of these differences, which are most apparent in the areas most heavily pumped in 1988, the simulated predevelopment conditions were similar to the 1988 conditions.

As an example of potential future development, the model was used to simulate a ground-water withdrawal rate of 15,000 acre-ft/yr and no return flow of irrigation water. No attempt was made to optimize the number, location, or pumping rates of the hypothetical wells. The results indicate that, if the assumptions on which the model is based are correct, the system would eventually reach a new equilibrium condition in which water levels would decline about 100 feet in the vicinity of the pumped wells, and as much as 40 feet at the California-Nevada State line (compared to 1988 water levels). In addition, net subsurface inflow across the western boundary would increase by about 230 acre-ft/yr, net flow eastward across the State line would increase by about 1,600 acre-ft/yr, evapotranspiration by native vegetation would decrease by about 10,000 acre-ft/yr, and subsurface outflow across the eastern boundary would decrease by about 4,000 acre-ft/yr.

The theoretical maximum volume of water that can be withdrawn from the aquifer system on a sustained basis and eventually result in a new equilibrium is limited by the amount of naturally occurring discharge that can be captured by pumping, plus any additional recharge that can be induced as a result of pumping. In Honey Lake Valley, most perennial streams are far from the proposed pumping areas so that the limit, for practical purposes, would be restricted to the amount of natural discharge that could be captured by pumping. Withdrawals might be further constrained because yields of aquifers composed of fine-grained lakebed sediments may be too low for sustained withdrawals at the maximum rate. The volume of ground water stored in fractured rocks may be small, so that pumping would cause unacceptable drawdowns in some parts of the basin.

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