GROUND-WATER HYDROLOGY AND SIMULATED EFFECTS OF DEVELOPMENT IN SMITH CREEK VALLEY, A HYDROLOGICALLY CLOSED BASIN IN LANDER COUNTY, NEVADA

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1409–E

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By JAMES M. THOMAS, STEPHEN M. CARLTON, *and* LAWRENCE B. HINES

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) Program was started in 1978 following a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which in aggregate underlie much of the country and which represent an important component of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and accordingly transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system and the changes brought about in it by human activities, and to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number, and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400 and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

Dallas L. Peck
Director

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[Plates are in pocket]

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CONVERSION FACTORS AND ABBREVIATIONS

Inch-pound units of measure used in this report may be converted to International System of units (SI) units by using the following factors:

Multiply inch-pound units	By	To obtain SI units
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)
foot (ft)	0.3048	meter (m)
square foot (ft^2)	0.09290	square meter (m^2)
cubic foot (ft^3)	0.02832	cubic meter (m^3)
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
foot per second (ft/s)	0.3048	meter per second (m/s)
inch (in.)	25.40	millimeter (mm)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
degrees Fahrenheit (°F)	°C=(°F−32)/1.8	degrees Celsius (°C)

REGIONAL AQUIFER-SYSTEM ANALYSIS—GREAT BASIN, NEVADA-UTAH

GROUND-WATER HYDROLOGY AND SIMULATED EFFECTS OF DEVELOPMENT IN SMITH CREEK VALLEY, A HYDROLOGICALLY CLOSED BASIN IN LANDER COUNTY, NEVADA

By JAMES M. THOMAS, STEPHEN M. CARLTON, and LAWRENCE B. HINES

ABSTRACT

Smith Creek Valley is one of 14 hydrologically closed, single-valley ground-water flow systems in the Great Basin. Gravity data indicate that the basin-fill aquifer is a complex bowl-shaped structure with a depression 5,500 feet deep beneath the main playa near the center of the basin and a depression 3,000 feet deep in the north part of the basin. Ground water recharges the basin-fill aquifer around the perimeter and flows toward a topographic low (playa) near the center, where it is discharged by evaporation from bare soil and transpiration from phreatophytic plants. Horizontal hydraulic gradients range from more than 100 feet per mile around the margin of the basin to only 1 foot per mile near the central part. An upward vertical hydraulic gradient in the discharge area is as much as 0.20 foot per foot in the top 75 feet of predominantly fine-grained basin-fill sediments.

Hydraulic conductivities of the basin-fill deposits range from about 1×10^{-7} foot per second for fine-grained playa materials (primarily silts and clays with thin sand stringers) to about 4×10^{-4} foot per second for coarse-grained deposits (primarily sands and gravels with thin lenses of silt and clay). Storage coefficients range from 0.00007 for a 75-foot-thick confined interval to 0.15 for coarse-grained deposits in the 50-foot-thick water-table interval.

Recharge was estimated to be 9,600 acre-feet per year by the Maxey-Eakin method and 8,300 acre-feet per year by a chloride-balance technique. Both estimates are based on a refined altitude-precipitation relationship for the valley. Natural discharge by evapotranspiration was estimated to be 8,300 acre-feet per year, on the basis of detailed phreatophyte mapping. About 200 acre-feet per year of flow from hot springs in the playa area is included in this estimate. In 1983, an additional 650 acre-feet was pumped for irrigation.

Calibration of a ground-water flow model showed that water levels throughout the basin are controlled by water levels in the discharge area, which in turn are controlled by the rate of natural discharge. Hydraulic gradients are controlled by grain-size and recharge distributions. Areas containing predominantly fine-grained deposits, or receiving large amounts of recharge, have steeper gradients than areas having predominantly coarse-grained deposits or small amounts of recharge.

The flow model was used to simulate effects of development under different hypothetical scenarios. Each simulation was for 600 years—300 years of pumping and 300 years of recovery. Initially, pumpage was set equal to the estimated average annual recharge and was distributed areally to efficiently reduce the natural discharge. For this scenario, as much as 96 percent of the ground water that would otherwise be discharged was captured and the maximum simulated drawdown was 20 feet averaged over a 2-square-mile model block. Doubling the pumping rate captured all ground water discharged by evapotranspiration in less than 25 years and produced block-wide drawdowns locally of more than 200 feet. When the pumping rate was varied by first assigning the value at twice the estimated recharge rate for 50 years and then reducing the rate to that of the estimated recharge for the next 250 years, all ground water discharged by evapotranspiration was captured in less than 25 years and the maximum block-wide drawdown was about 35 feet. When the pumping was located near one end of the basin and set at a rate equal to recharge, average drawdowns exceeded 100 feet and less than 75 percent of the natural discharge was captured. The other two pumping scenarios, both with pumping rates set equal to the recharge rate and concentrated near the discharge area, were relatively efficient in reducing natural discharge, and produced maximum block-wide drawdowns of less than 50 feet.

Model simulations indicate that natural discharge can be efficiently reduced with a maximum block-wide drawdown of less than 20 feet by pumping at the rate of recharge, if pumping wells are optimally located. Doubling or varying the pumping rate resulted in more rapid reduction of natural discharge but produced greater simulated water-level declines. Moving the pumping wells to more concentrated centers near the discharge area resulted in less efficient reduction of natural discharge and produced greater simulated water-level declines than when they were optimally located.

All six hypothetical development scenarios resulted in water-level declines and reduced natural evapotranspiration during pumping. All six showed water-level recoveries and increasing natural discharge after pumping ceased. However, the extent of water-level declines and the rate of natural-discharge reduction depended on the areal distribution and rate of pumping. The subsequent recovery of water levels depended on the location of pumping, the depth of drawdown in the discharge area, and the cumulative amount of water removed from storage. Pumping at greater rates, or in more concentrated centers, or in areas away from major sources of recharge produced greater drawdowns throughout the basin.

INTRODUCTION

Smith Creek Valley was studied as part of the Great Basin Regional Aquifer-System Analysis (RASA). The objectives of the Great Basin RASA study are described by Harrill and others (1983, p. 2–3).

PURPOSE AND SCOPE

The primary purpose of the Smith Creek Valley study was to investigate the geohydrology and to analyze ground-water flow in a hydrologically closed, single-valley system. Fourteen of the identified 39 major flow systems in the Great Basin are closed single valleys, so Smith Creek Valley was selected to be studied as a type area for these closed single-flow systems (Harrill and others, 1983). In these closed valleys, all recharge to and discharge from the ground-water aquifer occur within the topographic basins, because low-permeability rocks compose the surrounding mountain blocks and underlie the basin-fill deposits, and there are no surface outflow drainages. Knowledge gained about hydrologic processes in Smith Creek Valley should be transferable to similar valleys elsewhere in the Great Basin.

The study entailed collecting and evaluating field data; determining the extent, thickness, and hydraulic properties of the basin-fill aquifer; reevaluating earlier recharge and discharge estimates, modeling the ground-water flow system; and evaluating potential effects under different hypothetical development scenarios. Field work consisted of locating wells, measuring water levels, mapping phreatophytes in detail, inventorying pumpage, and determining the location and extent of significant geologic and hydrologic features.

Geologic and hydrologic data were evaluated to gain an understanding of the geometric and hydraulic properties of the basin-fill aquifer and the flow system. A three-dimensional, finite-difference ground-water flow model was used to simulate the flow system, to help characterize hydrologic processes, and to evaluate hydraulic properties of the ground-water flow system.

LOCATION AND GENERAL FEATURES

Smith Creek Valley is in west-central Nevada, approximately 125 mi east of Reno (fig. 1). The valley encompasses 583 mi^2. It is bounded on the west by the Desatoya Mountains, on the north by the New Pass Range, and on the east by the Shoshone Mountains, all of which are block-faulted masses composed primarily of Tertiary volcanic rocks. Ione Valley bounds Smith Creek Valley to the south. However, a ground-water divide occurs near the southern boundary of the study area, along a line of volcanic rock outcrops that protrude through a thin accumulation of basin-fill deposits (fig. 1). Ground water south of this divide flows into adjacent Ione Valley.

Land-surface relief in the Smith Creek topographic basin is greater than 4,000 ft. North Shoshone Peak reaches an altitude of 10,313 ft above sea level, and several peaks in both the Shoshone and Desatoya Mountains are over 9,000 ft in altitude. Smith Creek playa is the lowest topographic feature in the valley and has an altitude of approximately 6,035 ft.

PREVIOUS STUDIES

Published reports for Smith Creek Valley include a ground-water reconnaissance by Everett and Rush (1964) and geophysical reports by Herring (1967) and Salehi (1967). In addition, some data for Smith Creek Valley are listed in a State water-planning report (Nevada State Engineer, 1971). In the 1960's a significant amount of as yet unpublished data was collected in the central part of the basin by F. E. Rush, D. E. Everett, and J. R. Harrill of the U.S. Geological Survey. Their work included drilling, logging, and monitoring observation wells; leveling to obtain well and playa-surface altitudes; core sampling for laboratory analysis; and some water-quality sampling. These data are available at the U.S. Geological Survey office in Carson City, Nev.

WELL AND SPRING INVENTORY

Wells were located, water levels measured, and drillers' logs were examined, when available. Data on well construction and other pertinent information were entered into the U.S. Geological Survey WATSTORE (Water-data storage and retrieval system) data base. Springs and stream-sampling sites were also entered into the WATSTORE data base. All basic data are available at the U.S. Geological Survey office in Carson City, Nev.

Site locations are stored in the WATSTORE data base by latitude and longitude and by township, range, and section. The numbering of sites on the basis of township, range, and section uses the rectangular subdivision of land, referenced to the Mount Diablo base line and meridian. The numbering scheme consists of three units: (1) the township north of the base line, (2) the range east of the meridian, and (3) the section number. The section number is followed by as many as four letters that indicate quarter sections, quarter sections thereof, and so on, within the numbered section. The letters A, B, C, and D designate the northeast, northwest, southwest, and southeast quarters, respectively (fig. 2). A number following the letters indicates the sequence in which each well or spring in the section was recorded.

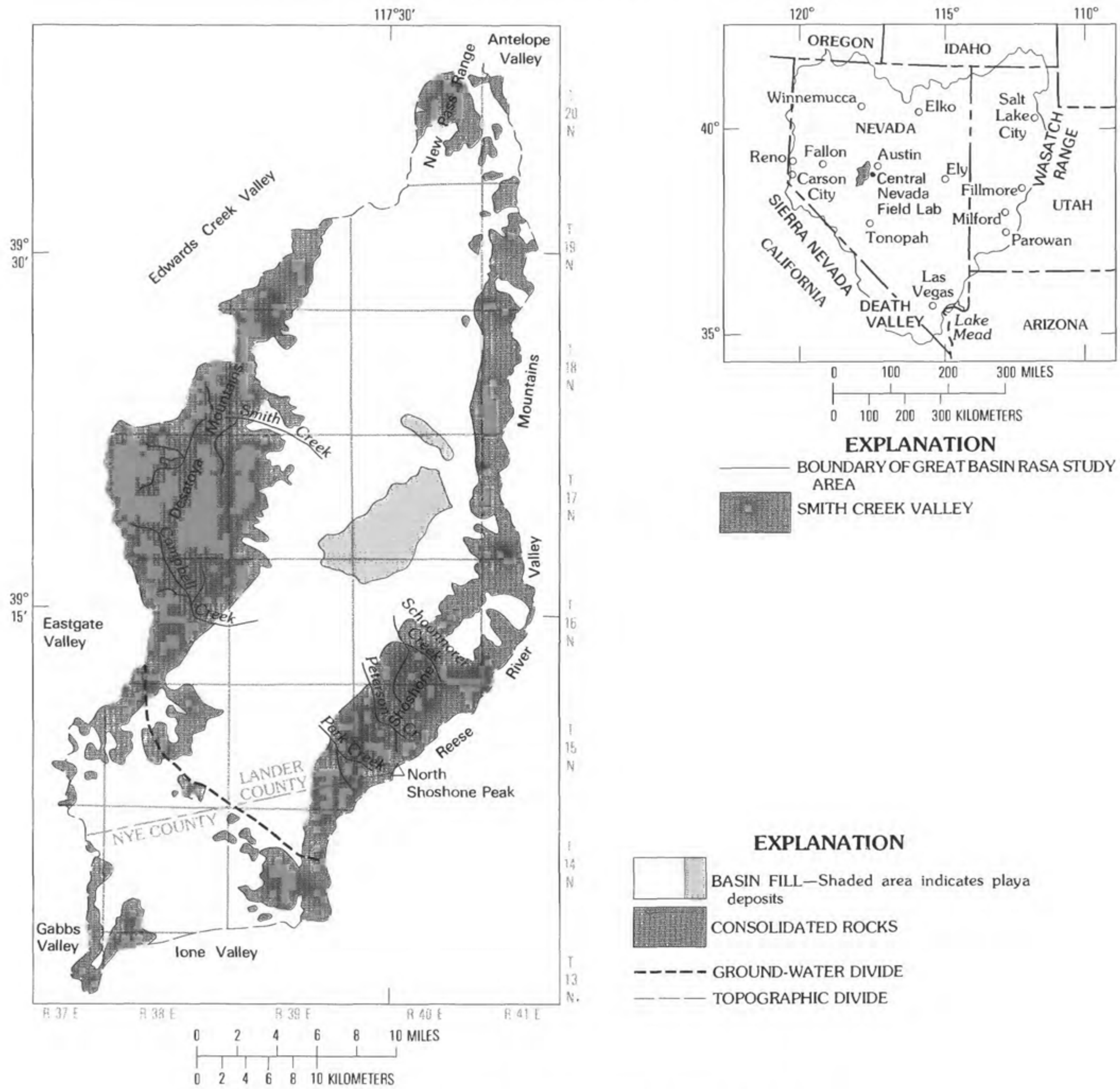


FIGURE 1.—Location and general features of Smith Creek Valley.

ACKNOWLEDGMENTS

The authors thank Hunt Oil Company of Nevada for supplying lithologic logs from geothermal temperature-gradient holes and Lori J. Connolly of the U.S. Geological Survey for her work on the illustrations in earlier drafts of this report.

GENERAL GEOLOGY

LITHOLOGY

For the purposes of this study, the rocks in Smith Creek Valley have been grouped into two main geologic units: Consolidated rocks and basin-fill deposits.

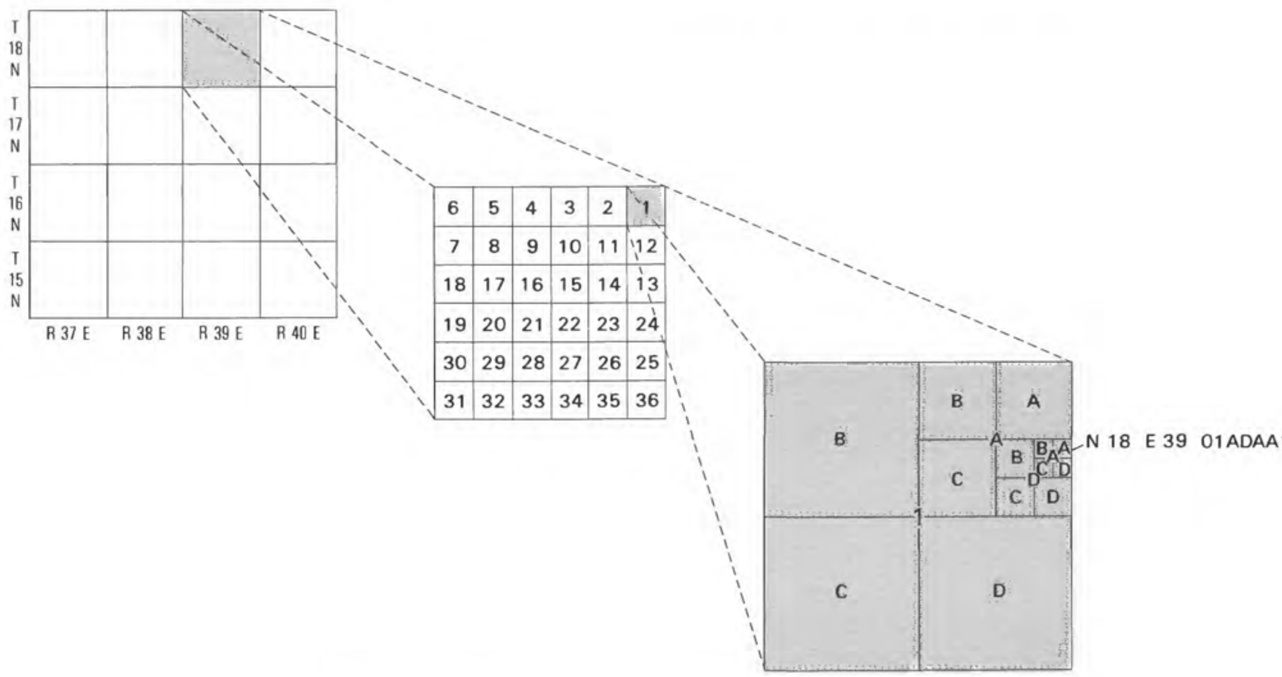


FIGURE 2.—Numbering system for wells and springs in Nevada.

Basin-fill deposits, in turn, have been further subdivided into playa and nonplaya deposits. Consolidated rocks form mountain blocks on the west, north, and east boundaries of the basin and underlie the basin fill (pl. 1). The consolidated rocks are predominately Tertiary rhyolite flows, ash flows, crystal-poor to crystal-rich ash-flow tuffs, and intrusive rocks (Stewart and McKee, 1977; Kleinhampl and Ziony, 1967).

Basin-fill deposits are composed primarily of Tertiary and Quaternary alluvial, colluvial, eolian, and lacustrine sediments. Thin layers of volcanic ash and minor amounts of chemically precipitated minerals and coatings are also present in the basin-fill deposits. Playa deposits are primarily Quaternary lacustrine sediments. The basin-fill deposits, which originated by erosion of adjacent mountains, range in grain size from clay in the playa to boulders in the alluvial fans. They generally grade in size from coarse near the mountain blocks to fine towards the center of the basin. Tertiary to lower Quaternary basin-fill deposits occupy the deeper parts of the basin and are exposed locally on the upper parts of alluvial fans around the margin of the basin. There they are highly dissected and are composed of poorly sorted gravel, sand, silt, and clay. Younger Quaternary basin-fill deposits overlie the Tertiary to lower Quaternary deposits in most of the basin and are composed of better sorted gravel, sand, silt, and clay (Everett and Rush, 1964). Quaternary lacustrine deposits formed during periods when a lake occupied the topographically lowest part of the basin near the center of the valley (Everett and Rush, 1964).

Lithologic maps and geologic cross sections for the basin-fill deposits were constructed from interpretations of well logs. The U.S. Geological Survey and Hunt Oil Company of Nevada drilled most of the wells in the basin. Consequently, terminology reported on most well logs is fairly consistent, and the deposits can be grouped according to grain size by using a standard set of terms to interpret the logs (Baker and Foulk, 1975, p. 62–66). Basin-fill deposits have been grouped into fine-grained, coarse-grained, and mixed deposits on the basis of the dominant lithology in each unit. The lithology included in each group is shown in table 1. Fine-grained deposits include some thin beds of gravel and sand, and coarse-grained deposits include some thin beds of silt and clay. Mixed deposits consist primarily of poorly sorted mixtures of clay, silt, sand, and gravel around the periphery of the valley floor and largely interbedded, finer and coarser layers surrounding the fine-grained playa deposits.

Areas where fine-grained, coarse-grained, and mixed deposits dominate for the intervals of 0–50 and 50–500 ft within saturated basin fill below the 1982 water table are shown in figure 3. The two maps in figure 3 were constructed by determining the average grain size from each available well log. The average grain size was ascertained by multiplying the thickness of a bed within the interval by an arbitrarily assigned value (fine-grained,

TABLE 1.—*Lithologic terms used in drillers' logs, grouped according to grain size*
[Modified after Baker and Foulk, 1975, p. 62-66]

Coarse-grained deposits	Fine-grained deposits	Mixed deposits[1]
Boulder	Anhydrite	Alluvium
Boulders and sand	Bentonite	Boulders, silt, sand, and clay
Cemented gravel	Caliche	Cobbles, sand, silt, and clay
Cobbles	Clay	Clayey sand
Cobbles and sand	Claystone	Gravel, sand, and silt
Conglomerate	Evaporite	Gravel, sand, and clay
Gravel	Gypsum	Sandy silt
Rubble	Limestone	Soil
Sand	Mud	
Sand and gravel	Mudstone	
Sandstone	Sandy clay	
	Shale	
	Silt	
	Siltstone	
	Silty clay	
	Tuff	
	Volcanics	

[1] Mixed deposits also include sequences of thinly interbedded coarse- and fine-grained deposits.

1; mixed, 2; and coarse-grained, 3), totaling the values for all beds in the interval, and dividing by the total thickness of the interval. These values were then plotted and contoured to produce maps as shown in figure 3 depicting areas of predominately fine-grained, mixed, and coarse-grained deposits for each interval.

Maps for both intervals show an area of fine-grained deposits beneath the playa that grades into a narrow band of mixed deposits around the margin of the playa and then into a coarse-grained area encompassing the central part of the basin outside the area of the playa. Few well logs are available around the margin of the basin, but surface observations indicate that this area consists mainly of coalescing alluvial fan deposits that generally contain poorly sorted mixtures of clay, silt, sand, and gravel except where perennial streams have deposited well-sorted alluvium.

The areas characterized by fine-grained deposits coincide closely with the area of the playas. However, the areal extent of fine-grained deposits for the 50- to 500-ft interval beneath the main playa is smaller than that of the 0- to 50-ft interval. Basin-fill deposits are interbedded finer and coarser layers beneath the small northern playa in the deeper interval. In addition, west of the fine-grained deposits in the 50- to 500-ft interval, coarse-grained deposits, which form a continuous unit in the 0- to 50-ft interval, are dissected by mixed deposits.

Lithologic sections *A–A'* and *B–B'* (figs. 4 and 5) show the vertical distribution of fine-grained, coarse-grained, and mixed deposits in the basin fill. Locations of these sections are shown in figure 3. The sections were constructed by plotting grain size from interpretations of individual well logs, and then correlating, where possible, the grain-size distributions from well to well. Both sections show mixed deposits grading abruptly into fine-grained playa deposits. The vertical and areal extent of the fine-grained playa deposits is unknown, but near the playa center they extend to a depth of at least 150 ft, and in the uppermost 100 ft, their areal extent decreases with depth. Section *B–B'* (fig. 5) shows the depositional sequence from near the consolidated-rock and basin-fill contact downslope to the playa. Basin-fill deposits change eastward from mixed deposits that overlie bedrock beneath the alluvial fan on the west side of the valley, to coarse-grained deposits 1 to 2 mi from the consolidated-rock and basin-fill contact, to mixed deposits around the margin of the playa, and finally to fine-grained deposits in the center of the basin.

GEOLOGIC STRUCTURE

The northeast-trending mountains that border the Smith Creek alluvial basin were uplifted along normal faults during the Tertiary and Quaternary Periods. These faults are the result of extensional stresses that developed during the last 17 million years (Stewart, 1980, p. 5), the period during which the present-day physiography of the Basin and Range province was formed. Stratigraphic units within the mountains are also faulted (pl. 1), forming blocks that dip gently eastward (Stewart and McKee, 1977).

Faulting has also resulted in an area of hot springs within the basin fill on the west side of the valley and has caused tilting of the playa (F. E. Rush and James R. Harrill, U.S. Geological Survey, written commun., 1966). The faults in the hot-spring area may act as conduits for upward flow of water or as barriers to ground-water flow that cause upward movement of the water.

The Smith Creek playa slopes gently westward because of Holocene downfaulting near the western edge of the playa (pl. 1). The east half of the playa slopes westward about 1 ft/mi. The west half also slopes westward but at a lesser gradient because of a depositional wedge that has formed on the lowest area of the playa since initial tilting of the nearly horizontal playa surface.

GENERAL HYDROLOGY

CLIMATE

Smith Creek Valley has a semiarid to subhumid climate, and the average yearly precipitation ranges from about 6 in. on the basin floor to more than 20 in.

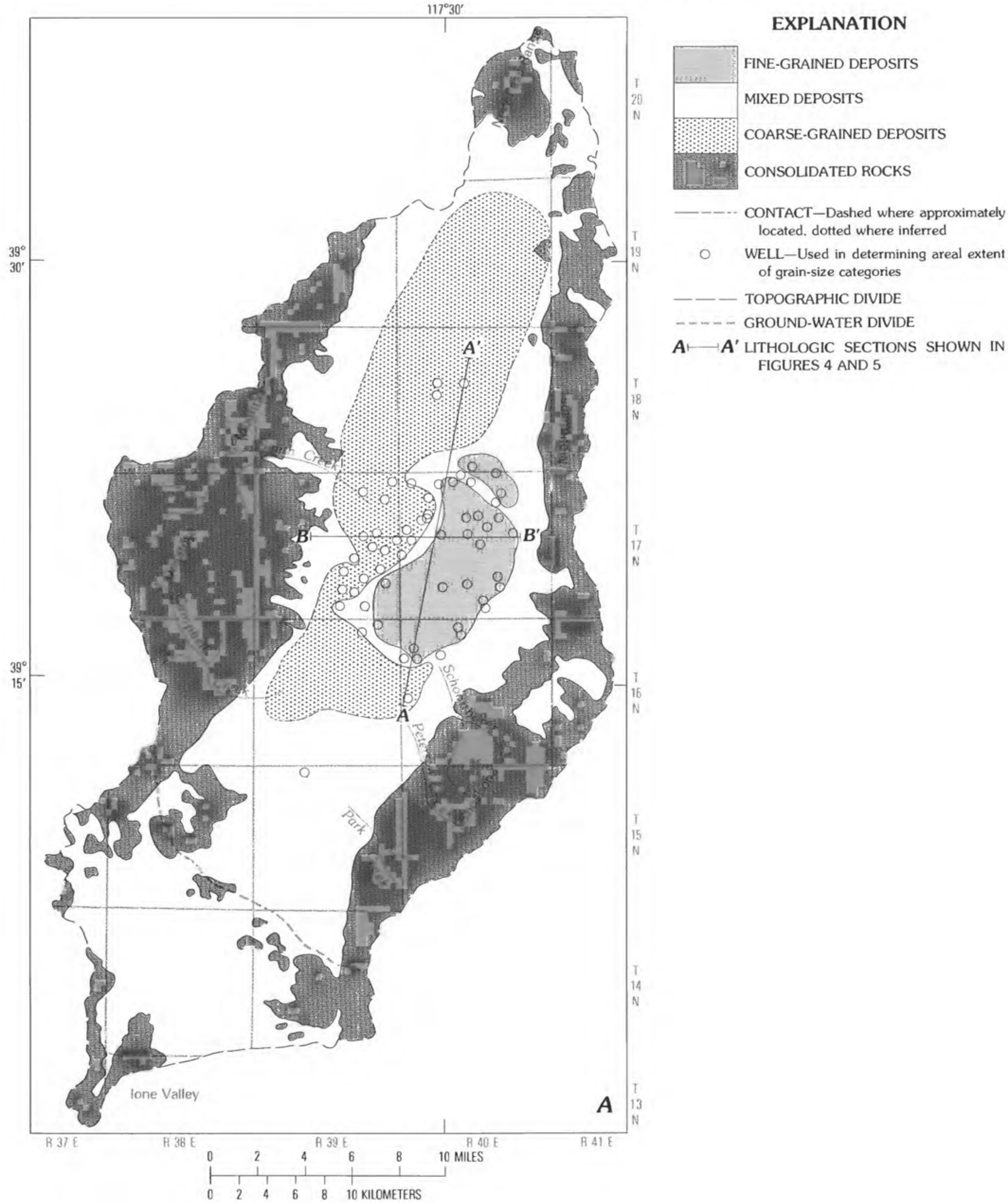
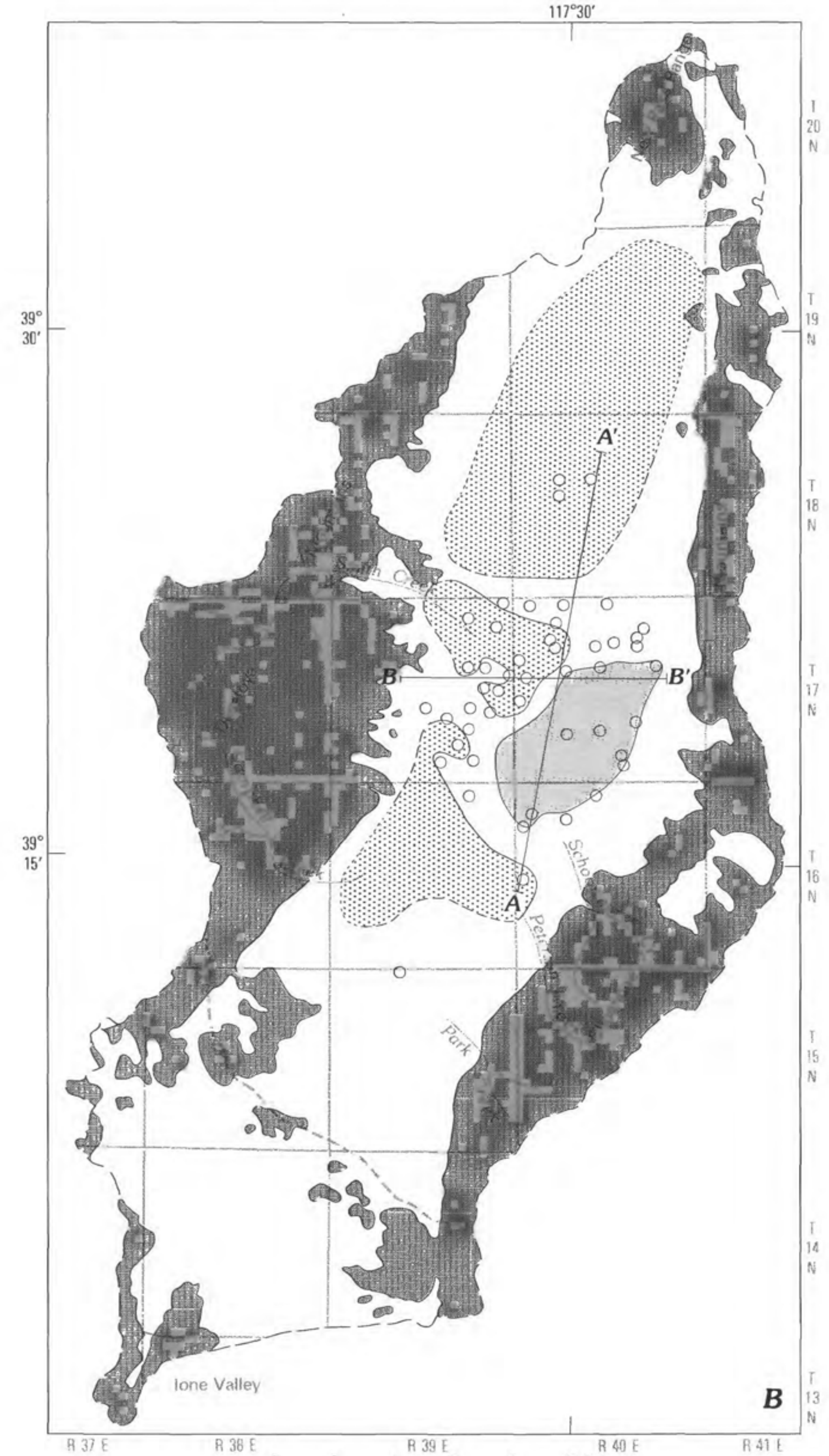


FIGURE 3 (above and facing page).—Fine-grained, coarse-grained, and mixed deposits in the (*A*) 0- to 50-ft and (*B*) 50- to 500-ft intervals of saturated basin fill below the 1982 water table. Deposits of the 50- to 500-ft interval are less than 450 ft thick around the margin of the basin.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0 2 4 6 8 10 KILOMETERS

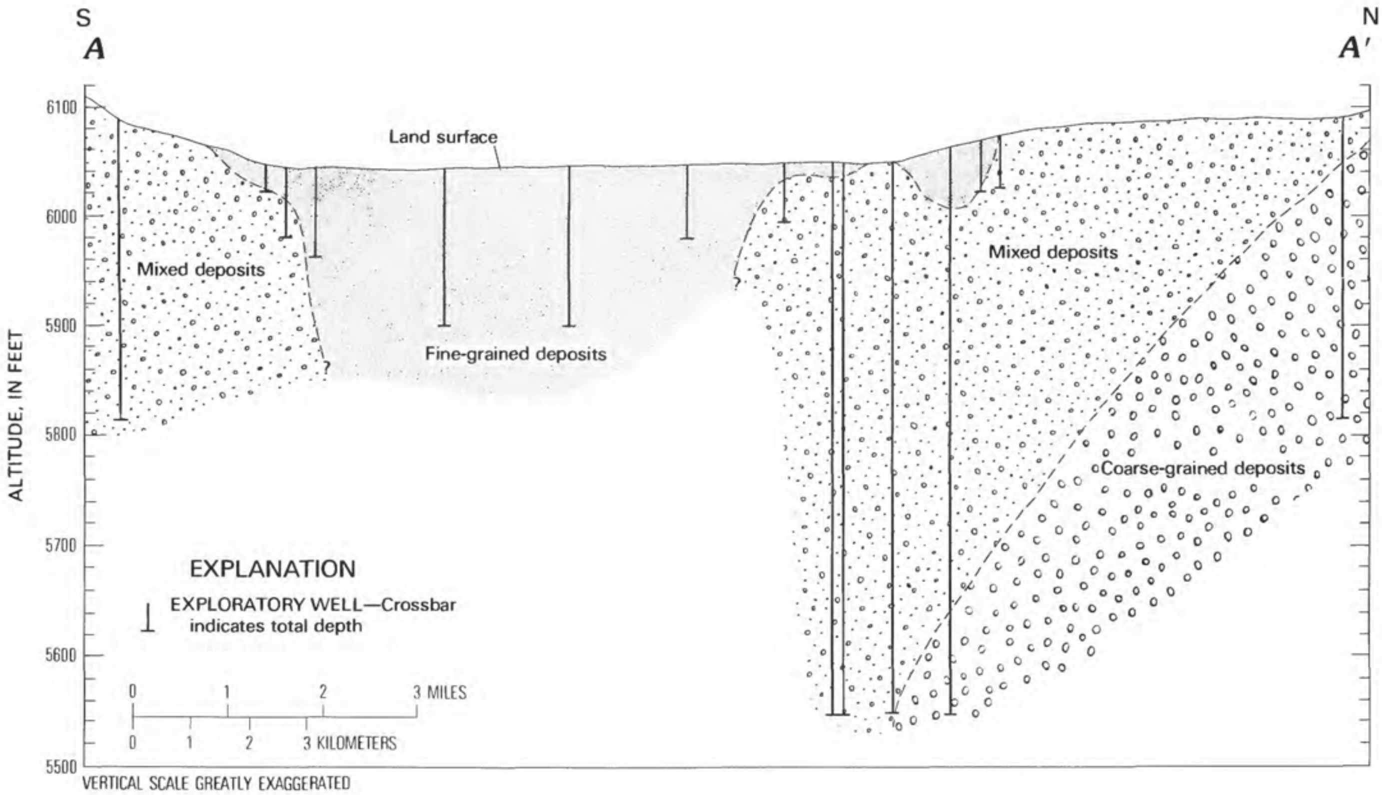


FIGURE 4.—Lithologic section *A-A′*. Line of section shown on figure 3.

in the surrounding mountains (tables 2 and 3, and fig. 6). The growing season in Smith Creek Valley is shorter than that in most other valleys in Nevada, due to high altitude; the season averages about 92 days, on the basis of the number of days between the last killing frost in the spring and the first killing frost in the fall (28°F).

Precipitation data for Smith Creek Valley include 17 yr of storage-gage records at Carrol Summit and only 1 yr of storage-gage records at seven other sites (table 3 and pl. 2). (Storage-gage containers are 3 ft high, 14 in. in diameter, and open at the top. They contain a small amount of a mixture of 60 percent methanol and 40 percent ethylene glycol, along with a layer of oil, to prevent freezing and evaporation.) Owing to the absence of long-term data for the valley, average precipitation has been approximated on the basis of data collected by the University of Nevada, Reno, Central-Nevada Field Laboratory in neighboring Reese River Valley and reported by the National Climatic Center. The laboratory is about 10 mi east of the Smith Creek Valley playa and approximately 5 mi east of a low topographic divide that forms the east boundary of the drainage basin. Climatological data for the Field Laboratory Station are available from 1965 to the present. Because the climate near Smith Creek Valley has been wetter during the last 17 yr than during the previous 76 yr, data from Austin, 25 mi east of the study area, were used to estimate long-term precipitation averages for the valley.

Precipitation generally increases with altitude in the study area (fig. 6). However, topographic effects may cause areal variations in precipitation throughout the study area. Thundershowers are the main source of precipitation in the summer, and precipitation falls mainly as snow in the winter.

Temperature also undergoes large daily and annual variations. Fifty degree diurnal temperature fluctuations are not uncommon in the study area.

SURFACE WATER

Most streams in Smith Creek Valley are ephemeral. However, Smith, Campbell, Peterson, Park, and Schoonnorer Creeks are perennial in the mountains and usually flow a short distance out from the mountain fronts onto alluvial fans (fig. 1 and pl. 2). For example, Campbell Creek flowed for a distance of nearly a mile from the mountain front onto the alluvial fan on April

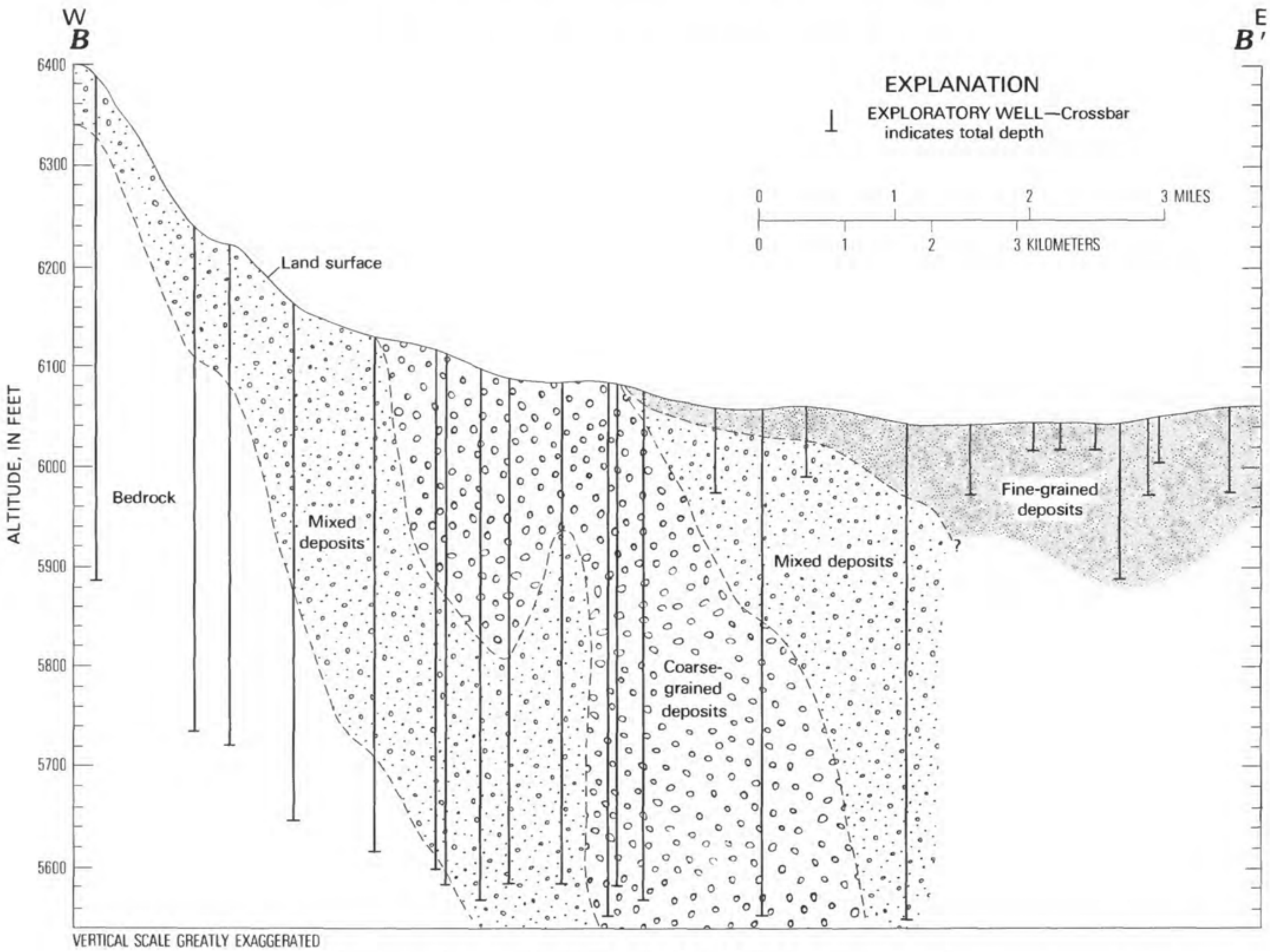


FIGURE 5.—Lithologic section *B-B'*. Line of section shown on figure 3.

10, 1981, and a few hundred feet on September 29, 1981. Surface water reaches the playa only during periods of high streamflow caused by intense rain storms or rapid melting of the snow pack in the surrounding mountains.

An estimated 8,800 acre-ft/yr (acre-feet per year) of surface water flows from the mountains and onto the valley floor (Nevada State Engineer, 1971). This estimate is based on runoff-altitude relationships developed for Nevada by Moore (1968). Although annual flows estimated by the runoff-altitude method are close to an estimate of annual flow based on a streamflow measurement for Smith Creek, these estimates are about two to four times higher than those based on flow measurements of the smaller Campbell and Peterson Creeks (table 4). Therefore, because the remaining drainages are smaller, lower in altitude, and are not perennial (with the exception of Park and Schoonnorer Creeks), runoff estimates by Moore for the remaining drainage areas of the basin probably are significantly greater than actual runoff. If runoff from the drainages not listed in table 4 is one-half to one-fourth of that estimated by the runoff-altitude method, actual annual runoff may be closer to 4,000 than the estimate of 8,800 acre-ft/yr by the Nevada State Engineer (1971).

Most of the runoff probably recharges the basin-fill ground-water aquifer. However, some water reaches the playa during periods of abundant streamflow and subsequently evaporates. Another component of runoff is evaporated and transpired along the stream channel without reaching either the ground-water aquifer or the playa.

GROUND WATER

All ground water in the Smith Creek drainage basin originates as precipitation. Precipitation that falls

TABLE 2.—*Precipitation data for the Central-Nevada Field Laboratory, Austin, and Carrol Summit*

[Precipitation, in inches; from published records of National Climatic Center, except as indicated]

Month	Central-Nevada Field Laboratory[1]	Austin[2]	Carrol Summit[3]
January----	0.64	0.89	1.07
February---	.45	1.01	.83
March------	.58	1.36	.82
April------	.59	1.47	1.39
May--------	.78	1.30	1.17
June-------	.89	1.33	1.41
July-------	.58	.51	.84
August-----	.82	.58	.91
September--	.60	.45	.83
October----	.54	.93	.81
November---	.44	.99	1.04
December---	.42	1.08	1.05
Annual-----	7.33	11.90	12.17

[1] Period of record: 17 years, 1966-82; all values contain estimated data or have missing data. Altitude 5,950 ft; lat. 39°23' N.; long. 117°19' W. (about 10 mi east of Smith Creek Valley playa).

[2] Data listed are for 30-year period, 1941-70. Annual average for 93-year period of record (1890-1982) is 12.17 inches. Altitude 6,605 ft; lat. 39°30' N.; long. 117°05' W. (about 25 mi northeast of Smith Creek Valley playa).

[3] U.S. Bureau of Land Management stations. Period of record: 17 years, 1964-80; all values contain estimated data. Altitude 7,700 ft; lat. 39°16' N.; long. 117°44' W. (about 15 mi southwest of Smith Creek Valley playa).

within the drainage basin infiltrates: (1) unconsolidated deposits as well as cracks and fractures in consolidated rocks in the mountains and (2) unconsolidated deposits in the valley floor. After infiltrating the water flows downgradient from areas of high head, usually in or near the mountains, to areas of low head in the vicinity of the playa near the center of the basin (pl. 2).

Ground water is unconfined in generally shallow, mixed- to coarse-grained deposits within the basin-fill, and becomes confined in areas containing fine-grained deposits. Deposits generally become finer toward the center of the basin. Alluvial fan deposits may also contain fine-grained materials that act as barriers to downward movement, producing zones of perched ground water locally.

Confined conditions in the playa area are caused by fine-grained sediments that are the result of (1) downstream sorting, whereby grain size becomes smaller due

TABLE 3.—*Annual precipitation at stations in and adjacent to Smith Creek Valley*

Number on plate 2	Station[1]	Altitude (feet)	Inches	
			1982[2]	Normalized[3]
1	East side of valley floor	6,050	8.25	5.79
2	West side of valley floor	6,180	9.71	6.80
3	East-side foothills------	6,210	9.37	6.56
4	Railroad Pass------------	6,470	7.93	5.55
5	West-side foothills-----	6,480	9.10	6.37
6	Carrol Summit-----------	7,440	19.10	13.37
7	Carrol Summit[4]----------	7,700	--	[5] 11.29
8	Basque Summit-----------	7,720	17.20	12.04
--	Central Nevada Field Laboratory[6]----------	5,950	9.40	[7] 6.36

[1] U.S. Geological Survey stations in Smith Creek Valley, except as indicated.

[2] Data for stations 1-6 and 8 collected by R. L. Carman, U.S. Geological Survey.

[3] Precipitation at stations 1-6 and 8 normalized to 93-year record at nearby Austin by using the following formula:

$$\text{(1982 precipitation at station)} \times \frac{\text{(93-yr avg. at Austin)}}{\text{(1982 total at Austin)}}$$

[4] U.S. Bureau of Land Management station (period of record, 1964-80).

[5] Average for period of record normalized to 93-year record at Austin by using the following formula:

$$\text{(1964-80 avg. at station)} \times \frac{\text{(93-yr avg. at Austin)}}{\text{(1964-80 avg. at Austin)}}$$

[6] National Climatic Center station in Reese River Valley (period of record, 1966-82).

[7] Average for period of record normalized to 93-year record at Austin by using the following formula:

$$\text{(1966-82 avg. at station)} \times \frac{\text{(93-yr avg. at Austin)}}{\text{(1966-82 avg. at Austin)}}$$

to decreased stream velocities as surface gradients flatten, and (2) deposition in Quaternary lakes.

Heads in the playa area increase with depth to at least 157 ft (the deepest well in the playa area), although the rate of head increase declines with depth. Figure 7 shows the upward vertical hydraulic gradient in the playa area between the depth intervals of 22 to 25 ft and 65 to 72 ft for pairs of adjacent observation wells completed in these depth intervals.

BASIN-FILL AQUIFER

AREAL EXTENT AND THICKNESS

The contact between consolidated rocks and basin fill around the periphery of the valley floor marks the outer

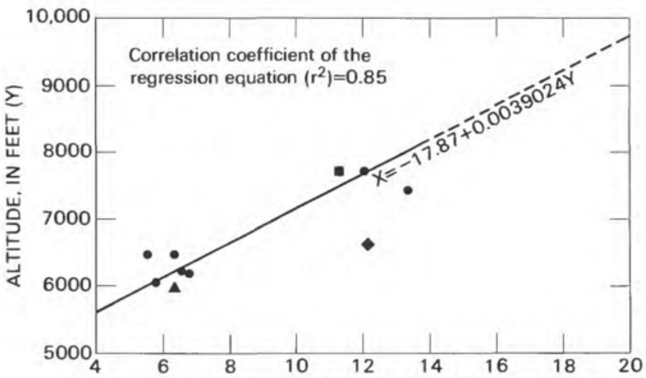


FIGURE 6.—Relationship between altitude and precipitation.

boundary of the basin-fill aquifer (pls. 1 and 2). Basin-fill deposits cover 340 mi² in the study area, which is approximately 60 percent of the total surface area of Smith Creek Valley.

The thickness of unconsolidated deposits in the basin (pl. 1) was determined by interpretation of the Bouguer gravity data of Erwin and Bittleston (1977). Basin-fill thicknesses determined by gravimetry were intermediate to values obtained from seismic and aeromagnetic data collected and interpreted by Herring (1967) and Salehi (1967) and similar to thicknesses determined in geothermal exploration wells drilled near the western margin of the basin by Hunt Oil Company of Nevada.

Gravity values indicate density variations in material beneath the land surface after corrections have been made to eliminate the effect of surrounding materials. Basin-fill deposits are generally less dense than adjacent and underlying consolidated rocks. This difference in densities produces gravity anomalies due to the presence of the low density basin-fill deposits. The magnitude of the gravity anomaly is related to the thickness of the low-density deposits overlying the high-density consolidated rocks (Telford and others, 1976).

Six gravity profile lines were constructed across the study area for determining basin-fill thickness (pl. 1). Residual gravity values were calculated along these profile lines by subtracting the regional gravity gradient from the Bouguer gravity values. These values were then contoured to produce a residual-gravity anomaly map that was used to estimate depths to the consolidated rock in the basin. The areal configuration of residual values is presumed to closely resemble the topography of the consolidated rock, underlying the basin-fill deposits.

The thickness of basin-fill deposits was then calculated from the residual gravity values along the six profile lines by using a two-dimensional profile model based on a technique described by Cordell and Henderson (1968). An assumed density difference of 0.50 g/cm³ (gram per cubic centimeter), (2.67 g/cm³ for consolidated rocks and 2.17 g/cm³ for basin-fill deposits) was used to calculate depth to consolidated rock for the

TABLE 4.—*Discharge of major ephemeral streams*[1]

Stream	Discharge measurements		Drainage-basin areas, by altitude zone (acres)				Estimated avg. annual discharge	
	Date	ft³/s	7,000 to 8,000 ft	8,000 to 9,000 ft	Above 9,000 ft	Total above 7,000 ft	On basis of streamflow measurements[2] (acre-ft/yr)	On basis of runoff-altitude relationships[3] (acre-ft/yr)
Smith Creek	8- 4-82	1.35	8,200	4,880	1,660	14,470	2,000	2,060
Peterson Creek	7-29-82	.304	1,210	1,660	760	3,630	330	740
Campbell Creek	4-10-81 9-29-81	.288 .049	10,650	2,400	200	13,250	230	950

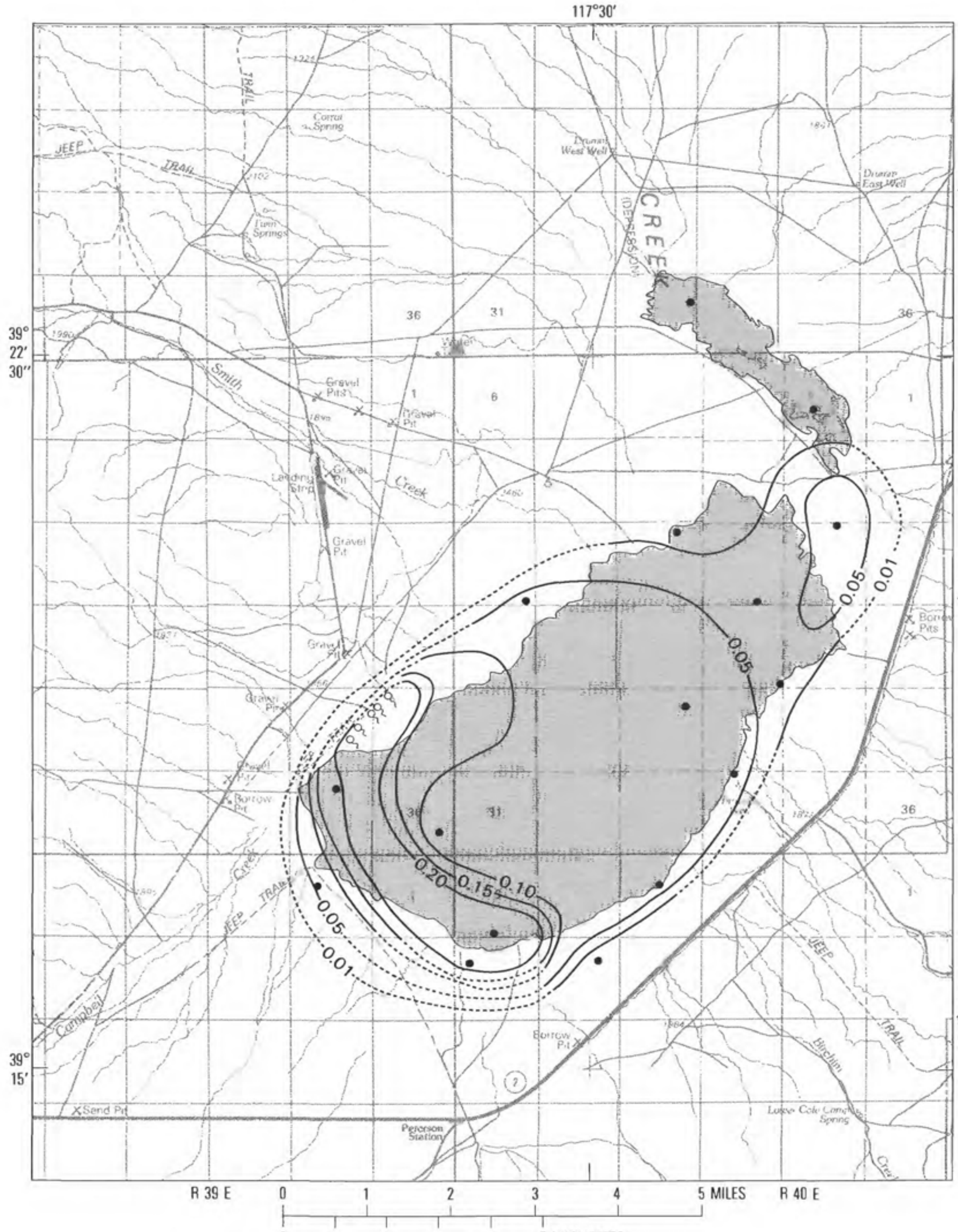
[1] Discharge at contact between consolidated rocks and basin fill at margin of basin.

[2] Estimated by comparing streamflow measurements with data for South Twin River, which is in nearby Toiyabe Range and drains Tertiary ash-flow tuffs similar to those surrounding most of Smith Creek Valley.

[3] Estimated from relationships developed by Moore (1968, page 33) by using the following equation:

$$Q = 0.4X_1 + 2.5X_2 + 5.6X_3 ,$$

where Q = average annual discharge, in acre-ft;
X_1, X_2, and X_3 = areas of altitude zones at 7,000–8,000, 8,000–9,000, and more than 9,000 ft;
and 0.4, 2.5, and 5.6 = regressional coefficients for the respective altitude zones (Moore, 1968, table 3).



EXPLANATION

BASIN FILL—Shaded area indicates playa deposits

—0.01--- LINE OF EQUAL VERTICAL HYDRAULIC GRADIENT—Approximately located, dotted where inferred. Interval, in feet per foot, is variable

• OBSERVATION-WELL PAIR—Used to determine hydraulic gradient

SPRING

FIGURE 7.—Upward vertical hydraulic gradient beneath the playa and adjacent areas. Steep gradient west of playa is along a fault scarp in the hot-spring area.

profile lines (Schaefer, 1983). The density value of 2.17 g/cm^3, which is used for basin-fill deposits elsewhere in the Great Basin, correlates well with borehole gravity data obtained from three wells in Dixie and Paradise Valleys, Nev. (Robbins and others, 1985).

Values for depth to consolidated rock were plotted on the profile lines and then contoured, using the residual gravity map as a guide, to produce a basin-fill thickness map for Smith Creek Valley (pl. 1). Maximum thickness of basin-fill deposits is about 5,500 ft, in a northeast-trending trough near the center of the basin beneath the playa. A shallower depression in the north-central part of the study area near the center of the basin contains basin-fill deposits of 3,000 ft. An east-trending consolidated-rock high with basin-fill thicknesses between 1,500 and 2,000 ft separates the two depressions. The lines of equal thickness on plate 1 suggest that the structural basin underlying the basin fill is nearly symmetrical. Depth to consolidated rock generally increases uniformly toward the deepest part of the basin near the center of the valley. Although the structural basin is bounded by faults along the mountain fronts and was produced by faulting, basin-fill thickness is contoured at too large an interval to detect individual faults that might exist in the bedrock beneath the basin-fill deposits. Basin-fill thicknesses in the southern part of the study area indicate that the basin-fill aquifer near the ground-water divide is relatively thin—less than 500 ft—compared to the rest of the basin.

Depth to consolidated rock in Smith Creek Valley was also estimated by using seismic and aeromagnetic geophysical techniques. Herring (1967) estimated a maximum depth of between 6,500 and 8,000 ft from three reversed seismic-refraction profiles made near the center of the basin (pl. 1). Salehi (1967) estimated a maximum depth of 3,700 ft and an average depth of 2,500 ft from an airborne magnetometer survey. The depths estimated in these studies bracket maximum depths estimated from the gravity data.

Geothermal temperature-gradient holes drilled to a depth of 500 ft near the western margin of the basin by Hunt Oil Company of Nevada penetrated consolidated rocks beneath basin-fill deposits near the mountain front. In this area, basin-fill thicknesses estimated from analysis of gravity data correspond reasonably well with consolidated-rock depths encountered in the wells. However, the configuration of the 500-ft contour has been slightly modified to account for the depths to consolidated rock determined by drilling (pl. 1).

HYDRAULIC PROPERTIES

HYDRAULIC CONDUCTIVITY

Basin-fill deposits in Smith Creek Valley generally have a greater horizontal than vertical hydraulic conductivity because they were deposited in relatively horizontal layers. Horizontal hydraulic conductivity of a basin-fill aquifer depends primarily on the high-permeability layers (Freeze and Cherry, 1979, p. 30–32). Low-permeability layers restrict flow, but, unless they compose a major part of the aquifer, they have little effect on the horizontal conductivity of the aquifer.

In contrast, the vertical hydraulic conductivity of a basin-fill aquifer depends primarily on the conductivities of the low-permeability layers because the vertical flow of water is approximately perpendicular to the layers. Furthermore, horizontal to vertical anisotropy due to preferential horizontal orientation of fine-grained particles can be three to one, or more (Freeze and Cherry, 1979, p. 32). Thus, within a basin-fill aquifer that contains layers of coarse- and fine-grained deposits, the horizontal hydraulic conductivity of the aquifer approximates the hydraulic conductivity of the coarse-grained deposits, whereas the vertical hydraulic conductivity approaches that of the fine-grained deposits.

No aquifer test data are available for Smith Creek Valley, so hydraulic-conductivity estimates for the basin-fill deposits are based on: (1) grain-size distributions within the basin fill, (2) aquifer test data from other valleys in the Great Basin, (3) a laboratory determination of vertical hydraulic conductivity, (4) calculations from moisture-retention, grain-size, and porosity analyses, and (5) water-temperature profiles.

Horizontal hydraulic conductivities estimated from specific capacities of 84 wells completed in coarse-grained basin-fill deposits (primarily sands and gravels with thin beds of silt and clay) in Paradise Valley, Nev., range from 9×10^{-5} to 2×10^{-3} ft/s (foot per second) and average 4×10^{-4} ft/s (David E. Prudic, U.S. Geological Survey, written commun., 1983). In addition, horizontal hydraulic conductivities determined from 28 aquifer tests of basin-fill deposits for 15 valleys in the Great Basin conducted and analyzed by Ertec Western, Inc. (written commun., 1982) range from 2×10^{-5} to 6×10^{-3} ft/s. Hydraulic conductivities determined from these aquifer test data generally fall within the range of values expected for coarse-grained deposits (Morris and Johnson, 1967, table 12). Therefore, an average value of 4×10^{-4} ft/s is thought to be representative of the horizontal conductivity of coarse-grained deposits in Smith Creek Valley.

Vertical hydraulic conductivities for coarse-grained deposits determined from 50 core samples in central California reported by Johnson and others (1968) range from 9.1×10^{-5} to 1.5×10^{-4} ft/s. The arithmetic mean is 1.2×10^{-4} ft/s. Therefore, a vertical hydraulic conductivity of 1×10^{-4} ft/s may be a reasonable estimate for coarse-grained deposits in Smith Creek Valley.

Horizontal and vertical hydraulic conductivities of the fine-grained deposits (primarily silts and clays with thin beds of gravel and sand) in the valley were determined from core samples and water-temperature gradients collected from well 17N 40E 29ACA near the central part of the playa (pl. 2). Four techniques were used to estimate the hydraulic conductivity of the playa deposits (table 5): (1) a laboratory test of a core sample using a falling-head permeameter; (2) the relationship between a soil-moisture retention curve and hydraulic conductivity (Marshall, 1958); (3) an estimate derived from the Kozeny-Carmen equation (Bear, 1972), which relates porosity and particle-size distribution to hydraulic conductivity, and (4) water-temperature profiles in the well (Michael L. Sorey, U.S. Geological Survey, written commun., 1971). Estimates of hydraulic conductivity range from 2×10^{-7} to 1×10^{-8} ft/s (table 5) with an average value of 7×10^{-8} ft/s. These values assume that the core samples are nearly isotropic, which may be reasonable because vertical hydraulic conductivities determined by a laboratory analysis and water-temperature profiles (table 5) are within the same range as hydraulic conductivities estimated from porosity, particle-size distribution, and moisture-retention analyses. Johnson and others (1968) reported an average hydraulic conductivity of 1.3×10^{-7} ft/s for 219 core samples of fine-grained deposits. Therefore, a hydraulic conductivity of 1×10^{-7} ft/s may be a reasonable estimate of the horizontal and vertical conductivity of fine-grained deposits in the valley.

Mixed deposits in the valley have hydraulic conductivities between those of the coarse-grained deposits (4×10^{-4} ft/s) and fine-grained deposits (1×10^{-7} ft/s). Conductivities for mixed deposits are probably lower in areas adjacent to the playa and higher away from the playa and along the margin of the valley floor because the deposits become coarser grained away from the playa. Representative values of hydraulic conductivity may be 1×10^{-5} ft/s for the marginal alluvial fan and coarser playa-area deposits and 1×10^{-6} ft/s for deposits adjacent to the playa, except directly south of the playa where mixed deposits contain more coarse material.

Hydraulic conductivities for saturated deposits in the depth intervals from 0 to 50 ft and from 50 to 500 ft can be estimated by using the grain-size distribution maps constructed for these intervals (fig. 3) and the average hydraulic-conductivity values for fine-grained, coarse-grained, and mixed deposits. No data exist for deposits in the basin-fill aquifer below a depth of 500 feet, but estimates of approximate conductivity were made for this interval by using the basin-fill thickness map (pl. 1) and assuming that the deposits below 500 feet are similar to those in the 50- to 500-ft interval, and that the hydraulic conductivity of these deposits decreases linearly with depth, at a rate of 25 percent for every 500-ft interval (Durbin and others, 1978).

TABLE 5.—*Physical and hydraulic properties of a core sample from well N17 E40 29ACA*

[Samples collected by F. E. Rush and J. R. Harrill, U.S. Geological Survey, in 1965; analyses by Geological Survey laboratories in Denver, Colo., and Menlo Park, Calif.]

Property	Value
Sample depth (feet below land surface)	27.5
Particle-size distribution (percent):	
Clay (<0.004 mm)	24
Silt (0.004-0.0625 mm)	53
Sand and coarser (>0.0625 mm)	23
Total porosity (percent)	50.6
Specific yield (percent)	6.5
Hydraulic conductivity (feet per second):	
Laboratory determined (vertical)	3×10^{-8}
Estimated from moisture-retention curves[1]	2×10^{-7}
Estimated from porosity and grain-size analyses[2]	5×10^{-8}
Estimated from water-temperature profiles (vertical)[3]	1×10^{-8}
Average hydraulic conductivity	7×10^{-8}

[1] Calculated by using equation of Marshall (1958, p. 4); calculation does not take sample orientation into account.

[2] Calculated by using Kozeny-Carmen equation (Bear, 1972, p. 166); calculation does not take sample orientation into account.

[3] Vertical hydraulic conductivity is estimated over entire interval of well (Michael L. Sorey, U.S. Geological Survey, written commun., 1971).

SPECIFIC YIELD

Approximate specific yields for intervals in the basin-fill aquifer can be estimated by using the lithology maps (fig. 3) and reported specific yields for coarse-grained, mixed, and fine-grained deposits. Specific yields of coarse-grained deposits can be as high as 27 percent for clean coarse sand (Johnson, 1967, table 29), but coarse-grained deposits in Smith Creek Valley contain thin beds of silt and clay, and in some places fine-grained material partly fills pore spaces within the larger grain-size matrix, thereby reducing specific yields. A specific yield of approximately 15 percent used in ground-water flow models for other valleys in the Great Basin (Harrill, 1982; Mower, 1982; and David S. Morgan, U.S. Geological Survey, written commun., 1983), was assumed to be the average specific yield of the coarse-grained deposits in the basin. A specific yield for mixed deposits in the valley may be approximately 10 percent (Harrill, 1982; and David S. Morgan, written commun., 1983). Fine-grained deposits in the playa area may have a specific yield of about 6 percent (table 5), (Johnson, 1967; Harrill, 1982; Mower, 1982, p. 22; and David S. Morgan, written commun., 1983). The estimated specific yields for the intervals in the basin-fill aquifer are assumed to have areal boundaries coinciding with those of grain-size distributions in figure 3.

STORAGE COEFFICIENT

Storage coefficients for the 50-ft interval below the water table are assumed to be the same as the specific yield for the same interval of the aquifer. Thus, for the 50-ft interval the storage coefficients were assumed to range from approximately 6 percent for fine-grained deposits in the playa area to 15 percent for coarse-grained deposits in the aquifer.

Storage coefficients for the confined intervals of 50 to 500 ft and 500 ft to consolidated rock of saturated basin fill were determined by multiplying the thickness of the interval by the specific storage, which is assumed to be 9.3×10^{-7} per ft of aquifer for coarse-grained and mixed deposits and 4.7×10^{-6} per ft for fine-grained deposits (Ireland and others, 1982, p. 28–35). Elastic compaction of fine-grained deposits is greater than for coarser deposits because of differences in pore structure; as a result, specific storage also is greater for the fine-grained deposits. Storage coefficients range from 2.1×10^{-3} to 4.2×10^{-4} for the 50- to 500-ft interval and from 7.0×10^{-5} for an interval 75 ft thick (that is, 500 to 575 ft below the water table) around the margin of the valley floor to 4.6×10^{-3} for an interval 5,000 ft thick near the center of the basin for deposits below 500 feet. The 50- to 500-ft interval is assumed to be under confined conditions; however, if the water table were to decline more than 50 ft below its present level, the deeper interval would become a water-table aquifer and the storage coefficient would equal the specific yield of the sediments in the interval.

GROUND WATER IN STORAGE

The amount of water stored in the basin-fill aquifer was estimated as a product of the area, thickness, and average specific yield for the fine-grained, coarse-grained, and mixed deposits. Storage estimates assume that all basin-fill deposits would be dewatered as water was withdrawn from the aquifer. Approximately 2.1 million acre-ft of water is stored in the uppermost 100 ft of saturated basin fill. This quantity is considered more probable than the earlier estimate of 1.5 million acre-ft (Nevada State Engineer, 1971), which did not take into account areal or vertical distribution of grain size. About 6.6 million acre-ft of water is stored in the uppermost 500 ft, and the entire basin-fill aquifer contains a total of about 21 million acre-ft of water. However, ground water that is suitable in quality and within economic reach for agricultural use (the uppermost 200 ft of saturated deposits) may amount to only about 3.3 million acre-ft, or roughly 15 percent of the aquifer total.

GROUND-WATER RECHARGE

Recharge to the basin-fill aquifer was estimated using the Maxey-Eakin method (Maxey and Eakin, 1949) and a chloride-balance technique. Both methods are based on total precipitation that falls within the recharge areas of the valley. According to Everett and Rush (1964), recharge in Smith Creek Valley originates in the mountains surrounding the valley floor above an altitude of 7,000 ft. Total precipitation above 7,000 ft was calculated by determining the areas within altitude zones of 7,000 to 8,000, 8,000 to 9,000, and 9,000 to 10,313 ft above sea level, obtained from 7½- and 15-minute topographic maps, and then multiplying the area of each zone by the average precipitation for the zone as determined from the relation between altitude and precipitation (fig. 6). Estimated precipitation, by altitude zone, is given in table 6.

The Maxey-Eakin method estimates recharge as a percentage of total precipitation for each altitude zone (table 6). The zone-by-zone percentage values are based on estimates developed by Maxey and Eakin (1949) by trial-and-error balancing of recharge and discharge estimates for 13 valleys in east-central Nevada. An estimated average annual recharge rate of approximately 9,600 acre-ft/yr was calculated for Smith Creek Valley using this method. This is less than the estimate of

TABLE 6.—*Estimated average annual ground-water recharge*

Altitude zone (ft)	Area[1] (acres)	Precipitation		Estimated percentage of precipitation that becomes recharge[3]	Recharge (acre-ft per yr)
		Feet per yr[2]	Acre-ft per yr		
7,000–8,000	64,800	0.95	62,000	7	4,300
8,000–9,000	18,000	1.28	23,000	15	3,500
9,000–10,313	4,200	1.65	7,000	25	1,800
Total	87,000	---	92,000	---	9,600

[1] Acreages differ from those listed by Everett and Rush (1964, p. 10) because (1) areas in the extreme southern end of the valley that generate recharge which flows south into Ione Valley are not included, and (2) areas listed herein are based on larger scale maps (1:24,000 and 1:62,500) than those used by Everett and Rush (1964, p. 10; 1:250,000).

[2] Precipitation is for average altitude of zone, and was calculated by using equation in figure 6. [Values used by Everett and Rush (1964, p. 10) were 1.12, 1.46, and 1.75 ft, respectively.]

[3] Percentages from Everett and Rush (1964, p. 10).

12,000 acre-ft/yr made by Everett and Rush (1964, p. 10). About 60 percent of the difference is due to the new altitude-precipitation relationship used in this study and about 40 percent is due to the smaller recharge area used herein because recharge in the extreme southern end of the valley flows into Ione Valley.

Using a chloride-balance technique, average annual recharge was estimated to be about 8,300 acre-ft/yr. The technique estimates recharge by comparing the total chloride input from precipitation and dry fallout in recharge areas to chloride concentrations in the ground water recharging the basin-fill aquifer (Vacher and Ayers, 1980; Ayers, 1981; Mandel and Shiftan, 1981; Irving, 1982; and Michael D. Dettinger, U.S. Geological Survey, written commun., 1983). Recharge estimates were calculated using the following equation:

$$R = P(Cl_P/Cl_R),$$

where R = recharge, in acre-feet per year;
P = total precipitation in the overall recharge area, in acre-feet per year;
Cl_P = chloride concentration of precipitation and dry fallout, in milligrams per liter; and
Cl_R = chloride concentration of ground water in the recharge area, in milligrams per liter.

The estimated volume of total precipitation for the recharge area is 92,000 acre-ft (table 6). Chloride concentration of precipitation and dry fallout in the recharge area is estimated by Michael D. Dettinger (U.S. Geological Survey, oral commun., 1983) to be 0.4 mg/L (milligrams per liter). The chloride concentration of ground water in the recharge area, an estimated 4.4 mg/L, is derived from the average chloride concentrations of Smith, Campbell, and Peterson Creeks sampled during low flows and weighted for the estimated average annual discharge of each stream at the mountain-block and basin-fill contact.

The areal distribution of recharge from mountainous areas surrounding the valley floor is the same for both methods of estimation, because each is based on total precipitation. Only the relative amount of recharge from each area differs depending on the method used. Approximately 55 percent of the recharge to the basin originates in the Desatoya Mountains along the western side of the valley (fig. 8); the Smith and Campbell Creek drainage basins supply most of this recharge. The other major source of recharge is the southern half of the Shoshone Mountains, which supplies about 40 percent of the total. The Schoonnorer, Peterson, and Park Creek drainage basins provide most of this recharge. The northern half of the Shoshone Mountains and the New Pass Range supply the remaining recharge.

GROUND-WATER DISCHARGE

Ground water in Smith Creek Valley is discharged from the basin-fill aquifer by evapotranspiration from phreatophytic areas around the margin of the playa, evaporation from bare playa soil, and discharge from wells that ultimately is lost by evapotranspiration from irrigated alfalfa fields.

EVAPOTRANSPIRATION BY PHREATOPHYTES

Plants in the study area may be classified as members of one of three groups: (1) xerophytes, (2) phreatophytes, and (3) crops. A xerophyte derives water primarily from surface-water infiltration, whereas a phreatophyte derives water primarily from the aquifer (Meinzer, 1927, p. 1). In some places, plants that are normally phreatophytic grow in xerophytic areas where surface runoff pools and is temporarily stored as perched ground water. So a rigorous definition of phreatophytic evapotranspiration is required to avoid erroneous incorporation of xerophytic areas in the discharge zone. In this study, a phreatophyte is considered to be a plant whose main source of water is from the basin-fill aquifer.

Crops in Smith Creek Valley rely principally on ground-water irrigation. Alfalfa is the only commercial crop grown in the valley. Although alfalfa roots commonly penetrate to depths of 33 ft or more (Zimmerman, 1969, p. 40), there are no fields in Smith Creek Valley where alfalfa roots reach the water table. Therefore, discharge of ground-water by crops is through irrigation and will be discussed later.

The presence of phreatophytes is typically governed by their ability to obtain water, a direct function of their root depth. However, phreatophytes may grow in areas where the surface-water supply is sufficient to cause seasonal, albeit temporary, fluctuation of the water table. Phreatophytic areas of this nature are prevalent in the west-central and southwest parts of the valley where Smith and Campbell Creeks enter the basin. In contrast to these areas having surface-water augmentation, the east side of the valley receives little surface runoff, and the phreatophyte-xerophyte boundary is better defined.

Phreatophytes in Smith Creek Valley are of three principal types: (1) big greasewood (*Sarcobatus vermiculatus*), (2) rabbitbrush (*Chrysothamnus*), and (3) saltgrass (*Distichlis stricta*) (fig. 9). Great Basin wildrye (*Elymus cinereous*), another phreatophytic grass, is also present as a successor plant in two abandoned fields, but appears to be yielding to rabbitbrush incursion. Because grasses constitute only a minor part of the total discharge area in Smith Creek Valley and because

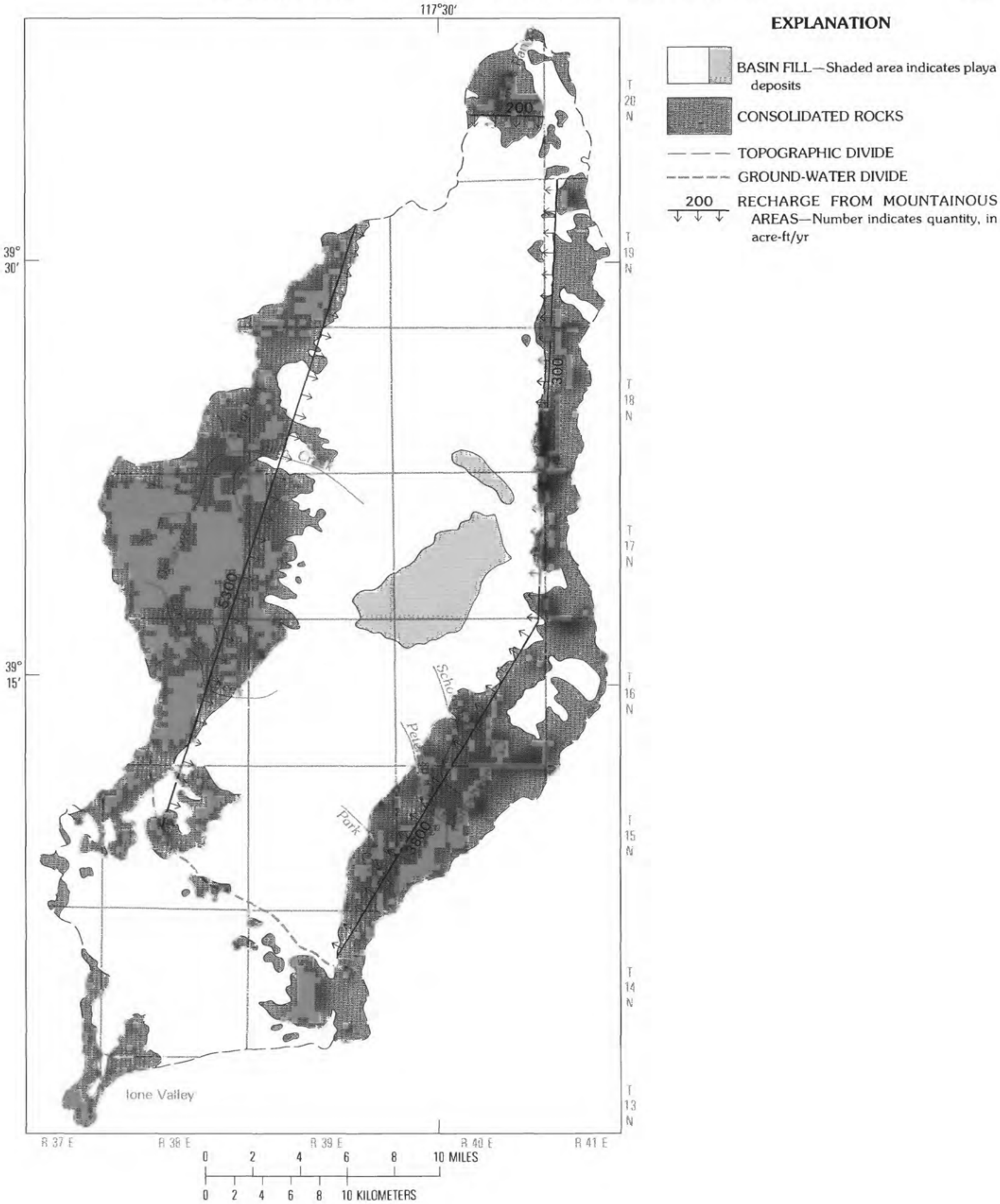


FIGURE 8.—Estimated areal distribution of recharge on the basis of the Maxey-Eakin technique (basin-wide total, 9,600 acre-ft/yr).

evapotranspirative rate information for wild phreatophytic grasses is minimal, wildrye is included with saltgrass to form a common plant-type zone.

Of the three principal phreatophyte types, greasewood has roots that penetrate deepest. In Smith Creek Valley, greasewood commonly thrives where the water table is 15 to 20 ft below land surface and becomes xerophytic where the depth to water is greater than 30 ft. A short distance from the eastern and northern margins of the main playa, the outer phreatophytic boundary coincides with a depth to water of 25 to 30 ft. Along this boundary, greasewood leaves become darker, greener, and twice as long in the phreatophytic area as in the adjacent xerophytic area. Greasewood also surrounds the smaller northern playa, but these plants may be sustained more by water ponded during sporadic storm events than by ground water. Consequently, plants in this area were considered to be more xerophytic than phreatophytic. Greasewood is the dominant plant along the east and north sides of the discharge area because the depth to water in this area is generally greater than 12 ft, which seems to be the approximate limit of root penetration of phreatophytic grasses and rabbitbrush in Smith Creek Valley.

In contrast to the east and north sides of the valley, the south and west sides are heavily wetted during the spring and early summer by Smith Creek, Campbell Creek, and numerous other, smaller streams. Because of this wetting, much of the area is dominated by dense stands of rabbitbrush which often exceed 3 ft in height.

Phreatophytic grasses grow where the water table is shallow, typically less than 7 ft, in abandoned hay fields, and in areas too saline for woody phreatophytes. Great Basin wildrye grows in two abandoned hay fields on the west side of the valley at the terminus of Campbell and Smith Creeks. Saltgrass grows in the hot springs area at the southwest edge of the playa.

The phreatophyte area was divided into zones having similar composition and foliage-volume density (table 7), by using methods described by Lawrence B. Hines (U.S. Geological Survey, written commun., 1983). The resulting distribution of phreatophytes is shown in figure 9.

After establishing vegetation-zone boundaries and assigning foliage volumes for each plant species, evapotranspiration rates were estimated for each zone occupied by either greasewood or rabbitbrush on the basis of foliage-volume density. Lysimeter studies by

TABLE 7.—*Foliage densities determined from vegetation transects*

Zone[1]	Phreatophyte type	Transect site[1]	Foliage density (cubic ft per square ft)					Percentage of total phreatophyte density	
			Greasewood	Rabbitbrush	Combined phreatophytes	Xerophytes	Total foliage	Greasewood	Rabbitbrush
1	Sparse greasewood	16	0.15	0.00	0.15	0.11	0.26	100	0
2	Moderate greasewood (includes some sagebrush)	1	(2)	(2)	.37	.39	.76	(2)	(2)
		7	.13	.21	.34	.20	.54	38	62
		4	.51	.10	.61	.03	.64	84	16
		6	.38	.06	.44	.00	.44	86	14
		9	.40	.00	.40	.01	.41	100	0
		10	.53	.05	.58	.18	.76	91	9
		12	.29	.15	.44	.17	.61	66	34
		13	.43	.00	.43	.04	.47	100	0
		15	.38	.17	.55	.13	.68	69	31
		17	.22	.21	.43	.05	.48	52	48
3	Dense greasewood and rabbitbrush	11	.31	.44	.75	.02	.77	41	59
4	Rabbitbrush	8	.00	.24	.24	.13	.37	0	100
		14	.13	.26	.39	.00	.39	33	67

[1] See figure 9.

[2] Non-foliated; transect performed before advent of growing season, so greasewood and rabbitbrush could not be distinguished from each other.

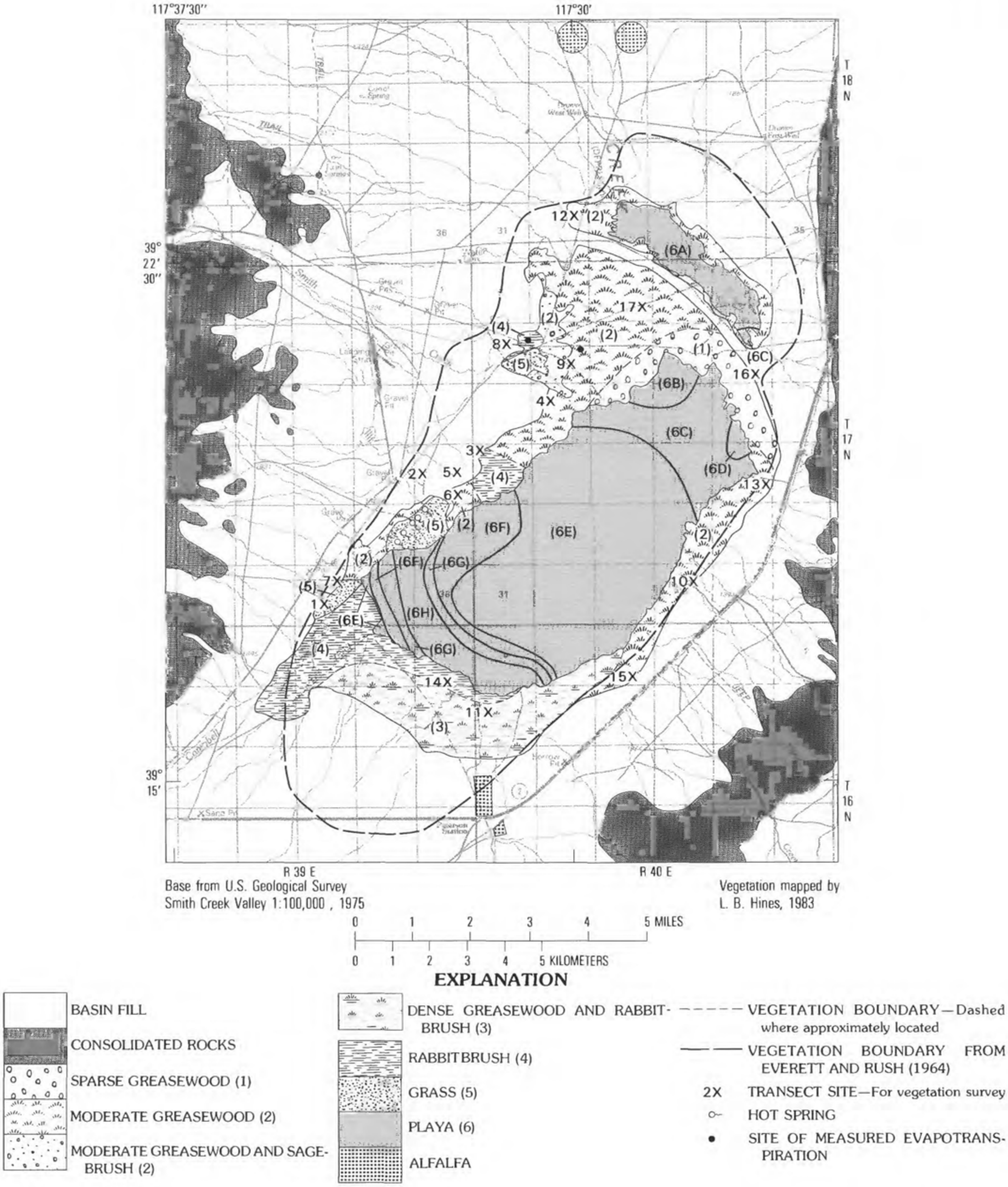


FIGURE 9.—Phreatophyte assemblages and areas of playa-surface evaporation. Parenthetical numbers and letters refer to evapotranspiration-rate zones listed in tables 7 and 8.

Robinson and Waananen (1970, p. 28) for the years 1963 through 1967 indicate average annual evapotranspiration rates of 0.7 and 1.1 ft^3 of water per ft^3 of foliage for greasewood and rabbitbrush, respectively. These rates were factored to accommodate areas of mixed vegetation by multiplying each rate by the fractional amount of foliage volume present (that is, the volume density of a species divided by the total volume density of all phreatophytes in the area under consideration). The products were then summed to obtain an overall evapotranspiration rate for the area. Annual evapotranspiration from an area was determined by multiplying the overall evapotranspiration rate by the respective acreage (table 8).

EVAPORATION FROM THE PLAYA

Evaporation rates for the playa were estimated as specific discharge. The playa was divided into 12 areas in which each is assumed to have uniform vertical hydraulic head gradient (figs. 7 and 9). Boundaries for these areas consist of lines of equal vertical head gradient at intervals of 0.05 ft/ft. Upward flow in each of these areas was estimated by multiplying the vertical hydraulic-conductivity value of 7×10^{-8} ft/s (determined for a sediment sample and water-temperature profiles for a well in the playa as discussed previously in the "Basin-Fill Aquifer" section) by the zonal area and the average vertical hydraulic gradient. Playa evaporation rates are summarized in table 8. Total evaporation from the playa was estimated to be approximately 2,400 acre-ft/yr. However, ground-water discharge from the playa may be less in some years because a significant part of the playa is commonly covered with overland runoff for 1 to 3 months during the spring and early summer. Previous work suggests that vertical head gradients are not affected by the standing water over the playa, perhaps because adjustment of heads proceeds more slowly than removal of standing water by evaporation. More research seems warranted with regard to playa evaporation, and results presented in this report should be considered as a first approximation.

TABLE 8.—*Natural evapotranspiration*

Zone	Zone type	Phreatophyte density[1] (Cubic ft per square ft)		Avg. annual evapotranspiration[2] (ft)	Area (acres)	Avg. annual discharge from aquifer (acre-ft)
		Greasewood	Rabbitbrush			
1	Sparse greasewood-----	0.15	0.00	0.11	1,170	130
2	Moderate greasewood---	.36	.11	.37	6,370	2,360
3	Dense greasewood and rabbitbrush---------	.75	.31	.87	2,460	2,140
4	Rabbitbrush-----------	.07	.25	.32	2,150	690
5	Saltgrass and wildrye[3]	---	---	.5	670	340
6A	Playa-----------------	---	---	.01	1,040	10
6B	Playa-----------------	---	---	.02	380	8
6C	Playa-----------------	---	---	.05	1,860	90
6D	Playa-----------------	---	---	.11	120	10
6E	Playa-----------------	---	---	.17	6,790	1,150
6F	Playa-----------------	---	---	.28	1,240	350
6G	Playa-----------------	---	---	.39	560	220
6H	Playa-----------------	---	---	.50	1,130	570
	Total (rounded)----------	---	---	---	26,000	8,100

[1] Statistical mean.

[2] Playa-surface evaporation rates were calculated as specific discharges (Darcian velocities) by using the equation: $q = (K_v)(I)$, where q is specific discharge (LT^{-1}; Darcian velocity); K_v is vertical hydraulic conductivity (LT^{-1}); and I is vertical hydraulic-head gradient (dimensionless).

[3] The average evapotranspiration rate used by Everett and Rush (1964, p. 11).

GROUND-WATER PUMPAGE

Ground water is pumped for domestic use, livestock watering, and irrigation of crops. In 1983, the combined volume of water for domestic and livestock use is estimated to have been less than 20 acre-ft and has, therefore, been omitted from the water-budget estimates. Only four fields were being irrigated during the 1983 growing season. Total net pumpage (that part of the pumpage that is consumed by crops) was estimated by multiplying the irrigated acreages by annual crop-consumption rates of 2.0 ft/yr for the two northern fields (250 acres) and 1.0 ft/yr for the two southern fields (150 acres). (Both rates are based on three cuttings per year and the lower rate reflects the use of surface water as well as ground water; see Rush, 1976, p. 7.) Thus, total net pumpage was an estimated 650 acre-ft/yr for irrigation in 1983.

SUMMARY OF DISCHARGE ESTIMATES

Estimates of ground-water discharge are summarized in table 9. Total natural discharge from the aquifer is about 8,300 acre-ft/yr, which compares reasonably well with the 6,900 acre-ft/yr estimated by Everett and Rush (1964, p. 12). An additional 650 acre-ft was consumed by irrigated crops in 1983. Pumpage for stock and domestic uses is estimated to have been less than 20 acre-ft in 1983.

TABLE 9.—*Summary of discharge from the basin-fill aquifer*

Component	Avg. annual discharge (acre-ft)
Natural evapotranspiration (table 8)	8,100
Hot-spring discharge[1]	200
Total natural discharge	8,300
Evapotranspiration induced by ground-water pumping	650
Domestic and livestock use of ground water	<20
Total discharge (rounded)	9,000

[1] From Everett and Rush (1964, page 12).

SIMULATION OF GROUND-WATER FLOW

CONCEPTUAL MODEL

Development of a conceptual model that can integrate the geology and hydrology of the Smith Creek Valley flow system is essential in evaluating the ground-water flow system. Simulation with a computer-based (mathematical) model, using the physical properties estimated from the conceptual model, leads to further refinements of the understanding of the flow system. An established and calibrated numerical flow model can be used to evaluate the potential impacts of future ground-water development.

The ground-water flow system in Smith Creek Valley described in the preceding sections is a hydrologically closed aquifer containing unconsolidated basin-fill deposits. The surrounding low-permeability consolidated rocks act as barriers to the ground-water flow and delineate the east, west, north, and bottom boundaries of the flow system. The southern boundary is a ground-water divide in the vicinity of volcanic rocks that protrude through the unconsolidated basin fill (pl. 2). The top of the flow system is the water table.

In general, ground water flows from the edges of the basin-fill aquifer toward the center, as illustrated in figure 10 (a simplified conceptualization of ground-water flow in Smith Creek Valley). Water recharges the basin-fill aquifer around the periphery and flows laterally toward the central playa area, where it moves primarily upward and is discharged by evaporation from the soil and transpiration by phreatophytes. Ground water exists under both confined and unconfined conditions. For purposes of simulation, ground water in the uppermost 50 ft of saturated basin fill was considered to be unconfined. Below the 50-ft depth, ground water was considered to be confined.

Hydraulic conductivity and, therefore, transmissivity are influenced by depositional processes. Basin-fill deposits generally show a decrease in grain size and better sorting toward the center of the basin. This depositional environment produces poorly sorted gravel and boulders in a fine-silt matrix beneath the upper parts of fans, grades into better sorted and finer grained deposits away from the mountains, and finally grades into very well sorted clayey silt deposits beneath the playa.

MATHEMATICAL FLOW MODEL

A mathematical flow model that describes the system is of great use in defining basin-wide hydrologic properties and evaluating potential impacts caused by future ground-water withdrawals. The mathematical model used to analyze ground-water flow in Smith Creek Valley is a three-dimensional flow equation which may be written as follows (McDonald and Harbaugh, 1984, p. 7):

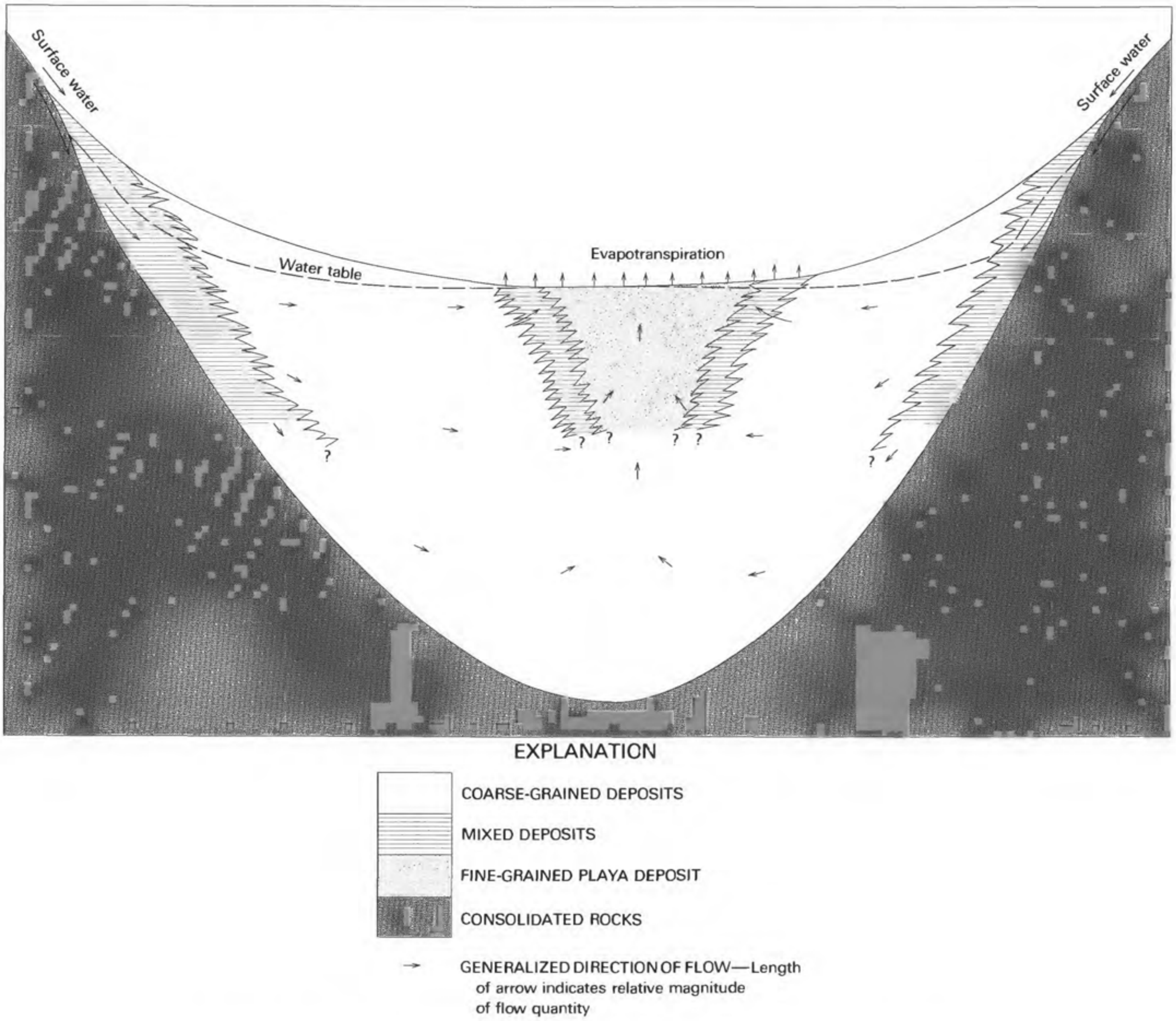


FIGURE 10.—Conceptualization of the ground-water flow system in Smith Creek Valley.

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial}{\partial z}\right)-W=S_s\frac{\partial h}{\partial t},$$

where K = hydraulic conductivity in the horizontal (x and y) and vertical (z) directions;
h = the potentiometric head;
W = volumetric flux per unit volume, which represents a source or sink of water;
S_s = specific storage of the porous material; and
t = time.

This equation is solved by a finite-difference approach using a computer program written by McDonald and Harbaugh (1984). The method of simulating ground-water flow by finite-difference techniques has been discussed by many authors including Remson and others (1971) and Wang and Anderson (1982).

Application of the computer program to solve ground-water flow problems requires knowledge of the following hydrogeologic conditions:

(1) Hydraulic properties of the deposits in the aquifer,

(2) the shape and physical boundaries of the aquifer system,

(3) flow conditions at the boundaries, and

(4) initial conditions of ground-water flow and water levels.

The accuracy of the calibrated mathematical model is dependent on the assumptions and approximations in the finite-difference numerical solution and the distribution and quality of data. Hydraulic properties of the aquifer deposits (estimated by model calibration) can be used to define the flow system and evaluate impacts that would be produced by changes in stress, such as pumping. However, three main limitations that constrain the validity of the model (Harrill, 1982) are:

(1) The inability of the numerical model to simulate all the complexities of the natural flow system. The assumptions used for construction of the model affect the output and are simplified relative to the natural conditions.

(2) The distribution of field data; for example, water-level or lithologic data may not be areally or vertically extensive enough to define the system adequately.

(3) The model probably is not unique. Many combinations of aquifer properties and recharge-discharge distributions can produce the same results, particularly because the model was calibrated for a predevelopment (steady-state) condition. For example, a proportionate change in Q (total sources and sinks of water) with respect to T (transmissivity) would result in the same steady-state model solution. One way to better calibrate the model is to use the water-level changes caused by documented historic pumping. This additional calibration would provide a check on the aquifer properties used in the model. Smith Creek Valley, however, has not as yet had sufficient pumpage and enough pumping time to make noticeable changes in the head distribution, so the model could not be calibrated against historic water-level changes. However, if the values used to represent hydraulic properties of the aquifer system are reasonable, then the model will be closely related to the flow system.

The computer program used for this study accommodates a multilayered heterogeneous aquifer with irregular physical and flow boundaries, discharge from wells, and evapotranspiration from the water table. Heterogeneity in the aquifer system caused by variation in the types of deposits was simulated by: (1) Varying the aquifer properties between model blocks and (2) averaging aquifer properties in each model block to represent the aggregate of the heterogeneity within that block. Model output was evaluated by comparing: (1) Simulated areal distributions and rates of evapotranspiration to estimated values, (2) simulated hydraulic heads to the measured heads, and (3) simulated head differences between the 50-ft and 50- to 500-ft intervals to the measured differences.

MODEL CONSTRUCTION

The aquifer system was divided into blocks by superimposing a grid over a map of the study area. The grid was aligned parallel to the elongate north-northeast-trending axis of the valley to minimize the number of inactive model blocks. The basin-fill aquifer is about 32 mi long and a maximum of about 11 mi wide. The grid was constructed with 1-mi spacing east to west and 2-mi spacing north to south, because of the elongate shape of the valley and the sparse spatial distribution of the hydrologic data. The initial grid was constructed by using the geologic map that delineates the boundary between basin fill and consolidated rocks (pl. 1). This grid was later refined to eliminate model blocks that did not contain saturated unconsolidated deposits. The resulting grid is 11 by 16 model blocks and contains 114 active blocks (fig. 11).

The Smith Creek Valley aquifer system was subdivided vertically into three layers for modeling purposes. The top layer (layer one) was chosen as the uppermost 50 ft of saturated basin fill because the vertical gradient between wells at depth intervals of 22–25 and 64–75 ft in the discharge area is upward, yet no confining layer is distinguishable. The approximate midpoint between the well groups was arbitrarily chosen as the bottom boundary for layer one. The top 50 ft of saturated basin fill is largely an unconfined water-table layer. The middle layer (layer two) was chosen as the 50- to 500-ft interval of saturated basin fill because lithologic information from wells is not available below a depth of 500 ft. The bottom layer (layer three) is from 500 ft to the top of consolidated rock. Layers two and three are treated as confined aquifers. Recharge in the model is input to the top layer. The gradient is downward and the water moves down into the second and third layers and laterally toward the playa. In the playa area, the vertical gradient is upward from the second and third layers to the top layer and water moves upward and discharges by evapotranspiration.

FLOW-BOUNDARY CONDITIONS

Specified-flux and head-dependent boundaries were used in the Smith Creek Valley flow model. All lateral and vertical boundaries between the consolidated rocks and basin-fill aquifer were specified as zero-flux boundaries. Recharge was introduced into the flow model as a specified-flux boundary in the first model block inside the no-flow boundary around the periphery of layer one. This was simulated so that the recharge occurred along the mountain front of the basin. The top phreatic boundary is head dependent; consequently, the discharge rate

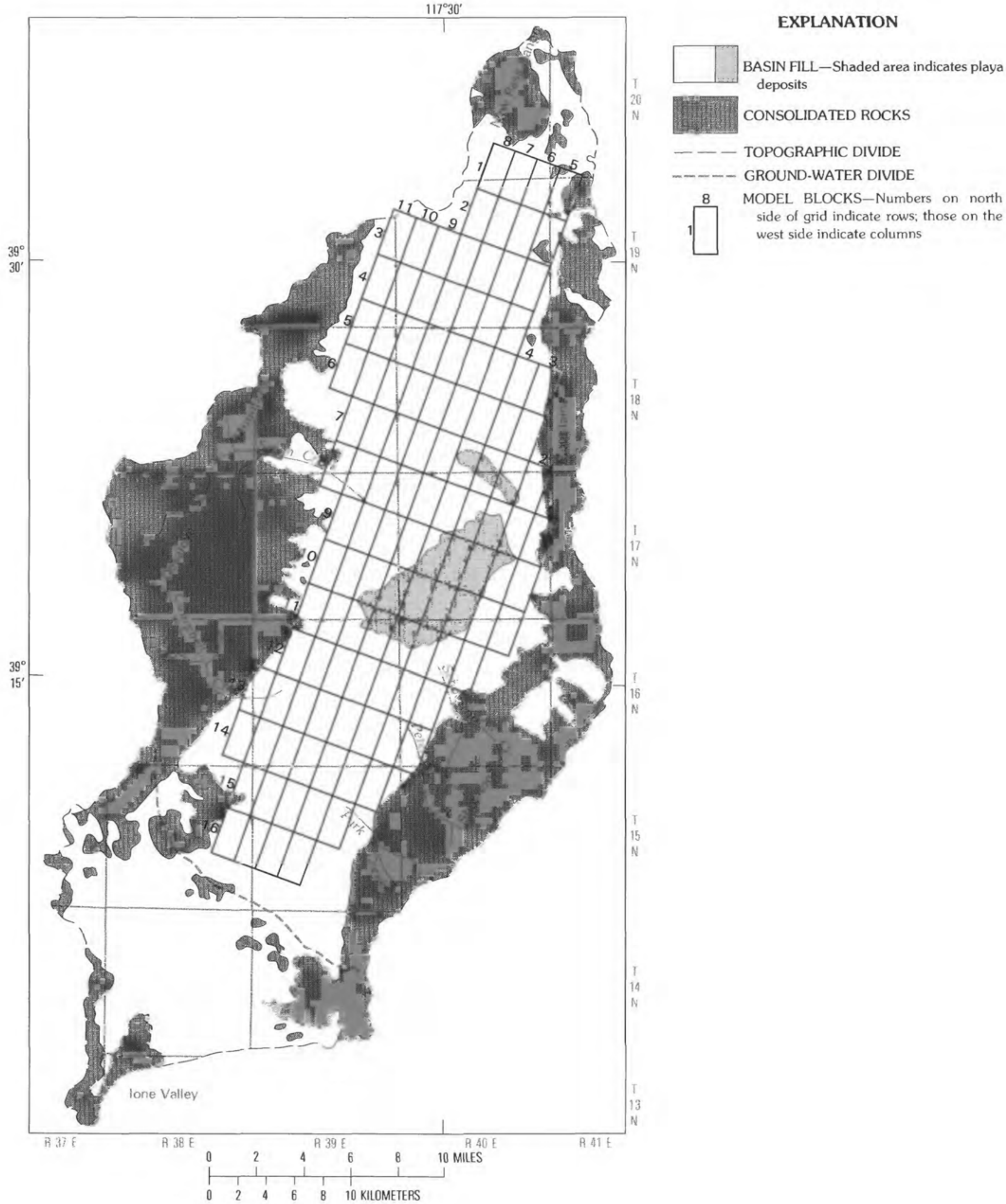


FIGURE 11.—Finite-difference grid for digital model.

depends on the depth to water. Maximum discharge occurs when water is at the land surface, and the rate decreases linearly with depth to zero at a set depth.

RECHARGE

Annual recharge estimates are 8,300 and 9,600 acre-ft, for the chloride-balance and Maxey-Eakin methods, respectively (see section "Ground-Water Recharge"). For a steady-state flow system, recharge should equal natural discharge which is 8,300 acre-ft/yr, as indicated in table 9. This suggests that the recharge estimates are probably reasonable. However, because the discharge rate is based on field observations, rather than the empirical values that were used to estimate recharge, the recharge rate for the model was set equal to the discharge rate of approximately 8,000 acre-ft/yr.

Adjusted recharge rates for each peripheral model block in layer one (the water-table layer) were determined by multiplying the Maxey-Eakin estimations of recharge for corresponding drainage areas by 0.833 (the ratio of estimated natural discharge to the Maxey-Eakin estimate of recharge) to obtain a reduced recharge rate that is equivalent to the more accurately known discharge rate.

DISCHARGE

Ground water is evaporated or transpired from the top 30 ft, or less, of basin fill in Smith Creek Valley. Therefore, evapotranspiration is limited to the top layer of the ground-water flow model. Evapotranspiration in the computer code is computed from a maximum evapotranspiration rate at land surface and a depth at which evapotranspiration ceases, the extinction depth. In this report extinction depth was assumed to be 12 to 30 ft below land surface and will be discussed later. Evapotranspiration of ground water is simulated in the model where the depth to water is less than the extinction depth. Depth to water was determined by subtracting the water-level altitude (pl. 2) from the estimated land-surface altitude. Land-surface altitudes were estimated for each model block in the discharge area from altitudes at wells in the playa area surveyed by F. E. Rush and J. R. Harrill (U.S. Geological Survey) during 1965–66. The accuracy of land-surface estimates in the discharge area is generally within 1 ft. The evapotranspiration rate is calculated from a rate-versus-depth relationship where the maximum rate is at land surface and the rate decreases linearly to zero at the extinction depth.

Extinction depth was assigned based on the presence or absence of phreatophytic vegetation and the depth to water. The extinction depth beneath the playa was set at 12 ft. Extinction depths for model blocks in the phreatophytic fringe surrounding the playa were set so that they were at the same altitude as those beneath the playa or at a maximum depth of 30 ft below land surface. This was done to maintain the flat water-level gradient in the central part of the basin. A maximum evapotranspiration rate of 1.5 ft/yr at land surface was assumed and used for the flow model, except for the playa, where 3.0 ft/yr was used.

Discharge from the hot springs is by direct evaporation or by transpiration of water recirculated to the shallow water table. Therefore, discharge in the hot-spring area was simulated as evapotranspiration from the top layer of the model.

INITIAL CONDITIONS AND HYDRAULIC PROPERTIES

The initial head distribution for model layer one (the water-table layer) is based on 1982 water-level measurements (pl. 2). This water-table surface is believed to represent the long-term steady-state condition. The initial head distribution for model layer two is assumed to be the same as layer one except in the discharge area, where the heads are from 0 to 14 ft higher in layer two than in layer one. No water-level data were available for model layer three, so the water levels in this layer were set equal to those in model layer two, even though the levels in layer three would be expected to be higher than those in layer two beneath the discharge area.

Transmissivities of the unconsolidated basin-fill deposits used in the model simulations are the product of the hydraulic conductivity and saturated aquifer thickness. The horizontal hydraulic conductivity in the top 50 ft and in the interval from 50 to 500 ft of saturated basin fill is distributed on the basis of lithology as shown in figure 3. In model layer one, coarse-grained deposits were assigned a conductivity value of 4×10^{-4} ft/s, fine-grained deposits beneath the playa were assigned a value of 1×10^{-7} ft/s, mixed deposits on alluvial fans and adjacent to coarse-grained deposits in the discharge area were assigned a value of 1×10^{-5} ft/s, and mixtures of fine-grained and mixed deposits adjacent to the playa were assigned a value of 1×10^{-6} ft/s. In model layer two, coarse-grained deposits were assigned a conductivity value of 4×10^{-4} ft/s, fine-grained deposits were assigned a value of 1×10^{-6} ft/s, and mixed deposits were assigned a value of 1×10^{-5} ft/s. Hydraulic conductivity was adjusted during the calibration process. Model layer three was assumed to have a lithology distribution similar to layer two, but hydraulic conductivity was assumed to decrease linearly with increasing depth due to overburden pressure. The resulting hydraulic conductivity in layer three was calculated by assuming that the conductivity decreased 25 percent for each 500-ft depth increase (Durbin and others, 1978) from the top of the layer to the consolidated-rock contact.

Two approaches were used to assign transmissivities in the model. Transmissivity in layer one, which is unconfined, was calculated by the computer program. For each model iteration, saturated thickness was calculated by subtracting the bottom altitude of layer one from the altitude of the water table. This value was then multiplied by the assigned hydraulic conductivity to obtain the transmissivity. Model layers two and three are confined; therefore, their saturated thickness remains constant. Transmissivities for these layers were calculated by multiplying the saturated thickness of the layer by the assigned hydraulic conductivity for layers two and three.

Vertical hydraulic conductivity is incorporated into a leakance term. Leakance (L) as defined by Lohman (1972, p. 30) is the ratio of the vertical hydraulic conductivity of the confining bed to the thickness of the bed. Leakance is used in the computer model to simulate vertical flow between model layers. Because the sediments in Smith Creek Valley are composed of numerous discontinuous lenses of coarser and finer grained deposits, an equivalent leakance value between model layers, rather than a single leakance value for each bed, was estimated using the following equation:

$$L' = K_z'/b \ ,$$

where L' = equivalent leakance;
K_z' = equivalent vertical hydraulic conductivity of the sediments;
and b = thickness between the centers of two adjacent model layers.

The equivalent vertical hydraulic conductivity of the sediments was calculated from the method proposed by Freeze and Cherry (1979, p. 34, equation 2.31) assuming that:

(1) Ground-water flow was perpendicular to the changes in hydraulic conductivity;

(2) each type of deposit (coarse, fine, and mixed) was isotropic; and

(3) the distribution of deposits in model layer three (lowest layer) was the same as that in model layer two.

The estimated leakance values were adjusted during model calibration to attain agreement between the simulated vertical head gradients in the discharge area and those measured in wells.

Storage values used for transient simulations are given in the "Basin-Fill Aquifer" section of this report. Specific yields were used for layer one, an unconfined aquifer, and storage coefficients were used for the confined aquifers, layers two and three.

STEADY-STATE SIMULATION

Calibration of the steady-state flow model was based on the relation between simulated and measured or estimated head values, ground-water budget, vertical gradients between model layers one and two and distribution of evapotranspiration. The model was considered calibrated when:

(1) All simulated heads were within 10 ft of measured heads (averaged over 2 mi^2 model block);

(2) mean absolute departure of simulated heads from measured heads was close to zero and the standard deviation was minimal for the 32 model blocks containing wells in layer one;

(3) mass balance of water into and out of the system had a minimal error;

(4) the average of the simulated head differences between model layers one and two was less than 2 ft of head differences derived from field measurements; and

(5) the simulated areal distribution of evapotranspiration matched the estimated distribution, and the simulated rate of discharge equaled the estimated rate.

Calibration of the model involved adjusting initial estimates of hydrologic properties to attain a best fit. The recharge rate, which is based on discharge estimates from field observations, was initially held constant because discharge was considered the best known of the hydrologic factors used in the model. First, the hydraulic conductivities were adjusted by changing the hydraulic conductivity of a specific lithologic unit, such as coarse-grained deposits, or by changing conductivities of all the lithologic units by the same factor. Second, the vertical leakance values between layers one and two were adjusted until the model could simulate closely the measured head differences between the layers. To be consistent, the leakance values between layers two and three were also adjusted by the same amount as those between layers one and two. Finally, discharge and recharge rates were adjusted to determine whether different rates could produce a better fit. An adjustment of one parameter often required the adjustment of other parameters; for example, a change in transmissivity would necessitate a change in leakance to maintain the observed head differences between model layers one and two.

Hydraulic properties of the aquifer system used in the model simulations were averaged over large (1-mi by 2-mi) model block areas; therefore, an estimate of a property for a model block can contain a wide range of values. The final calibrated transmissivities for all model layers and leakance values between model layers were calculated using the final adjusted hydraulic conductivities of 2.5×10^{-4} ft/s for coarse-grained deposits, 1.5×10^{-5} ft/s for mixed deposits, 2×10^{-6} ft/s for deposits consisting of mixtures of fine-grained and mixed deposits around the margin of the playa in model layer one and fine-grained deposits in layer two, and 1×10^{-7} ft/s for fine-grained playa deposits in model layer one. The final distribution of hydraulic conductivity for layer one is shown in figure 12. The final calibrated transmissivities of layers two and three are shown in figures 13 and 14, respectively. The final leakance values used in the calibrated steady-state model between model layers one and two are shown in figure 15.

All simulated heads for layer one in the steady-state solution are within 10 ft of measured heads (averaged over a 2 mi^2 model block), and most are within 5 ft (fig. 16). The mean absolute departure of the simulated heads from measured heads, for the 32 model blocks in layer one containing wells, was +0.22 ft, with a standard deviation of 4.5 ft. Simulated heads north of the playa area are all slightly higher than the measured heads because the nearly flat hydraulic gradient in the northern half of the basin is difficult to simulate. Increasing the transmissivities of the coarse-grained deposits in the valley to lower these simulated heads caused too much water to move toward the discharge area, which in turn caused an unrealistic ground-water mound around the margin of the lower-transmissivity playa. In addition, a slight mounding of water simulated by the steady-state model near the southern margin of the playa is due to the four-order-of-magnitude difference in hydraulic conductivities between the coarse-grained deposits and fine-grained playa deposits there. Simulated heads in the playa area are generally a little lower than measured heads. Water-level altitudes in these model blocks are controlled primarily by extinction depths, so the simulated heads could be raised by decreasing the extinction depths. However, this would also cause heads north of the playa area to increase, which would result in a worse overall fit of the model simulation. Therefore, the final distribution of hydraulic properties is a best fit for measured versus simulated head, considering the assumptions that went into construction of the model by using the criteria stated at the beginning of this section.

Simulated head differences between layers one and two are controlled primarily by the leakance values. These values were adjusted, within a lithologically realistic range, until simulated head differences were matched as closely as possible with head differences derived from field measurements. The average of simulated head differences between model layers one and two was less than 2 ft of the measured head differences (fig. 17). However, an exact match was never obtained because of the multiple interdependent hydrologic properties that influence vertical gradients. Simulated head differences in the fine-grained playa deposits were slightly less than the measured differences. This disparity is due to the small flux of water into this area as a result of the extremely low hydraulic conductivity of the playa deposits. In addition, the southeasternmost model block, containing upward vertical flow, has a simulated head difference of 5 ft as compared to a measured difference of 1 ft, because the block is adjacent to the low transmissivity playa deposits and it receives a large flux of water. This results in an increase in water levels in the block, which produces a greater head difference.

Total evapotranspiration simulated by the steady-state model equals total recharge, but the simulated distribution of discharge is different from the distribution estimated on the basis of phreatophyte mapping (fig. 18). The simulated area is slightly larger than the mapped area, as evidenced by the two model blocks north of the mapped discharge area that contain simulated evapotranspiration. This simulated enlargement of the discharge area is primarily the result of the flat water-level and land-surface gradients in this part of the basin and the fact that simulated heads are 5 to 7 ft higher than measured heads in the northern part of the basin. These simulated heads were high enough in two model blocks north of the area of estimated discharge to bring water levels above the 30-ft extinction depth for evapotranspiration, thereby inducing simulated discharge. The contribution of these two additional model blocks to the total simulated evapotranspiration was 7.5 percent, so the two blocks had little impact on the overall distribution of discharge.

SENSITIVITY ANALYSIS

A sensitivity test was made by changing values of the hydraulic properties uniformly after the flow model was calibrated. Hydraulic conductivities, recharge rates, and discharge rates used in the calibrated steady-state model were tested, individually, to determine the effect of doubling the original values at each parameter and the effect of reducing the original values by one-half. In addition, a spatial sensitivity analysis was performed by changing the hydraulic conductivity of just the coarse-grained deposits (conductivity was both

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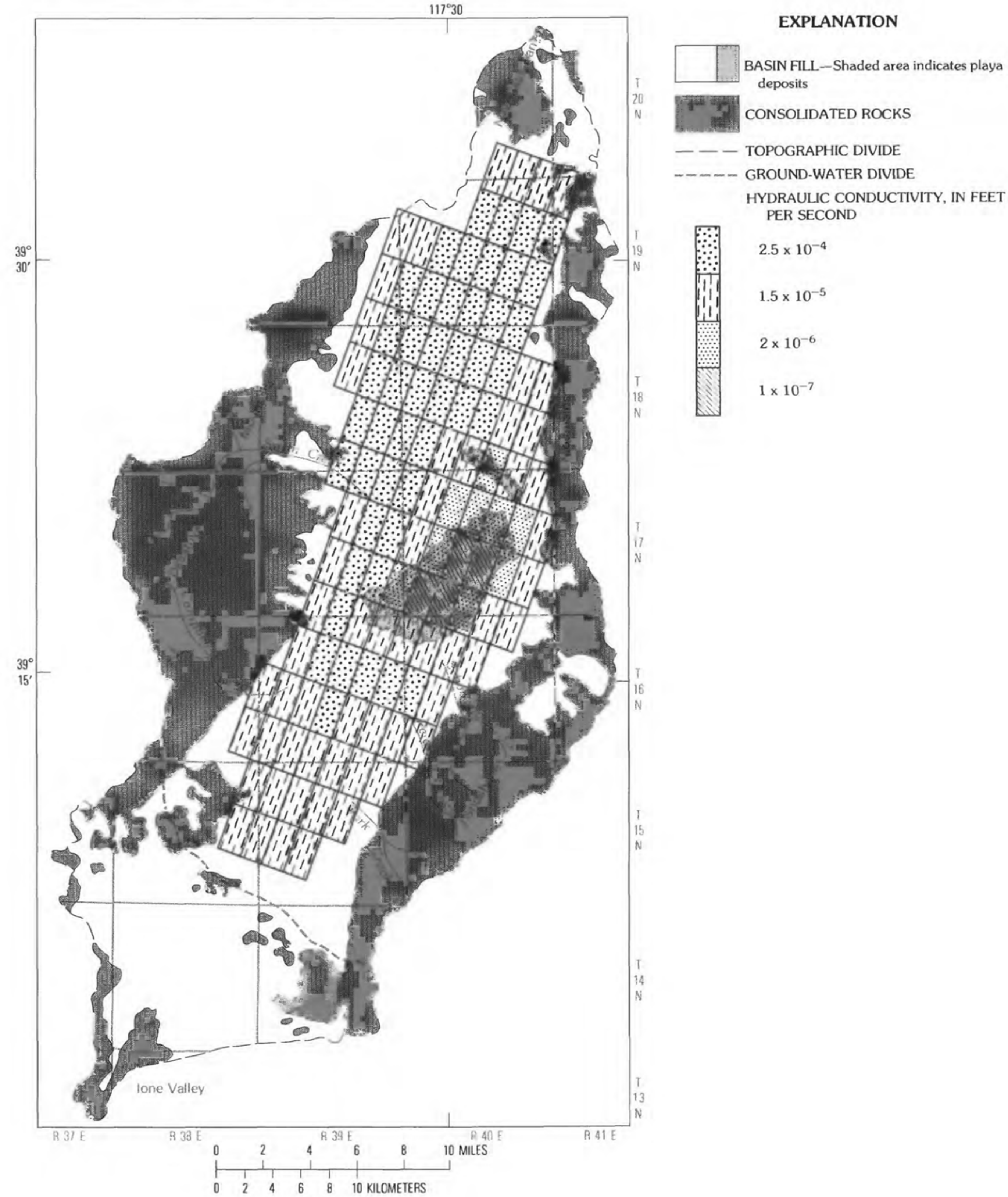


FIGURE 12.—Hydraulic conductivities for upper 50 ft of saturated basin fill (model layer one), based on steady-state simulation.

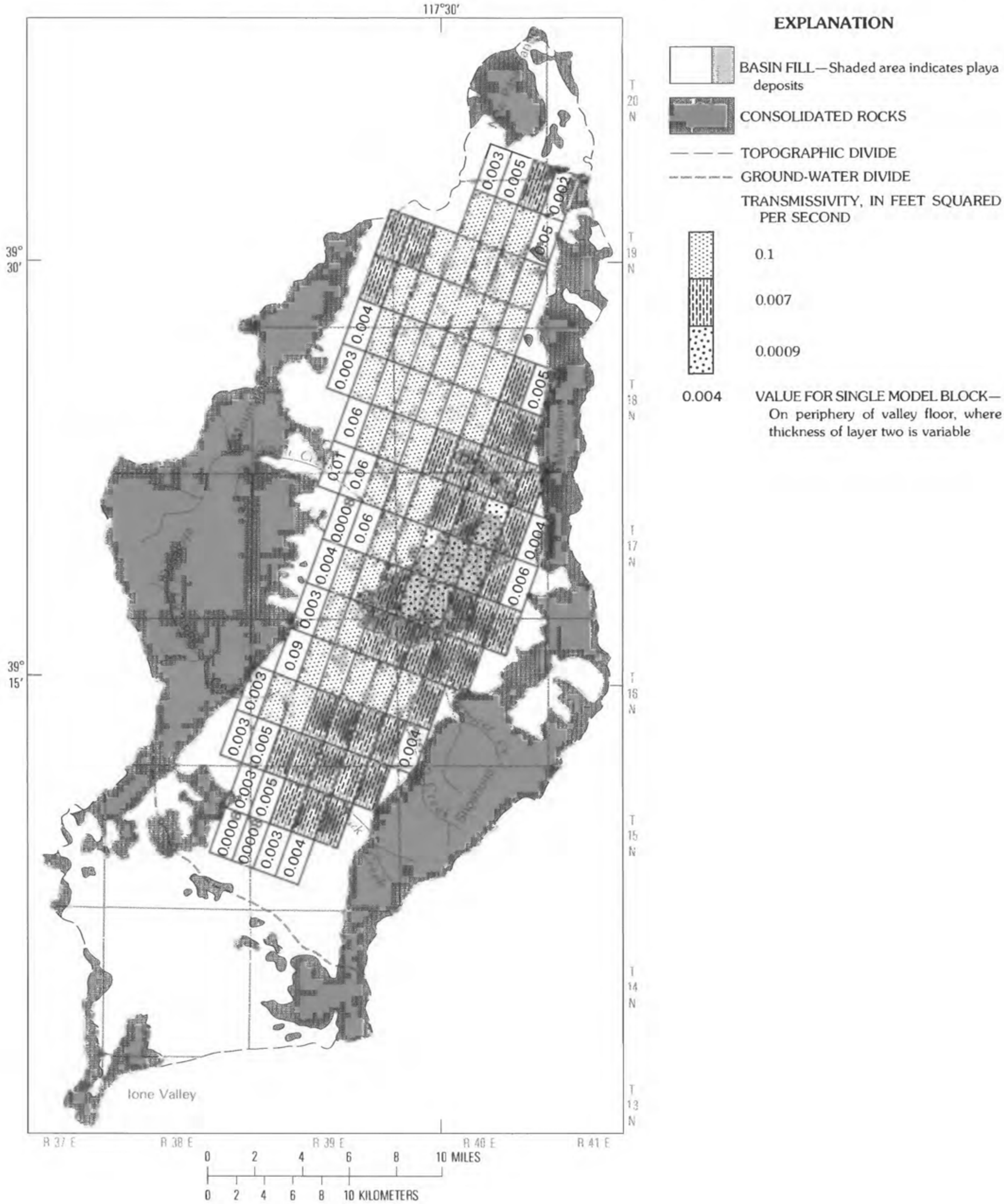


Figure 13.—Simulated aquifer transmissivities for 50- to 500-ft interval of saturated basin fill (model layer two), based on steady-state simulation.

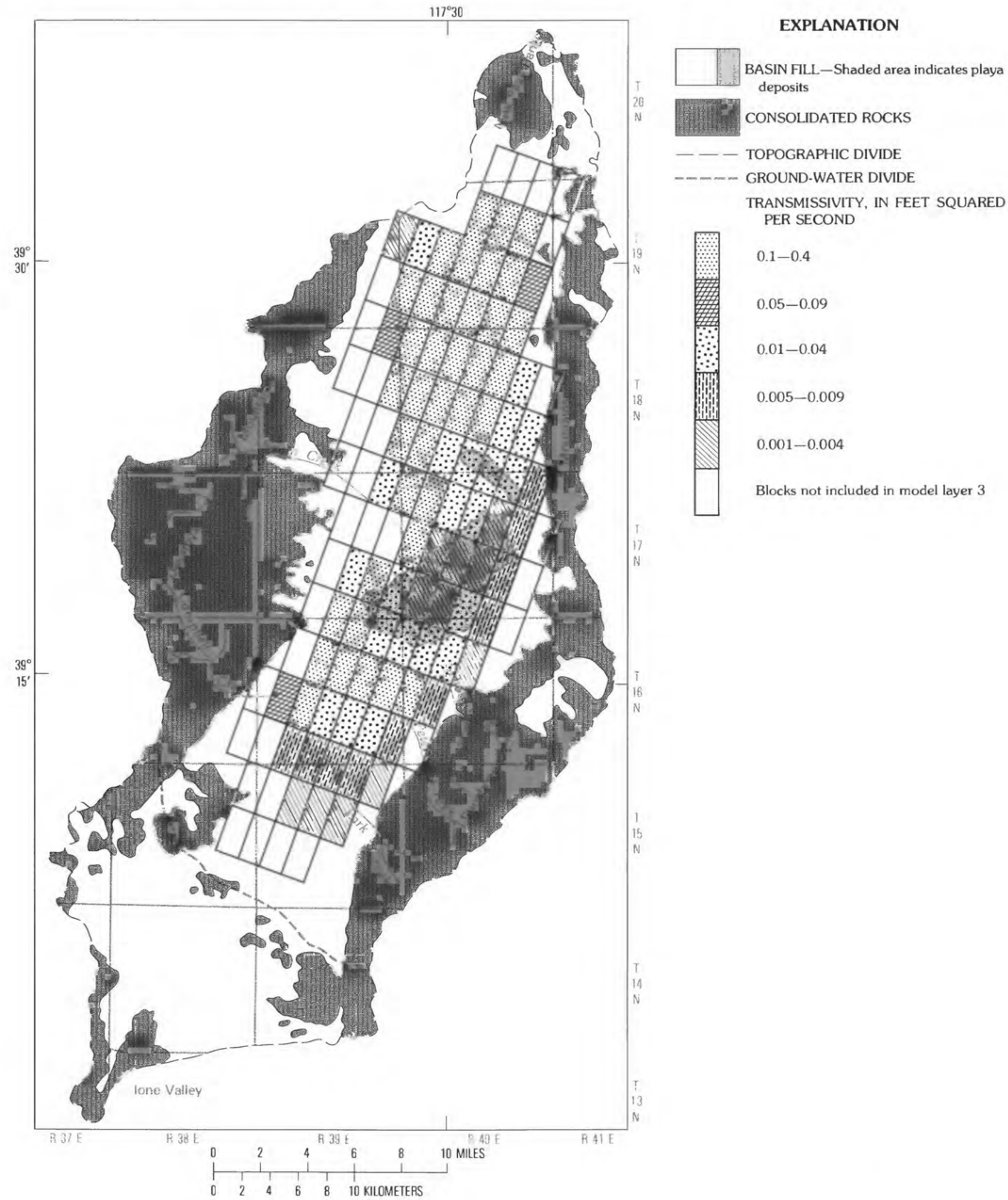


FIGURE 14.—Simulated aquifer transmissivities for 500 ft to consolidated rocks of saturated basin fill (model layer three), based on steady-state simulation.

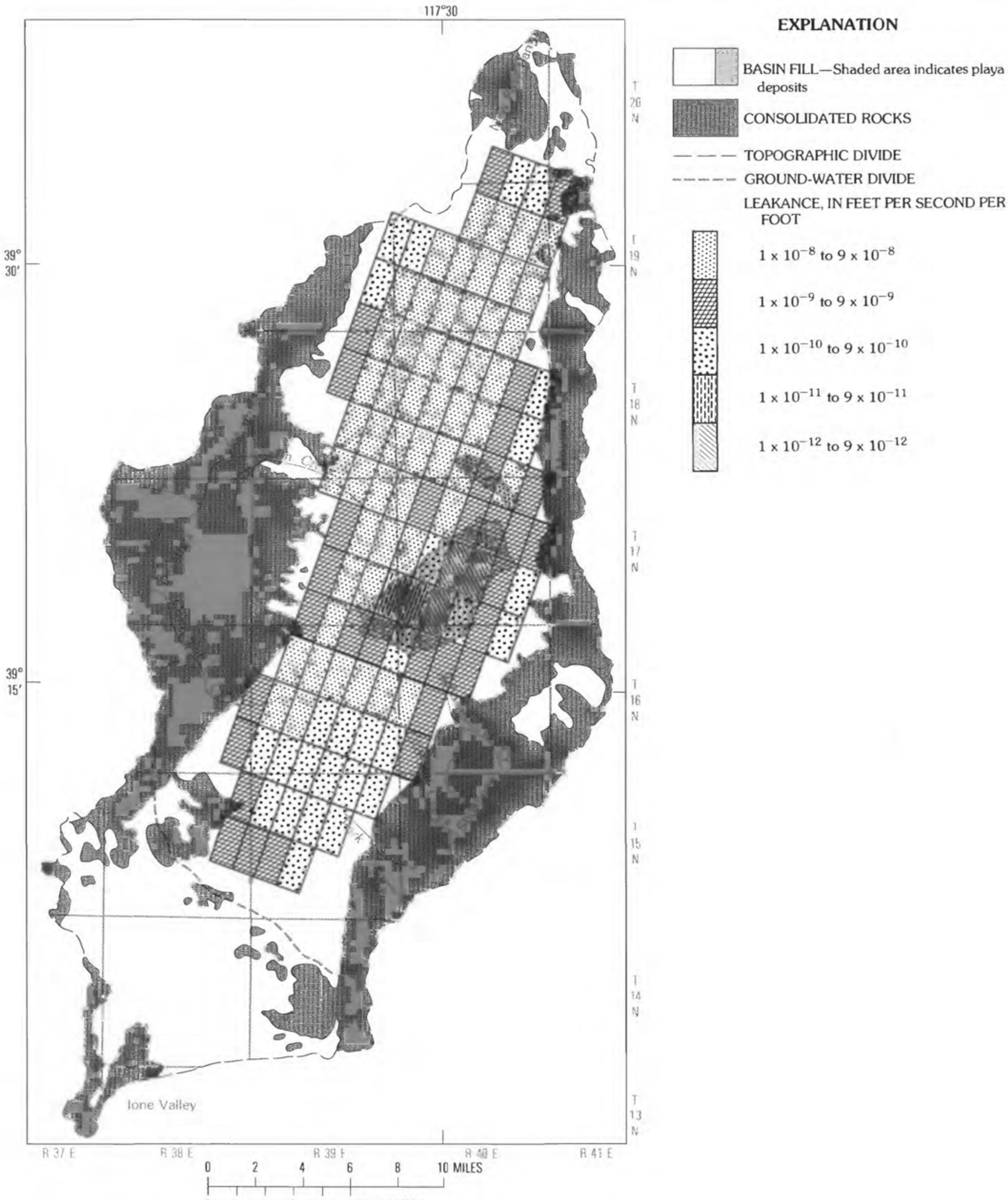


FIGURE 15.—Simulated leakance values between model layers one and two for the steady-state simulation. Initial calculated values were adjusted to minimize difference between simulated and measured head differences.

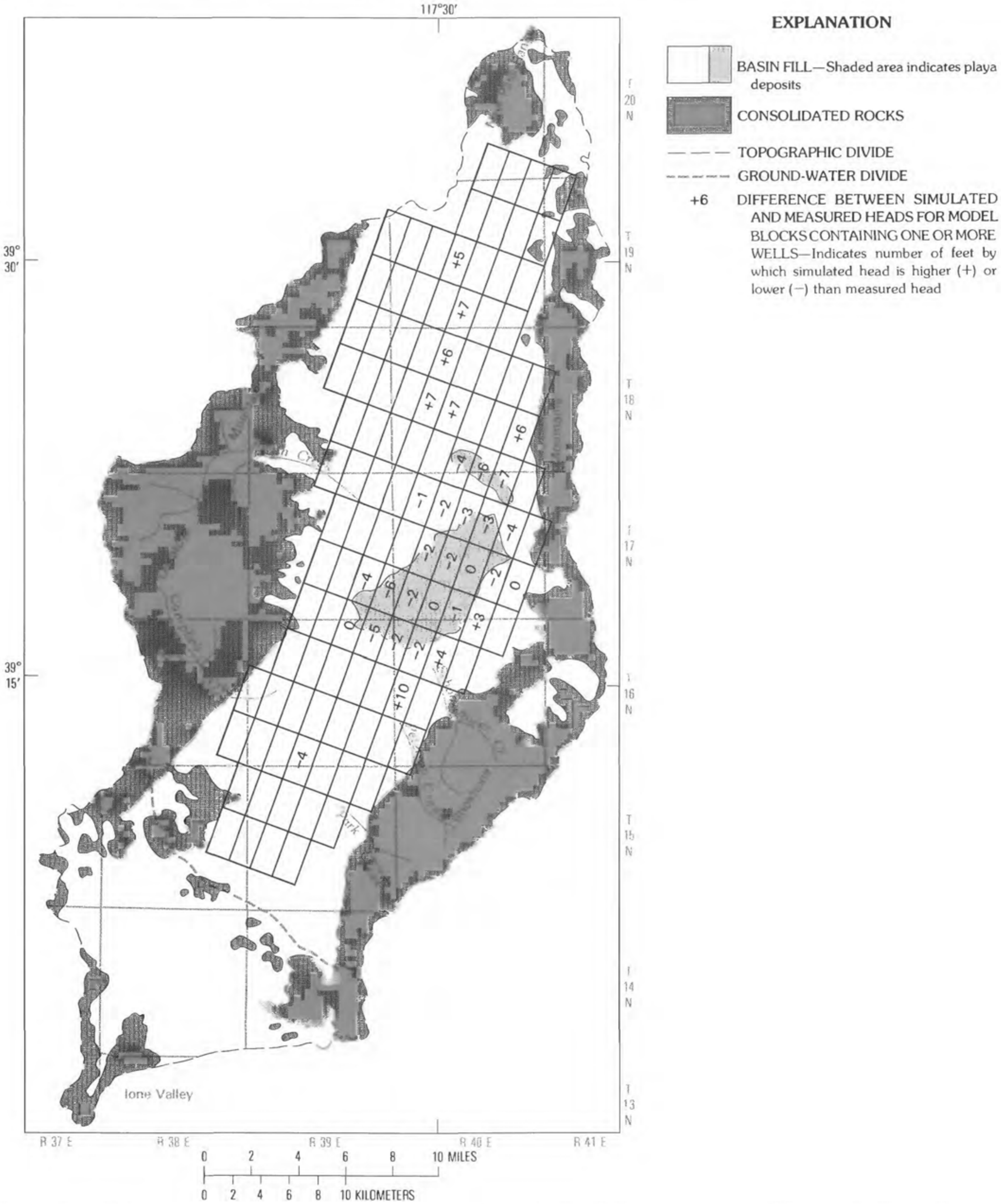
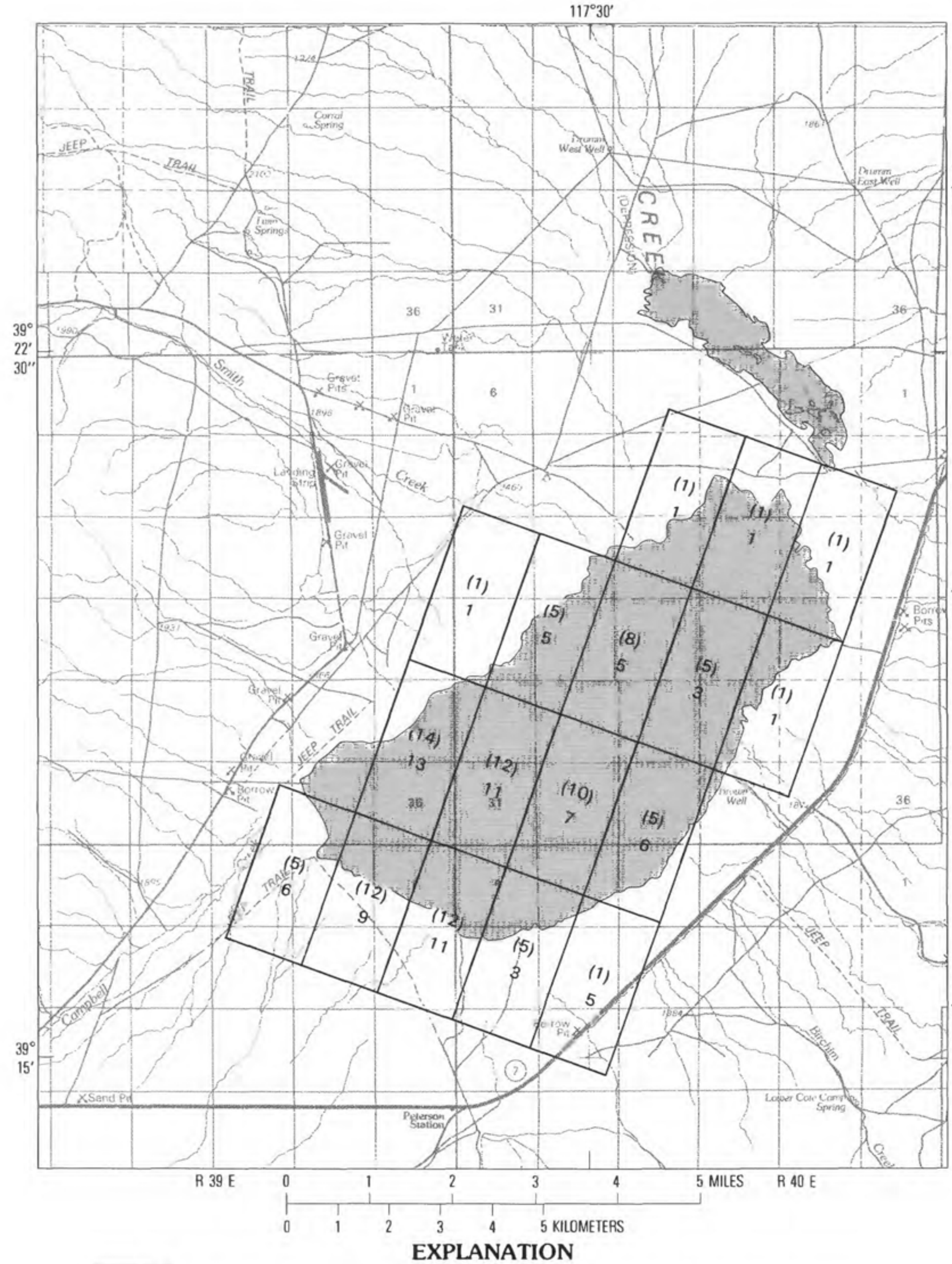


FIGURE 16.—Difference between simulated and measured (averaged over 2 mi² model blocks) heads in model layer one for the steady-state solution.



EXPLANATION

BASIN FILL—Shaded area indicates playa deposits

(14) MEASURED HEAD DIFFERENCE

13 SIMULATED HEAD DIFFERENCE

FIGURE 17.—Head differences between model layers one and two caused by fine-grained deposits in the playa area. In all blocks, head in layer two is higher than head in layer one.

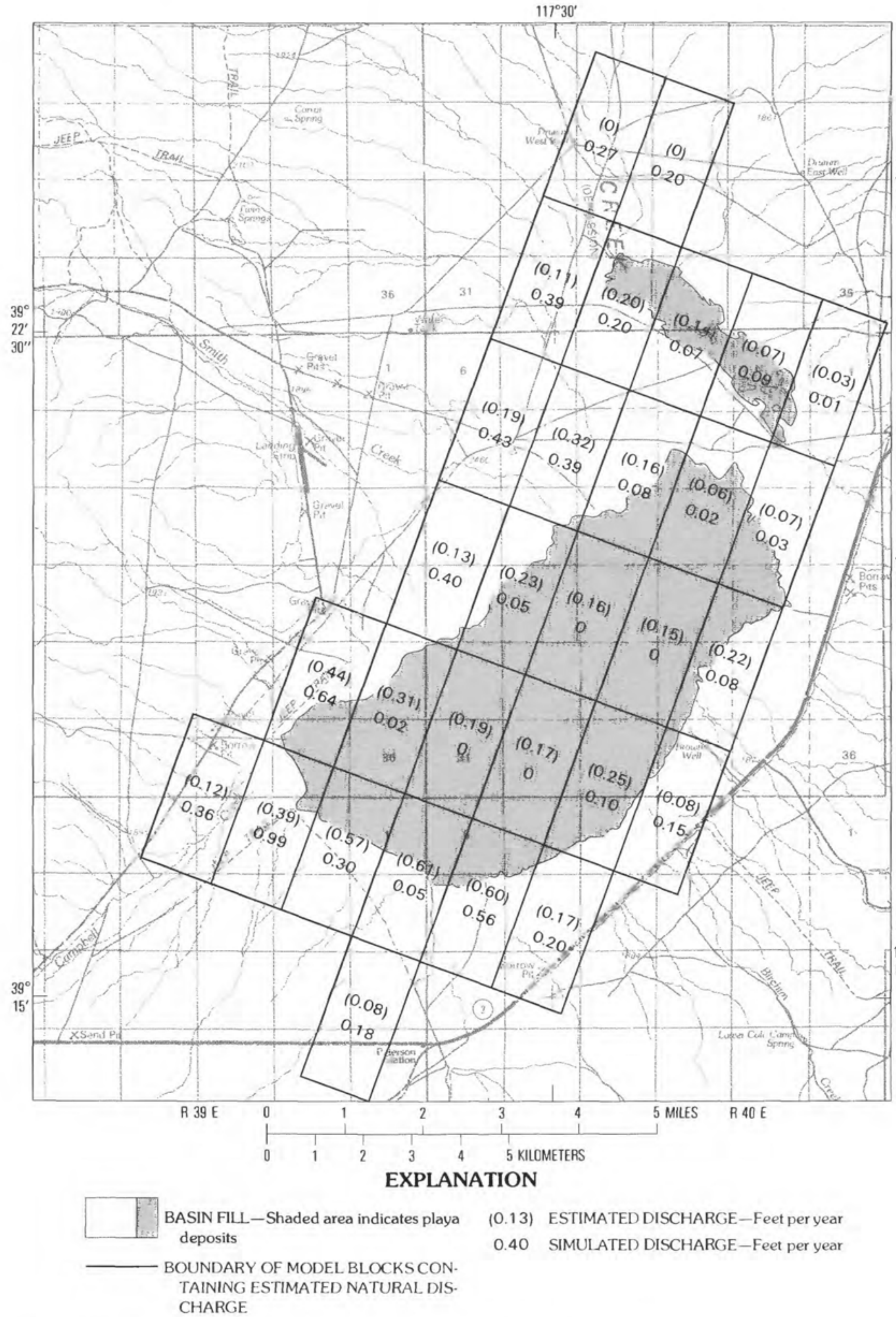


FIGURE 18.—Comparison between estimated and simulated natural discharge for the steady-state simulation.

doubled and reduced by one-half). The sensitivity of these properties was evaluated against (1) simulated versus measured heads in model layer one, (2) simulated versus measured head differences between model layers one and two, (3) water-budget balance, and (4) simulated versus estimated rates and areal distribution of discharge (table 10).

As presented in table 10, all changes in hydraulic properties produced an increase in (1) the absolute mean head difference, standard deviation, and maximum head difference above or below measured values in model layer one, and (2) the mean head difference between model layers one and two. Water-budget balances varied little, but the discharge area increased significantly to maintain the balance for some simulations. No single value, with the exception of discharge, was significantly more sensitive to change than another. The greater sensitivity of discharge to increasing evapotranspiration rates is due to the small thickness of the top model layer. Decreasing discharge produced model reactions similar to those observed for changes in other properties.

LIMITATIONS OF THE MODEL

The computer model represents a simplification of the ground-water flow system in Smith Creek Valley and was calibrated only for steady-state conditions. Furthermore, only a limited amount of hydrologic data exists for the valley, so hydrologic conditions had to be inferred for some parts of the basin-fill aquifer, especially around the margin of the valley and at depths greater than 500 ft in the basin fill. Consequently, the simulation results can be used to evaluate general responses of the basin-fill aquifer to pumping, but actual pumping may produce significantly greater or lesser responses than those that have been simulated.

SIMULATED RESPONSE OF THE GROUND-WATER SYSTEM TO HYPOTHETICAL DEVELOPMENT SCENARIOS

Drawdown and recovery of ground-water levels in Smith Creek Valley were simulated using the calibrated flow model to determine long-term responses of the aquifer system to hypothetical pumping scenarios. Model simulations were run for an arbitrary period of 600 yr—300 yr of pumping and 300 yr of recovery—to allow the flow system to respond to stresses and to approach a new equilibrium after termination of the pumping.

Hypothetical development scenarios were constrained with regard to the positioning of hypothetical pumping

TABLE 10.—*Sensitivity analysis of the steady-state simulations*

Model simulation	Difference between simulated and measured heads in layer 1 (ft): Absolute mean	Standard deviation	Maximum above measured	Maximum below measured	Approximate mean difference between simulated and measured heads for 17 model blocks in playa area between layers 1 and 2 (ft)	Mass balance of water entering and leaving the flow system[1] (percent discrepancy)	Number of model blocks containing simulated evapotranspiration
Calibrated simulation-----	0.2	4.5	10	7	1.4	0.73	29
Horizontal and vertical hydraulic conductivities:							
100-percent increase--	2.3	7.0	4	38	2.8	1.16	28
50-percent decrease---	1.8	8.1	27	8	3.0	.27	40
Recharge rate:							
100-percent increase--	2.3	7.0	4	38	2.8	1.16	28
50-percent decrease---	3.7	7.3	2	41	3.3	1.16	27
Evapotranspiration rate:							
100-percent increase	-------- 14 model blocks go dry in the discharge area, prohibiting solution ------------						
50-percent decrease---	2.4	5.6	14	6	2.1	.68	33
Hydraulic conductivity of coarse-grained deposits:							
100-percent increase--	.3	4.6	8	9	1.7	.98	30
50-percent decrease---	1.3	7.4	8	17	1.9	.04	36

[1] All mass balances are negative (that is, water leaving system exceeds water entering system).

wells. The wells were not located (1) in areas of fine-grained deposits having low transmissivity, (2) within a mile of saline water, (3) where the depth to water exceeds 200 ft, (4) where the thickness of saturated basin fill is less than 200 ft, (5) where model blocks are bounded on two or more sides by consolidated rocks, and (6) where land-surface slopes are greater than 200 ft/mi.

All model simulations were based on net pumpage, to avoid the necessity of quantifying and simulating the recirculation of pumped water to the aquifer (irrigation return flow). In addition, by locating the pumping wells more than a mile from saline water in the playa area, the migration of the saline water into areas of pumping should not be a problem because:

(1) The low hydraulic conductivity of the fine-grained playa deposits would greatly retard the lateral movement of the saline water from the playa area to the pumping area. With such slow movement, the migration of the saline water to a pumping well more than a mile away should be negligible.

(2) Considering the upward movement of fresh ground water beneath the playa area, the downward movement of shallow saline water from the playa into the lower aquifer and then moving toward the pumping area would be improbable. Even when most of the natural discharge is captured by pumping, some ground water will still flow to the playa area, maintaining a slight upward hydraulic gradient beneath the playa.

Simulated development scenarios were:

1. Pumping at a rate equal to the estimated average annual recharge, with (a) a distribution of pumping that would most efficiently reduce evaporation and transpiration in the discharge area (*scenario A*), (b) a center of concentrated pumping between the recharge and discharge areas, west of the playas (*scenario B*), (c) a center of concentrated pumping in the northern part of the valley, away from the discharge area (*scenario C*), and (d) two centers of concentrated pumping near the discharge area (*scenario D*).

2. Pumping at twice the recharge rate using a well distribution that would most efficiently reduce evaporation and transpiration in the discharge area (*scenario E*).

3. Pumping at twice the recharge rate for the first 50 yr, then pumping at the recharge rate for the subsequent 250 yr, with the same well distribution as used in scenarios A and E (*scenario F*).

Development scenarios A–D were simulated to test the feasibility of the concept of sustained yield, in which pumping at a rate equal to the average annual recharge efficiently reduces evapotranspiration. This process of reducing evapotranspiration was defined by Lohman and others (1972) as capture. This approach was chosen not only to test the application of the sustained-yield concept, but also because the States of Nevada and Utah generally allocate water rights on the basis of the average annual recharge to a valley, to prevent long-term water-level declines in a valley.

The six scenarios chosen are thought to be representative, in a general way, of reasonable alternatives for possible future development of the ground-water resources in Smith Creek Valley. Neither economic considerations nor the possibility of a salt buildup caused by recirculation of water in the pumping area were addressed in the simulations, but they could become problems in the future.

SCENARIO A

The first scenario was that simulated pumping was distributed so as to maximize capture of ground water by reducing natural evapotranspiration. The pumping was distributed based on optimum location of irrigation wells. Current ground-water irrigation in the valley is by center-pivot circular sprinklers, so the hypothetical future development was chosen to follow this pattern, in which one well irrigates 125 acres within a 160-acre allotment. A net pumping rate of 2.0 ft/yr (described earlier in the "Discharge" section of this report) results in a pumpage of 250 acre-ft/yr for each well. Under these assumptions, 32 wells are required to pump at an aggregate rate equivalent to the average annual recharge. Therefore, a model block, which is 1,280 acres, could contain a maximum of eight wells. Model simulations were run for several areal configurations of pumping until a distribution was attained that efficiently eliminated natural discharge (fig. 19).

Pumping produced a decline in water levels throughout the entire valley as the aquifer system approached a new equilibrium after 300 yr of pumping (fig. 20*A*). Water-level declines were exceeding 15 ft in a broad area in the northern half of the valley and in two smaller areas near the south margin of the playa after 300 yr of pumping. Generally, declines in the southern part of the valley were less than those in the northern part because: (1) More water is recharging the southern part and (2) the basin fill in the southern part is generally finer grained. Fine-grained sediments beneath the playa damped the effect of pumping in the central and east-central parts of the valley, resulting in less drawdown there.

Drawdowns in the 20 model blocks containing pumped wells (fig. 19) averaged about 14 ft and ranged from 20 ft to about 7 ft (fig. 20*B*) after 300 yr of simulated pumping. The drawdown in a block represents the average water-level decline over the entire block, so it is less than drawdowns near or within the individual

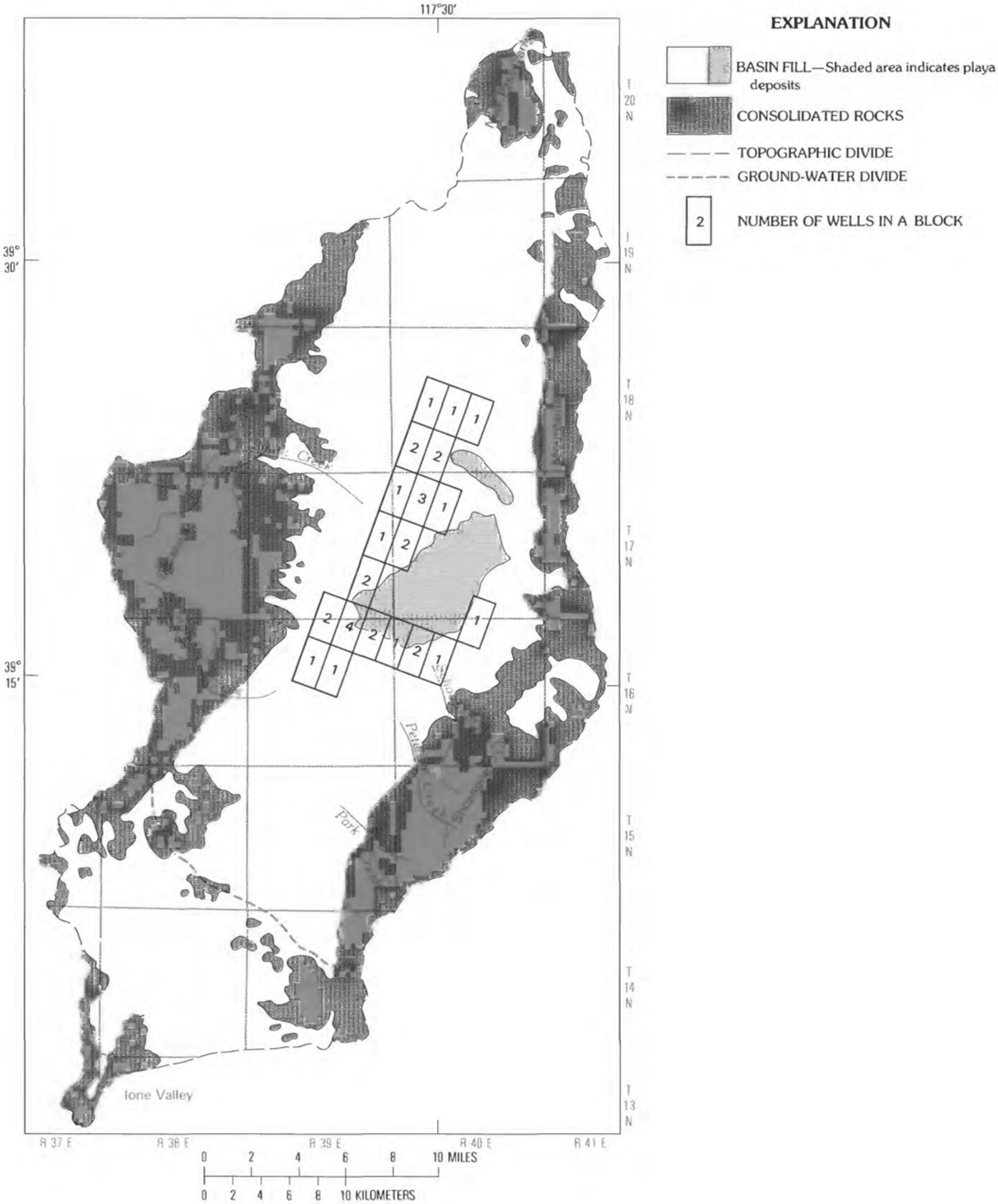


FIGURE 19.—Distribution of simulated pumping wells used to most effectively reduce natural discharge (scenario A).

pumped wells. Drawdowns for the remaining model blocks that do not contain pumped wells range from less than 1 ft to about 16 ft.

Water levels recovered rapidly after pumping ceased (fig. 20*B*). After 300 yr, simulated water levels were within 1 ft of their original water level (pre-pumping level), except near the extreme southern and northern parts of the valley (fig. 20*A*). These two areas were slow to recover due to the lack of appreciable nearby recharge.

Initially during the simulation, all water being pumped from the aquifer was coming out of storage. As pumping continued water levels declined in the discharge area resulting in a decrease in the rate of evapotranspiration. Subsequent pumping continually reduced the rate of evapotranspiration until most of the water discharged by pumping was the water that would have otherwise been discharged by evapotranspiration. Within 25 yr, as much as 75 percent of the ground water discharged by evapotranspiration had been captured by pumping (fig. 20*C*). The pumping simulation continued for another 275 yr, during which the natural discharge was further reduced but never completely eliminated.

Most of the eliminated natural discharge was upward vertical flow because simulated flow between model layers one and two was initially 10.5 ft^3/s, which represents about 95 percent of the water naturally discharged from the playa area. Twenty-five years of pumping reduced the upward flow to about 2.1 ft^3/s, which is about 19 percent of the original natural discharge.

Natural discharge increased rapidly when simulated pumping stopped; and within 25 yr, about 70 percent of the original, pre-pumping discharge rate had been regained (fig. 20*C*). Initially after pumping stopped, water recharging the aquifer accumulated in storage (fig. 20*D*), causing water levels to rise. As water levels continued to rise, however, natural discharge increased and the proportion of recharge water remaining in storage decreased. After 300 yr of recovery, natural discharge had almost returned to its original rate.

SCENARIO B

In simulation of scenario B, the pumping rate was assumed as 8,000 acre-ft/yr and concentrated in an elongated area, which was located between the recharge and discharge areas west of the playas (fig. 21A), to determine the effect of such a pumping configuration compared to that of scenario A (fig. 19). Scenario B was implemented in the model by assigning equal amounts of pumpage (667 acre-ft/yr/block) to 12 blocks.

Concentrating pumping in one area for 300 yr caused water levels to decline more than 40 ft in and adjacent to the northern half of the pumping area and more than 30 ft for the entire northern half of the valley (fig. 21*A*). Declines were generally least to the south and east. In fact, water-level declines were less than 5 ft east of the playas, where the fine-grained deposits act as a barrier to ground-water flow.

After 300 yr of simulated pumping, the drawdown averaged almost 38 ft and ranged from about 43 ft to about 15 ft in the pumped area (fig. 21*B*). Simulated drawdowns outside the pumped area ranged from a little more than 1 ft in a block east of the playas to about 42 ft in a block adjacent to the pumped area.

After the simulated pumping was terminated, water levels initially recovered rapidly (fig. 21*B*), with most water recharging the aquifer remaining in storage (fig. 21*D*). After 300 yr of recovery, water levels in the central part of the basin were within 1 ft of the pre-pumping water levels, and the maximum difference between the original and final water levels was about 4 ft near the extreme northern part and about 3 ft near the extreme southern parts of the valley (fig. 21*A*).

This scenario was less efficient in reducing natural discharge than was scenario A (figs. 21*C* and 20*C*). Consequently, more water was removed from storage (figs. 21*D* and 20*D*), so replenishment of the basin-fill aquifer and recovery of water levels in the discharge area required more time. This resulted in a slower increase in natural discharge during the recovery period than in scenario A.

SCENARIO C

Scenario C simulated an area of concentrated pumping, at the rate of 8,000 acre-ft/yr, in the northern part of the valley away from the area of natural discharge. Pumpage was distributed equally in nine model blocks (fig. 22*A*).

The 300 yr of simulated pumping produced a cone of depression in which water-level declines exceeded 100 ft, in the north part of the basin (fig. 22*A*). The average drawdown for the nine model blocks containing pumped wells was more than 100 ft, with a range from about 106 to 98 ft (fig. 22*B*). Drawdowns outside the pumped area ranged from less than 2 ft at the south end to about 100 ft at the north end of the valley.

Three-hundred years after the simulated pumping ceased, water levels had not recovered to their original prestressed levels in most of the basin (fig. 22*A*). Water levels near the extreme north end of the basin were still more than 10 ft below their original levels.

Pumpage initially came entirely out of storage, and even after 300 yr of pumping, less than 75 percent of the ground water lost to natural discharge was captured (figs. 22*C* and 22*D*). Therefore, when pumping ceased, more than 25 percent of the water recharging the aquifer was still being lost to evapotranspiration and water levels were still declining at a moderate rate in the pumped area (fig. 22*B*). Furthermore, natural discharge had recovered to only about 86 percent of the original (pre-pumping) rate 300 yr after pumping ceased, because of the large area of water-level decline in the northern part of the valley (fig. 22*A*).

SCENARIO D

Scenario D simulated pumpage of 8,000 acre-ft/yr distributed equally in four blocks north of and four blocks south of the playa (fig. 23*A*).

As expected, the configuration of pumping for this scenario produced two cones of depression (fig. 23*A*). The northern depression was deeper and more extensive than the southern one because the northern part of the valley receives considerably less recharge than the southern part (fig. 8). As in previous simulations, the fine-grained playa deposits acted as a barrier to ground-water flow, restricting drawdown east of the playa. After 300 yr of pumping, average drawdown for the eight model blocks with wells was about 36 ft, with a range from 42 to 29 ft (fig. 23*B*). Drawdown for model blocks outside the pumped area ranged from less than 1 ft east of the playa to about 40 ft adjacent to the northern pumping area.

Recovery of water levels was almost as fast for this scenario as for scenario A (figs. 23*B* and 20*B*, respectively). Water levels after 300 yr of recovery were within 1 ft of their original levels in the center of the valley and within about 4 ft near the northern part and 5 ft near the southern part of the valley (fig. 23*A*).

Natural discharge was almost completely eliminated after 300 yr of pumping (fig. 23*C*); however, the reduction was not as rapid or as complete as in scenario A (fig. 20*C*). Natural discharge increased quickly as water levels rose after the pumping stopped (fig. 23*C*), so that less than 50 percent of the recharge water was accumulating in storage after 50 yr of recovery (fig. 23*D*).

SCENARIO E

In simulation of scenario E, the aquifer was pumped for 300 yr at twice the rate (16,000 acre-ft/yr) used in scenario A, using the well distribution pattern shown in figure 19 (that most efficiently reduced evapotranspiration).

Doubling the pumping rate increased the rate and magnitude of water-level decline throughout the valley. After 300 yr of pumping at a rate of 16,000 acre-ft/yr, a prominent cone of depression in blocks containing pumping wells was about 12 times greater than that produced by pumping at the rate of 8,000 acre-ft/yr (figs. 24*A* and 20*A*, respectively). The maximum decline exceeded 200 ft and the average drawdown for all model blocks containing pumped wells was about 169 ft, with a range from 136 to 217 ft (fig. 24*B*). Drawdowns in model blocks outside the pumped area ranged from a minimum of about 60 ft, in the southernmost part of the valley, to a maximum of about 177 ft, south of and adjacent to the south boundary of the pumping area.

After 300 yr of recovery, water levels rebounded to within 5 ft of the original levels in the central part of the valley, but in the northern and southern parts of the valley water levels were still more than 40 ft below the original levels (fig. 24*A*).

Doubling the rate of pumping also completely eliminated the natural discharge in less than 25 yr (fig. 24*C*); thereafter, all pumpage in excess of recharge was derived entirely from storage (fig. 24*D*). When pumping ceased, all recharge accumulated in storage for more than 100 yr, until water levels rose sufficiently in the former discharge area to initiate some natural discharge; and even after 300 yr of recovery, natural discharge was only about 40 percent of its original rate (fig. 24*C*).

SCENARIO F

Scenario F was simulated at a pumping rate of 16,000 acre-ft/yr for 50 yr and then at a rate of 8,000 acre-ft/yr for the remaining 250 yr (fig. 25). The well distribution of scenario A (fig. 19) was used for this simulation.

The most significant response to decreasing the pumping rate after 50 yr was a rapid rise in water level in model blocks that contained pumping wells (fig. 25*C*). Pumping the valley for 50 yr at the rate of 16,000 acre-ft/yr produced rapid water-level declines, which exceeded 76 ft in a model block south of the playa (figs. 25*A* and 25*C*). The cone of depression became shallower when the pumping rate was reduced by one-half of the initial rate. After 250 yr of pumping at the reduced rate, maximum drawdown was about 35 feet and water-level declines exceeding 20 ft encompassed most of the valley (fig. 25*B*). These 20- to 30-ft declines are about 20 to 150 ft less than would have been produced if pumping had continued at the higher rate for the entire 300-yr period (figs. 24*A* and 25*B*).

After 300 yr of recovery, water levels were within 1 ft of their pre-pumped levels in the center of the valley, but they were still more than 8 ft lower than the original levels near the southern part and more than 4 ft in the northern part of the valley (fig. 25*B*). Basin-wide recovery of water levels was generally several feet less for this scenario than for scenario A because of the increased withdrawal of water from storage due to the higher initial pumping rate.

[Text continues on p. E53.]

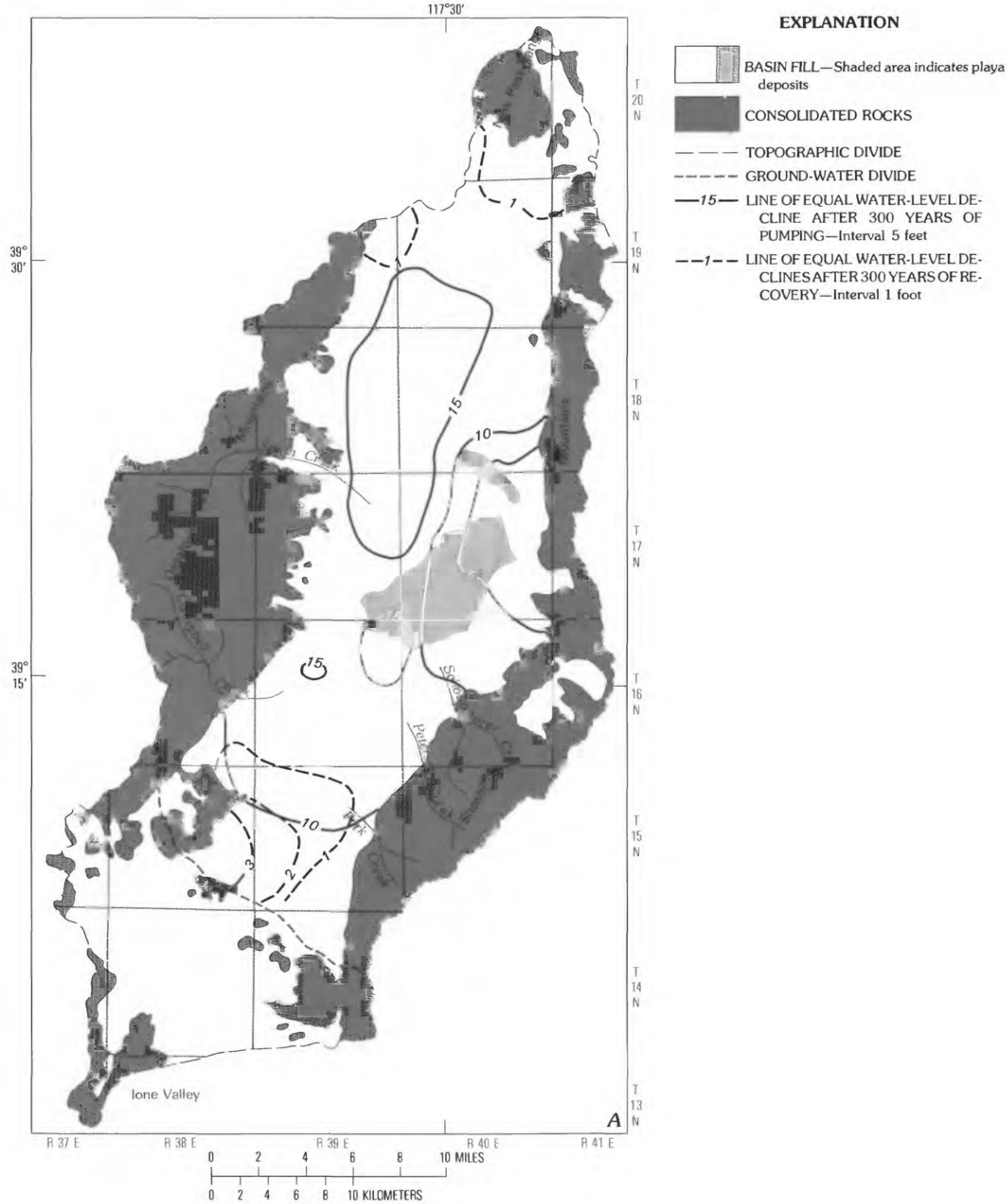


FIGURE 20 (above and facing page).—Results of simulated pumping and subsequent recovery for scenario A. Wells are distributed to most effectively reduce natural discharge, and are pumped at a combined rate equal to the sustained yield. *A*, Areal distribution of drawdown after 300 yr of pumping and after 300 yr of recovery.

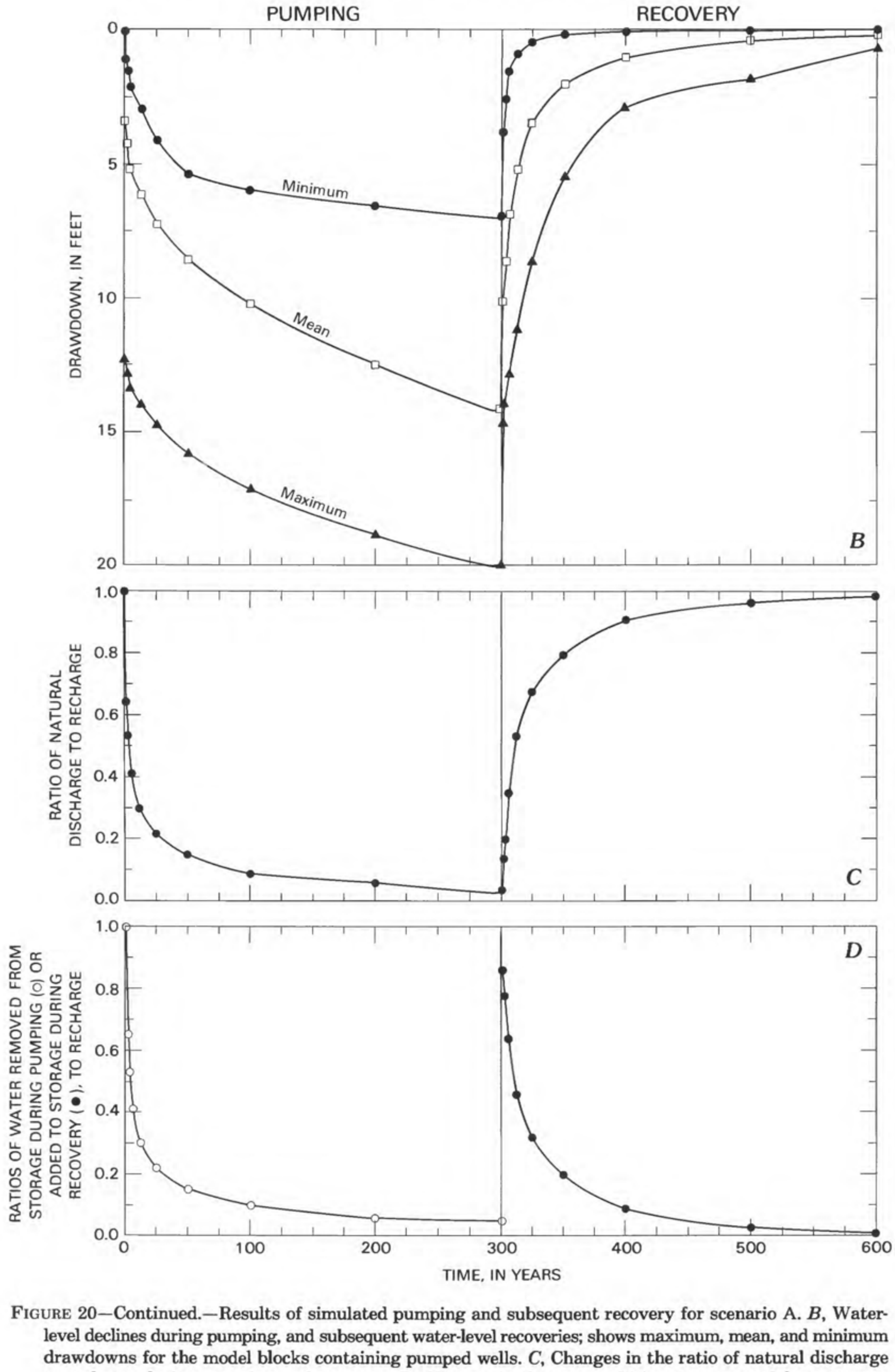


FIGURE 20—Continued.—Results of simulated pumping and subsequent recovery for scenario A. *B*, Water-level declines during pumping, and subsequent water-level recoveries; shows maximum, mean, and minimum drawdowns for the model blocks containing pumped wells. *C*, Changes in the ratio of natural discharge to recharge during pumping and recovery. *D*, Changes in the ratios of (1) water removed from storage during pumping to recharge and (2) water added to storage during recovery to recharge.

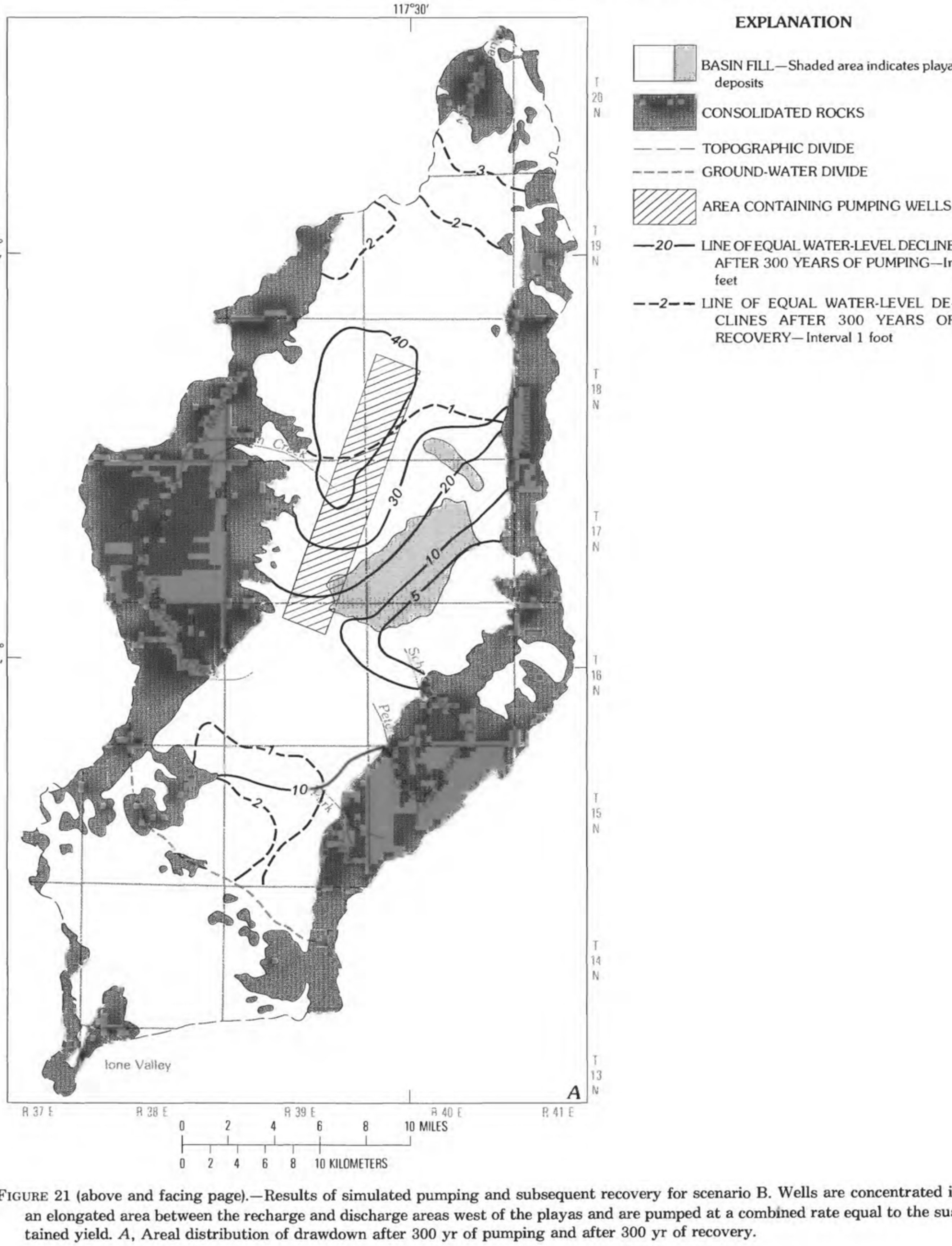


FIGURE 21 (above and facing page).—Results of simulated pumping and subsequent recovery for scenario B. Wells are concentrated in an elongated area between the recharge and discharge areas west of the playas and are pumped at a combined rate equal to the sustained yield. *A*, Areal distribution of drawdown after 300 yr of pumping and after 300 yr of recovery.

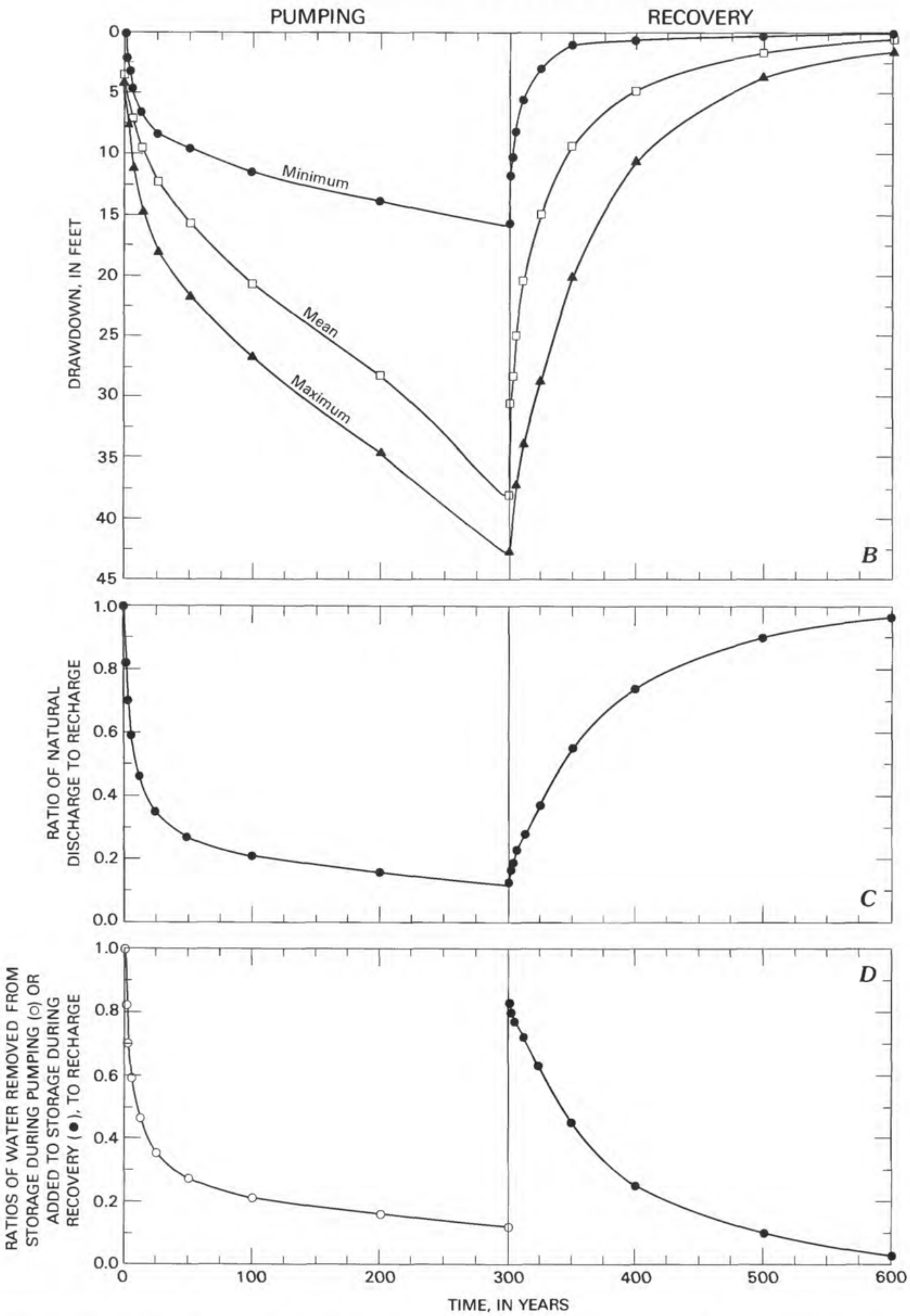


FIGURE 21—Continued.—Results of simulated pumping and subsequent recovery for scenario B. *B*, Water-level declines during pumping, and subsequent water-level recoveries; shows maximum, mean, and minimum drawdowns for the model blocks containing pumped wells. *C*, Changes in the ratio of natural discharge to recharge during pumping and recovery. *D*, Changes in the ratios of (1) water removed from storage during pumping to recharge and (2) water added to storage during recovery to recharge.

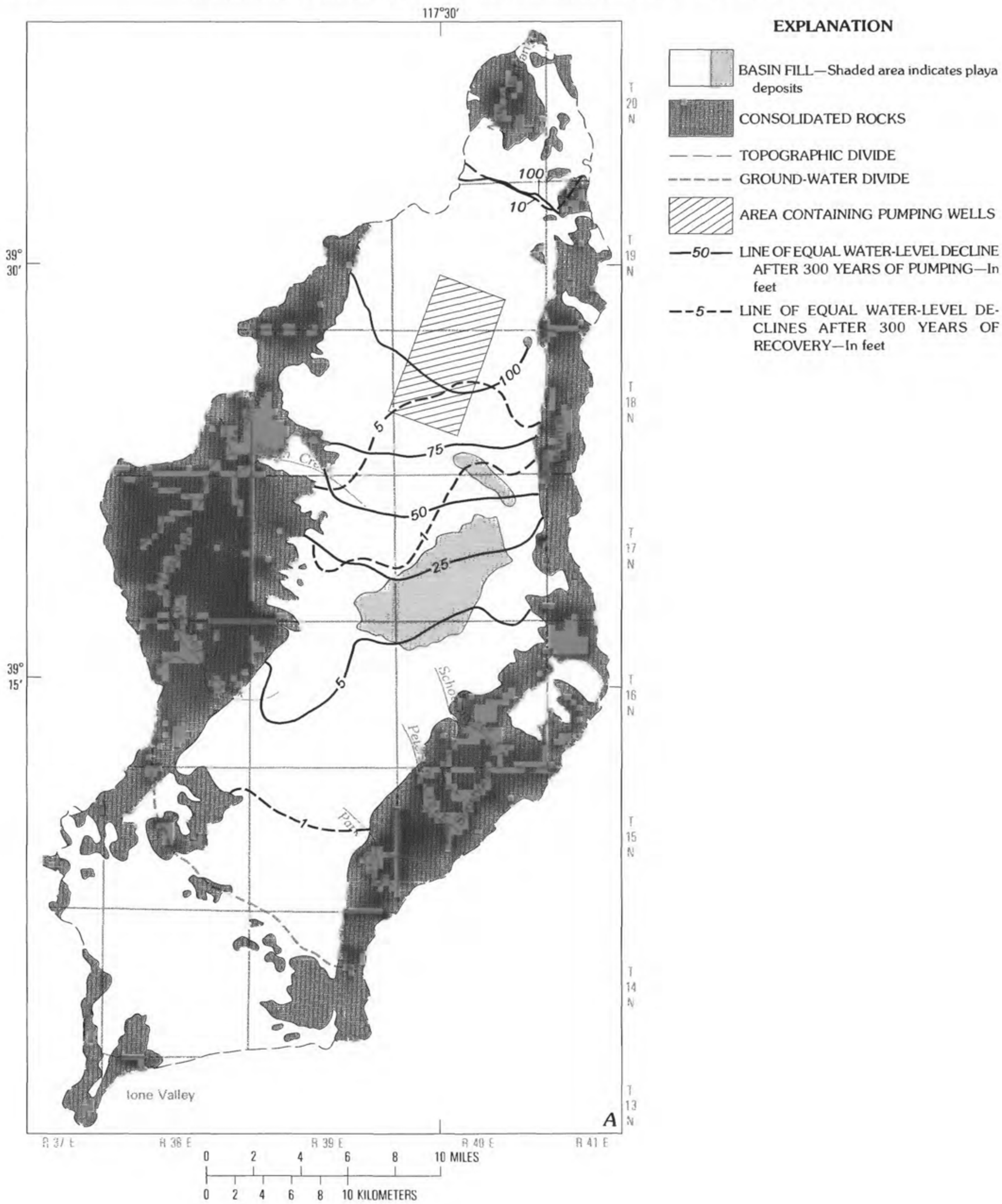


FIGURE 22 (above and facing page).—Results of simulated pumping and subsequent recovery for scenario C. Wells are concentrated in one area in the northern part of the basin, and are pumped at a combined rate equal to the sustained yield. *A*, Areal distribution of drawdown after 300 yr of pumping and after 300 yr of recovery.

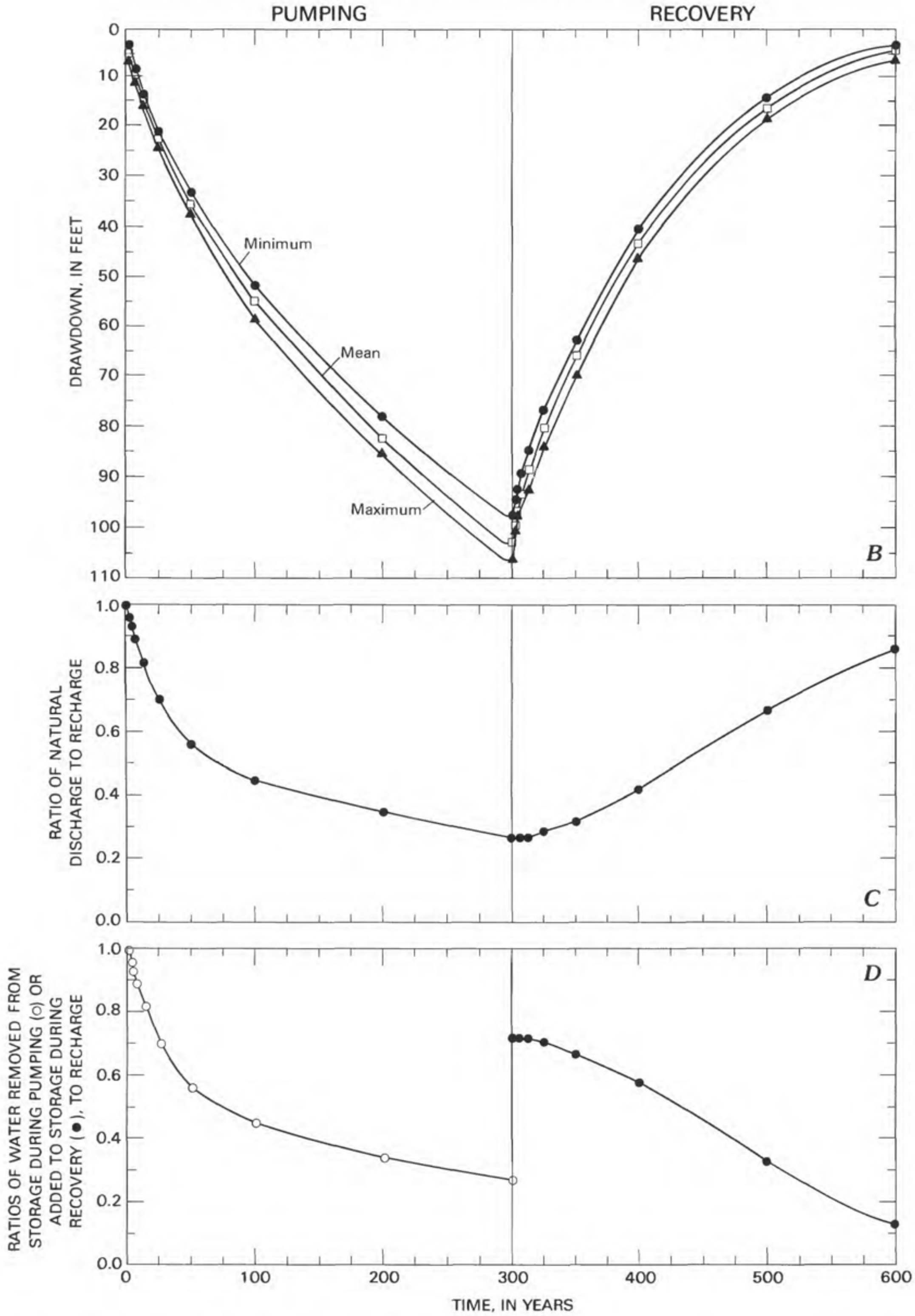


FIGURE 22—Continued.—Results of simulated pumping and subsequent recovery for scenario C. *B*, Water-level declines during pumping, and subsequent water-level recoveries; shows maximum, mean, and minimum drawdowns for the model blocks containing pumped wells. *C*, Changes in the ratio of natural discharge to recharge during pumping and recovery. *D*, Changes in the ratios of (1) water removed from storage during pumping to recharge and (2) water added to storage during recovery to recharge.

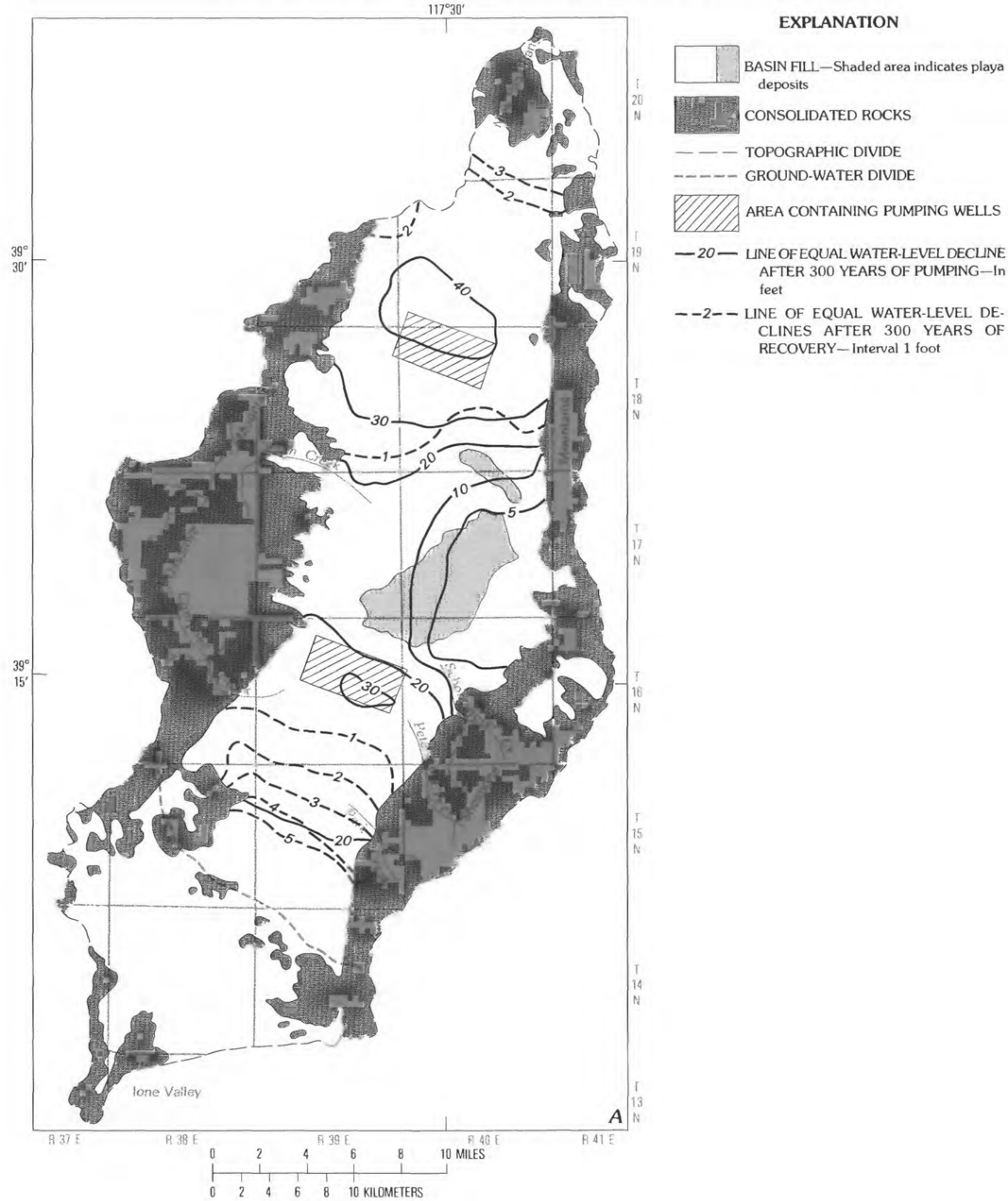


FIGURE 23 (above and facing page).—Results of simulated pumping and subsequent recovery for scenario D. Wells are concentrated in two areas near the natural-discharge area, and are pumped at a combined rate equal to the sustained yield. *A*, Areal distribution of drawdown after 300 yr of pumping and after 300 yr of recovery.

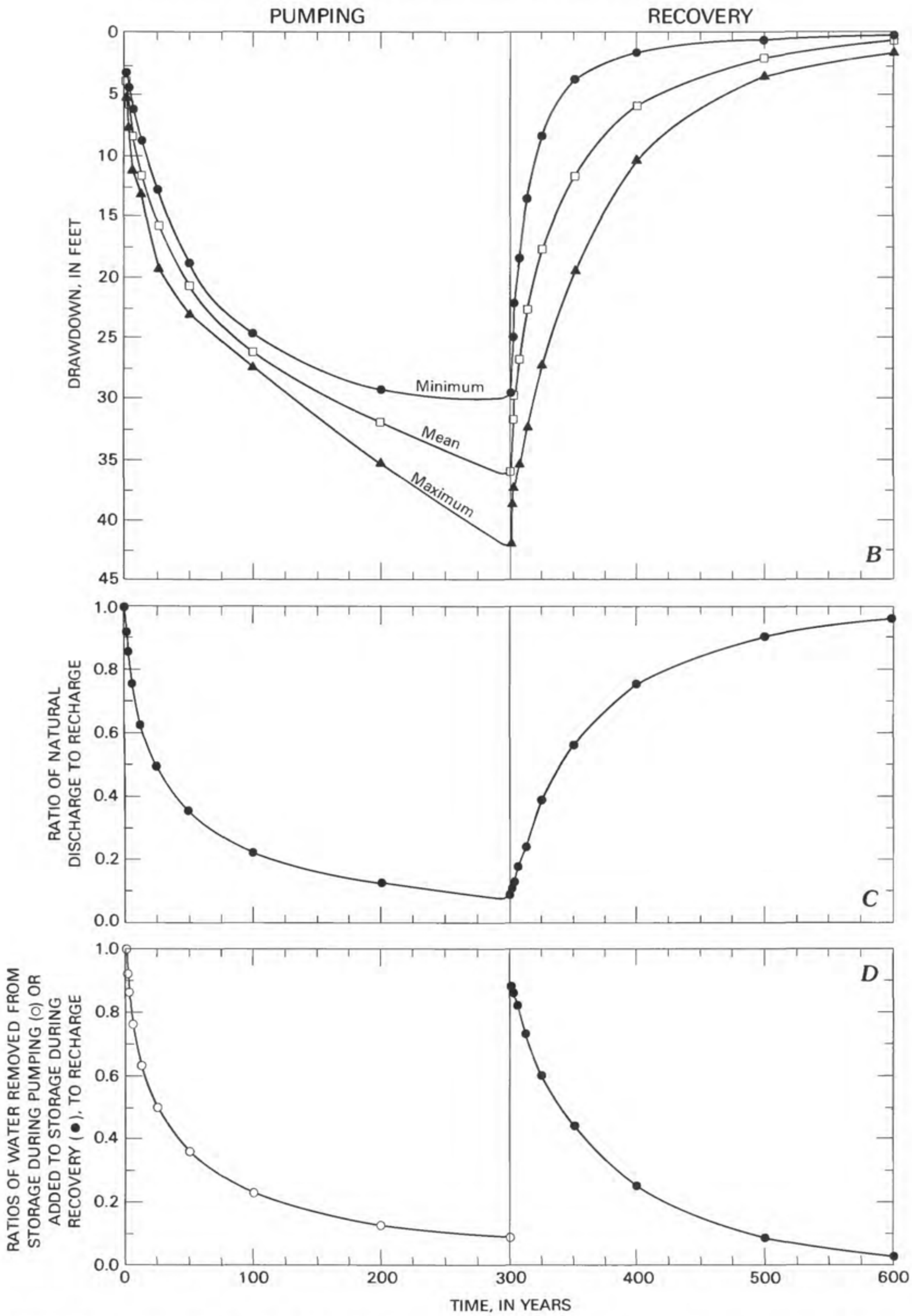


FIGURE 23—Continued.—Results of simulated pumping and subsequent recovery for scenario D. *B*, Water-level declines during pumping, and subsequent water-level recoveries; shows maximum, mean, and minimum drawdowns for the model blocks containing pumped wells. *C*, Changes in the ratio of natural discharge to recharge during pumping and recovery. *D*, Changes in the ratios of (1) water removed from storage during pumping to recharge and (2) water added to storage during recovery to recharge.

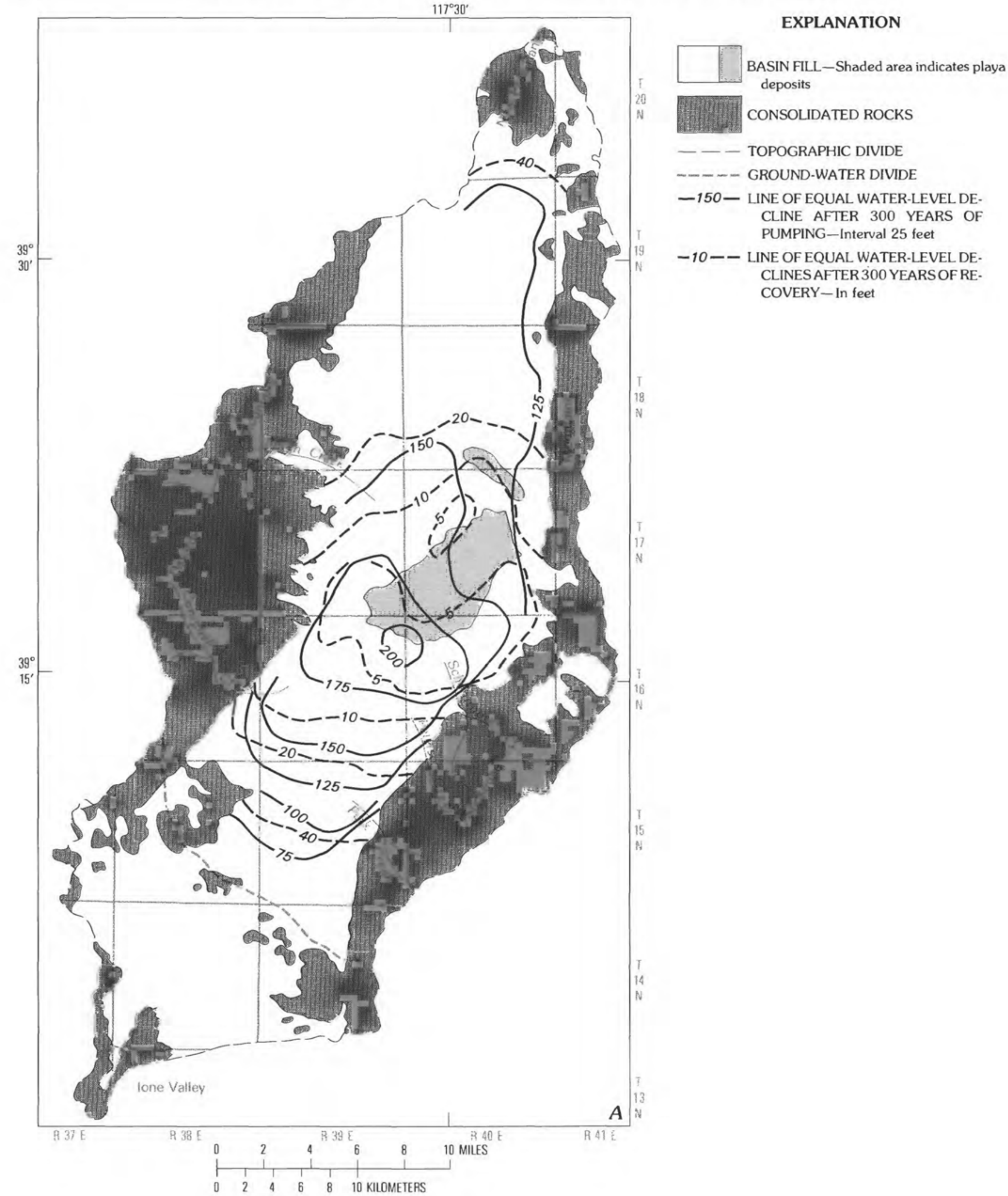


FIGURE 24 (above and facing page).—Results of simulated pumping and subsequent recovery for scenario E. Wells are distributed to most effectively reduce natural discharge, and are pumped at a combined rate equal to twice the sustained yield. *A*, Areal distribution of drawdown after 300 yr of pumping and after 300 yr of recovery.

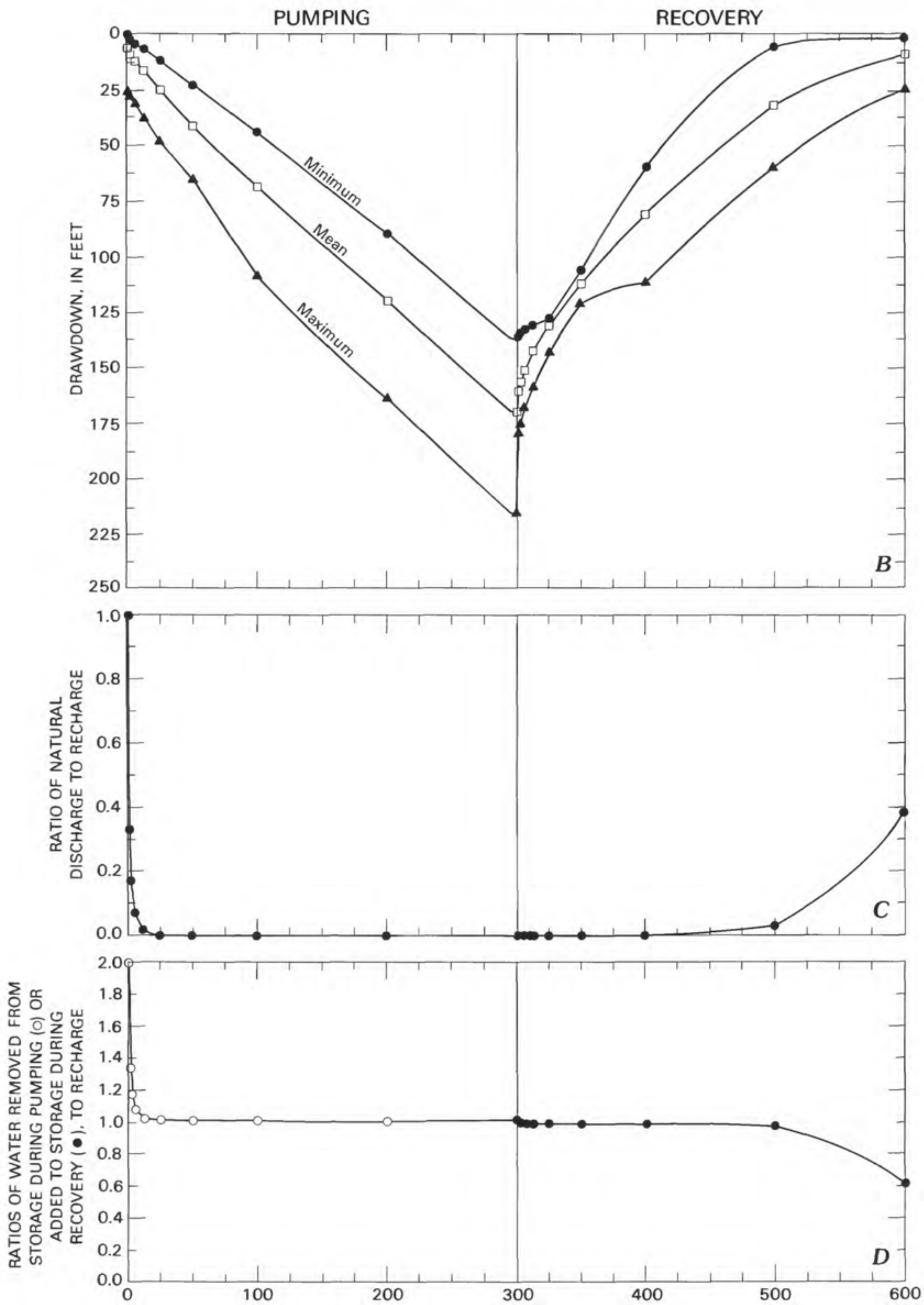


FIGURE 24—Continued.—Results of simulated pumping and subsequent recovery for scenario E. *B*, Water-level declines during pumping, and subsequent water-level recoveries; shows maximum, mean, and minimum drawdowns for the model blocks containing pumped wells. *C*, Changes in the ratio of natural discharge to recharge during pumping and recovery. *D*, Changes in the ratios of (1) water removed from storage during pumping to recharge and (2) water added to storage during recovery to recharge.

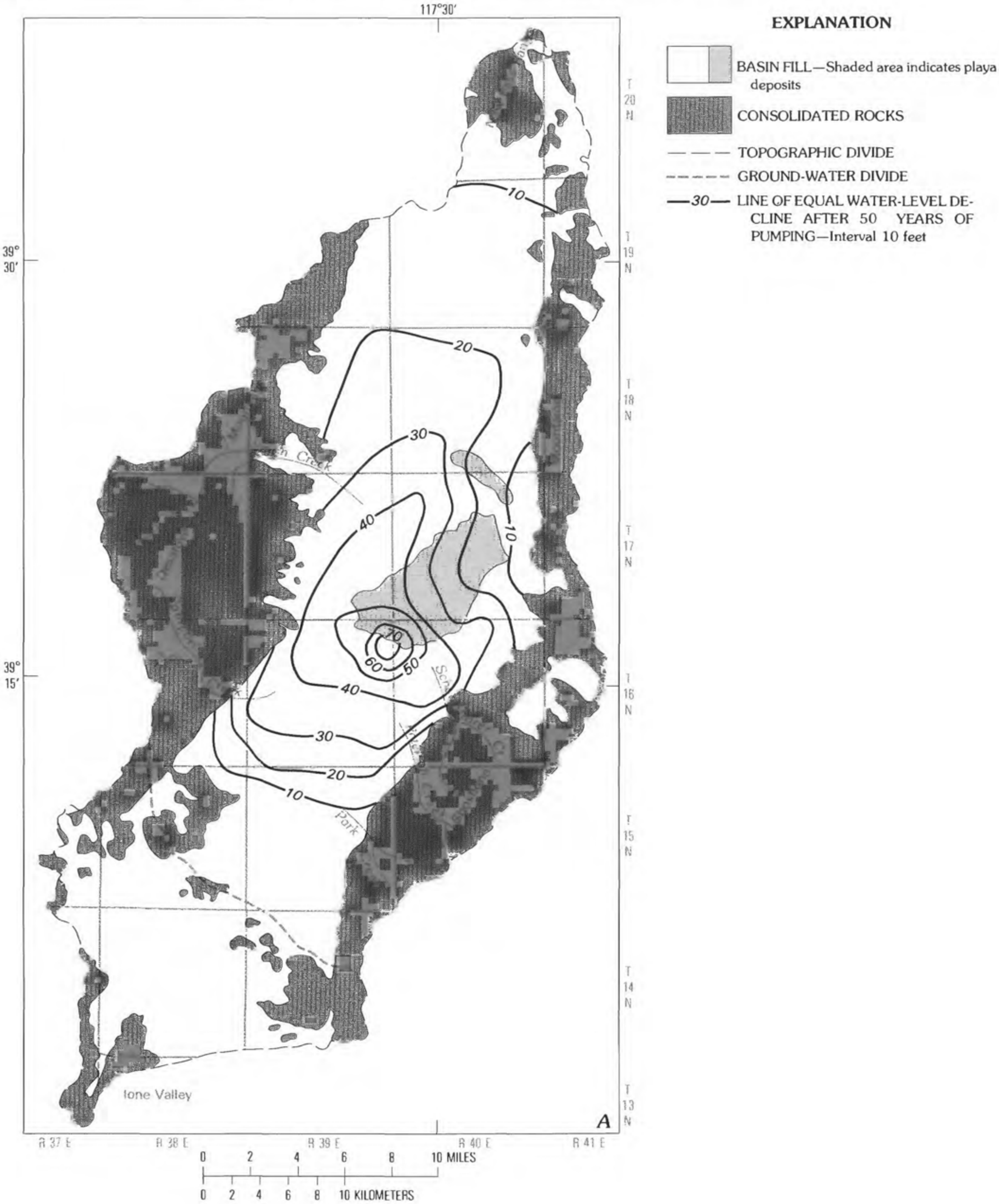


FIGURE 25 (above and following pages).—Results of simulated pumping and subsequent recovery for scenario F. Wells are distributed to most effectively reduce natural discharge, and are pumped at a combined rate equal to twice the sustained yield for 50 yr and then at a combined rate equal to the sustained yield for 250 yr. *A*, Areal distribution of drawdown after 50 yr of pumping.

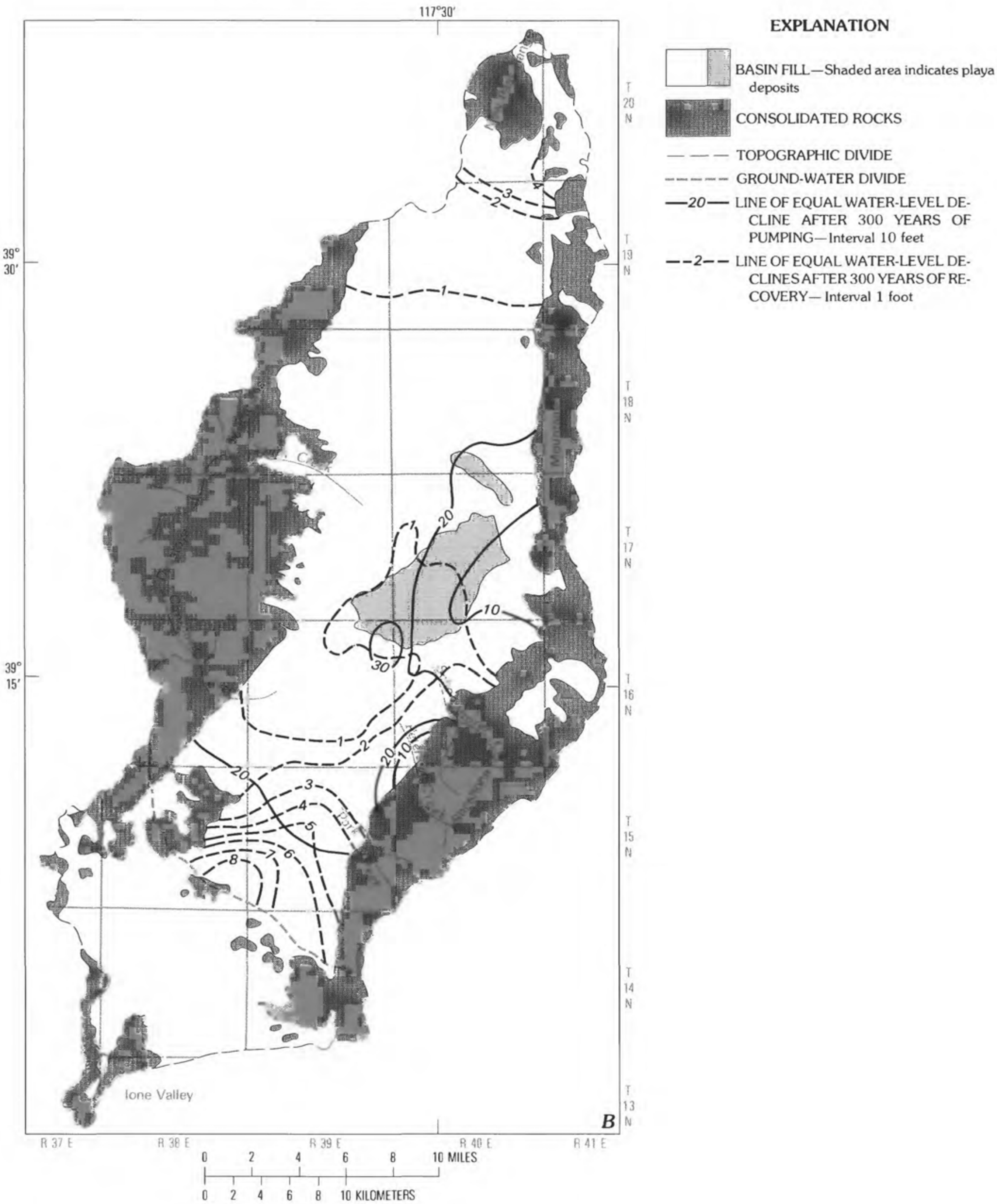


FIGURE 25—Continued.—Results of simulated pumping and subsequent recovery for scenario F. *B*, Areal distribution of drawdown after 300 yr of pumping and after 300 yr of recovery.

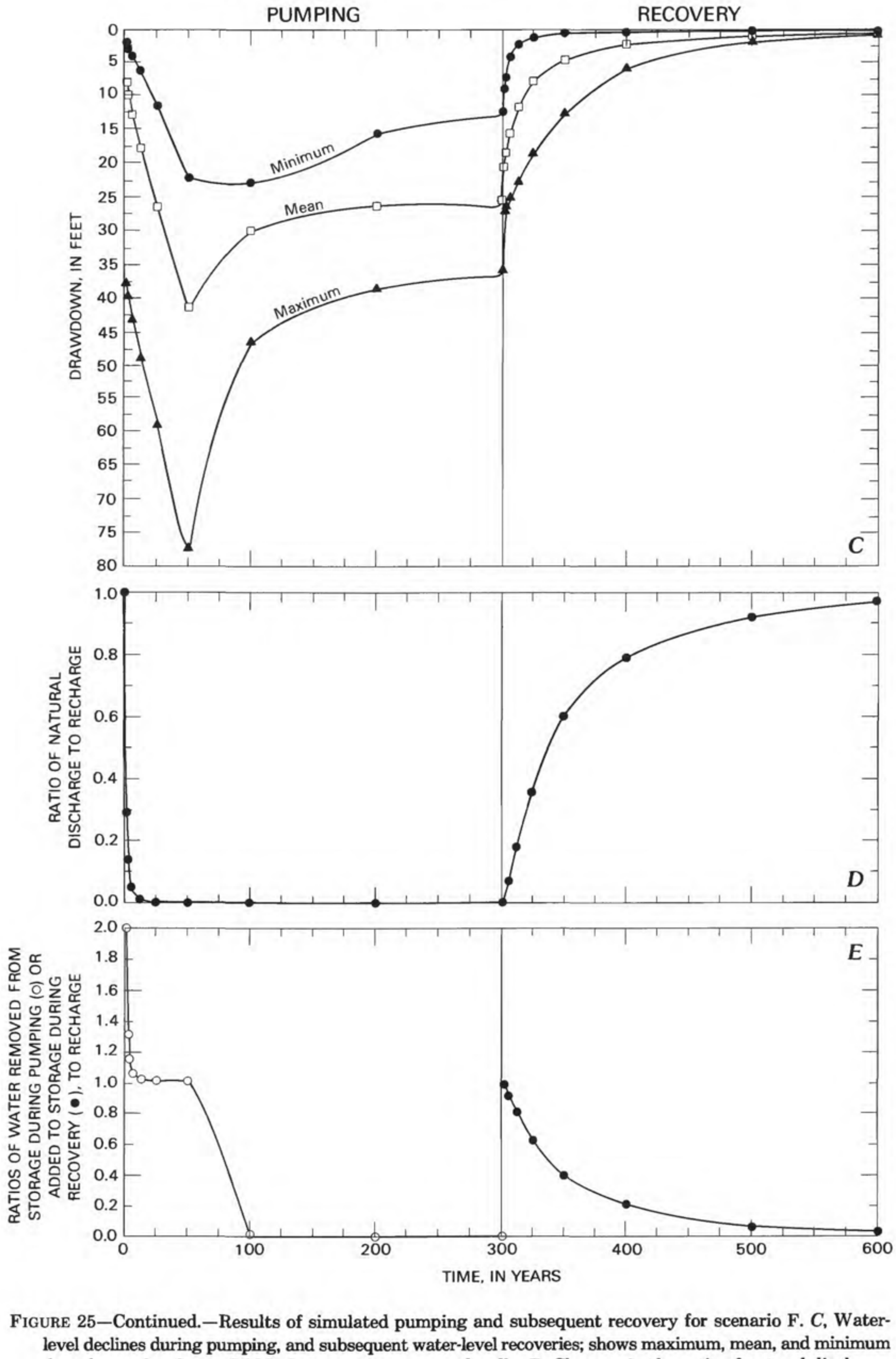


FIGURE 25—Continued.—Results of simulated pumping and subsequent recovery for scenario F. *C*, Water-level declines during pumping, and subsequent water-level recoveries; shows maximum, mean, and minimum drawdowns for the model blocks containing pumped wells. *D*, Changes in the ratio of natural discharge to recharge during pumping and recovery. *E*, Changes in the ratios of (1) water removed from storage during pumping to recharge and (2) water added to storage during recovery to recharge.

Pumping at the higher rate for 50 yr captured all ground water that would otherwise have been discharged by evapotranspiration in less than 25 yr, so the subsequent pumping at a reduced rate removed only an amount of ground water equal to recharge (fig. 25*D* and *E*). However, water-level declines in the valley during this simulation were greater than for scenario A in which ground water lost by evapotranspiration was captured more slowly. Consequently, resumption of natural discharge was slower for scenario F, because aquifer replenishment and water-level recovery in the discharge area required more time.

EVALUATION OF THE HYPOTHETICAL DEVELOPMENT SCENARIOS

Simulations of six development scenarios resulted in water-level declines and reduced natural evapotranspiration during pumping, as well as water-level recoveries and increased natural discharge approaching pre-pumping rates after pumping ceased. The extent of water-level declines and the rate of reduction of natural discharge depended on the areal distribution of well locations and rate of pumping. The subsequent recovery of water levels and natural discharge depended on the location of pumping, the degree of water-level declines in the discharge area, and the cumulative amount of water removed from storage. Pumping at greater rates, or in more concentrated areas, or in areas away from major sources of recharge or discharge produced greater drawdowns. Recovery of water levels and resumption of natural discharge were slower and not as complete for scenarios that (1) produced large cones of depression away from major recharge sources, (2) caused greater drawdowns in the natural-discharge area, or (3) removed greater volumes of water from storage.

Several pumping patterns were used in sustained-yield simulations to determine: (1) efficiency in capturing natural evapotranspiration, (2) effect on water-level declines, and (3) effects on recovery of water levels and natural discharge after pumping ceases. Natural discharge is probably reduced more efficiently by the model than it would be in an actual pumping situation because the simulated distribution of evapotranspiration did not include significant discharge from the playa (fig. 18). In addition, the possibility of continued discharge due to increased penetration by phreatophytic roots as water levels decline may not be totally accounted for by the model. Therefore, the simulated long-term trends are considered valid, but more time may be required in an actual pumping situation to obtain the amounts of reduction in the natural discharge that are indicated by the simulations.

Scenario A, with pumping strategically distributed to capture ground-water flow before it was naturally discharged, may eliminate as much as 96 percent of the evapotranspiration as the aquifer approaches a new equilibrium, with an average water-level decline in the pumping area of only about 14 ft. The sustained-yield concept of pumping an amount of ground water equal to recharge to reduce the natural discharge while producing only minor drawdowns probably is a viable development alternative. Recovery of water levels would occur rapidly when pumping ceased.

Scenarios B, C, and D produced greater water-level declines and were less effective than scenario A in capturing ground-water flow before it was naturally discharged. Distributing pumping in one (scenario B) or two (scenario D) concentrated areas near the natural-discharge zone would be the second best alternatives for reducing natural discharge while minimizing drawdown. Recovery of water levels after pumping ceased would also be faster and more complete for these alternatives than for scenario C. In contrast, concentrating pumping in one area away from the discharge zone (scenario C) produced the greatest water-level declines among all sustained-yield scenarios and was least effective in reducing natural discharge. Furthermore, recovery of water levels for this scenario was the slowest and least complete.

Doubling the pumping rate (scenario E) produced rapid water-level declines, even with the best well distribution as discussed in scenario A. The higher pumping rate would completely and more quickly capture ground water before it is discharged if reduction of discharge occurs as efficiently as indicated by the model. However, this would result in extensive water-level declines and depletion of storage. Recovery of water levels after pumping ceased would be slow. Even after 100 yr of recovery, the average water level in blocks containing pumping wells would still be about 90 ft below its pre-pumping level.

Varying the pumping rate (scenario F) from 16,000 acre-ft/yr down to 8,000 acre-ft/yr, after 50 yr of pumping, resulted in a rapid rise of water levels in the pumping area. Declines of water levels outside the pumping area were about 20 to 150 ft less than would have occurred if pumping had continued at the higher rate (16,000 acre-ft/yr) for the entire 300-yr period. A pumping rate that is initially higher than the rate of natural discharge has the advantage of capturing ground water discharged by evapotranspiration more rapidly than is the case for pumping equal to the discharge rate, but the increased pumping also removes more water from storage, which results in greater water-level declines. Therefore, the advantage of more quickly reducing evapotranspiration is offset by the disadvantages of greater water-level declines and removal of more water from storage. After 300 yr of dual-rate pumping, the average water-level decline in the pumped area is almost twice that for the constant rate which is equal to the discharge rate (scenario A). On the other hand, if the aquifer were pumped at a rate higher than the natural discharge rate, a subsequent reduction in the pumping rate to that of the natural discharge rate would result in (1) a decrease in the rate of water-level declines, and (2) a possible shallowing and broadening of the area of water-level declines. Eventually, the aquifer would approach a new equilibrium that would involve only a small average water-level decline. Recovery for this scenario would be slower and less complete than for the scenario of pumping a rate equal to the natural discharge rate (sustained yield), but faster and more complete than if pumping had continued at twice the rate of the natural discharge.

SUMMARY AND CONCLUSIONS

Hydrologic data were collected and phreatophytes mapped to define and quantify the hydrologic system of Smith Creek Valley. New and existing data were evaluated to determine the extent and thickness, hydraulic head distribution, and hydraulic properties of the basin-fill deposits, and to estimate rates of ground-water recharge and discharge in the valley.

Configuration of the basin-fill aquifer was determined by interpreting gravity data. Gravity interpretations depicted two basin-fill depressions underlying the valley—one about 3,000 ft thick, in the northern part of the valley, and the other about 5,500 ft thick, beneath the main playa in the central part of the valley. The thickness estimates obtained from gravity interpretations correlate reasonably well with thicknesses previously estimated from seismic and aeromagnetic data.

Horizontal hydraulic gradients are rather flat throughout most of the valley and are extremely flat—generally about 1 ft/mi—north of the playa. Vertical hydraulic gradients, produced by confinement in the playa sediments, can be as great as 0.20 ft/ft in the upper 75 ft of the sediments.

Ground water in the valley flows from alluvial fan deposits around the higher margin of the valley to the playa, which is topographically the lowest point in the valley. Flow is primarily horizontal, except in the discharge area where a strong upward vertical component exists. In the recharge areas around the margin of the valley, a downward component occurs, but little data are available to document the vertical gradient.

Hydraulic properties of the basin-fill deposits were estimated from lithologic logs from observation, irrigation, and stock wells; geothermal temperature-gradient holes; and aquifer test data from other valleys in the Great Basin. In addition, the average hydraulic conductivity of the fine-grained playa deposits was determined from one falling-head permeameter test performed in the laboratory and calculated from existing laboratory determinations of moisture retention, grain size, and porosity. An estimate of vertical hydraulic conductivity for the fine-grained deposits was also made from temperature profiles measured in a well. Model simulation suggests that average hydraulic conductivities for the different types of basin-fill deposits vary by approximately four orders of magnitude, from 1×10^{-7} ft/s for fine-grained playa deposits to 2.5×10^{-4} ft/s for coarse-grained deposits. Specific yields of basin-fill deposits were estimated to range from 6 percent for fine-grained playa deposits to 15 percent for coarse-grained deposits based on other ground-water flow modeling studies in the Great Basin (Harrill, 1982; Mower, 1982; and David S. Morgan, U.S. Geological Survey, written commun., 1983), a compilation of specific yields for various materials (Johnson, 1967), and a laboratory measurement of specific yield for a fine-grained soil sample (table 5). Storage coefficients were estimated to range from 7.0×10^{-5} for a 75-ft-thick interval of the confined aquifer to 0.15 for parts of the water-table aquifer, based on specific yields of the water-table aquifer and calculated from model-layer thicknesses by using assumed specific storage values (Ireland and others, 1982, p. 28–35) for the confined intervals.

Recharge estimates range from 8,300 to 9,600 acre-ft/yr, which are about 4,000 to 2,000 acre-ft/yr less than the previous reconnaissance estimate. Natural discharge is estimated at 8,300 acre-ft/yr, on the basis of detailed phreatophyte mapping. This estimate is about 2,000 acre-ft/yr higher than the previous reconnaissance estimate. Total discharge from the basin-fill aquifer is approximately 9,000 acre-ft/yr, including about 650 acre-ft/yr of pumpage. Annual rates of discharge are probably more reliable estimates than rates of recharge because discharge is derived from field observations rather than from an empirical relationship between altitude and precipitation as used in the two methods of estimating recharge.

All of the data were used in a ground-water flow model to (1) help define the hydrologic system, (2) evaluate hydraulic properties, and (3) simulate the long-term effects of the hypothetical development scenarios on the aquifer system. The simulations showed that water levels are controlled primarily by evapotranspiration in the discharge area. An increase in the evapotranspiration rate would cause water levels to decline in the discharge area and therefore would result in lower water levels throughout the valley. On the other hand, a decrease in evapotranspiration rates would cause water levels to rise in the discharge area, which would result in higher water levels throughout the valley. Raising or lowering the depth at which evapotranspiration ceases (extinction depth) in the flow model would have a similar effect on water levels.

Grain-size, which controls hydraulic conductivity, and recharge distributions also affect hydraulic gradients in the basin-fill aquifer. Parts of the aquifer outside the discharge area that contain finer-grained deposits or receive major amounts of recharge have steeper hydraulic gradients than parts of the aquifer having coarse-grained deposits or receiving small amounts of recharge.

Calibration of the model involved adjustment of initial estimates of hydraulic properties and boundary conditions, within reasonable limits, until an acceptable steady-state simulation was attained.

After the model was calibrated to predevelopment steady-state conditions, several hypothetical development scenarios were simulated, for 300 yr of pumping and 300 yr of recovery, to evaluate the effects on the ground-water system caused by (1) pumping from the aquifer at a rate approximately equal to the estimated recharge rate (8,000 acre-ft/yr; scenarios A–D), (2) pumping at a rate greater than the recharge rate (scenario E), and (3) pumping 50 yr at twice the rate and the remaining 250 yr at the recharge rate (8,000 acre-ft/yr; scenario F). The hypothetical pumping wells were not located (1) in areas of fine-grained deposits having low transmissivity, (2) within a mile of saline water, (3) where the depth to water exceeds 200 ft, (4) where the thickness of saturated basin fill is less than 200 ft, (5) where model blocks are bounded on two or more sides by consolidated rocks, and (6) where land-surface slopes are greater than 200 ft/mi.

The concept of sustained yield was evaluated by simulating pumping at the estimated rate of recharge, with several alternative areal distributions of pumping. Initially, a pumping distribution that would efficiently capture ground water before it is naturally discharged was simulated to determine whether the ground water lost to natural discharge could be captured while producing only minor drawdowns (scenario A). Simulation of this scenario resulted in rapid reduction of natural discharge and a maximum drawdown of 20 ft (averaged over a 2 mi^2 model block) as the aquifer approached a new equilibrium. Concentrating the same amount of pumpage in one (scenario B) or two (scenario D) areas near the discharge zone was almost as efficient in reducing natural evapotranspiration as scenario A, but it created average block-wide drawdowns exceeding 40 ft for the pumping period. Finally, placing a pumping center in the northern part of the valley (scenario C) resulted in significantly slower and less complete reduction of natural evapotranspiration than for all the previous scenarios (A, B, and D), and it created a large storage depletion and average block-wide water-level declines of greater than 100 ft in the pumped area.

When the pumping rate was doubled (twice the estimated recharge rate) and the well distribution was the same as in scenario A, ground water discharged by natural evapotranspiration was completely captured in less than 25 yr. After 300 yr of pumping, however, average block-wide water-levels in the pumped area had declined more than 200 ft and were still declining at approximately the same rate as when pumping started (scenario E).

When pumping was set at twice the rate of the estimated recharge for 50 yr and then reduced to the rate of estimated recharge for the next 250 yr, with the same well distribution as in scenario A (scenario F), ground water discharged by natural evapotranspiration was completely captured in less than 25 yr. When pumping was reduced after the first 50 yr, water levels in the pumping area rose. Maximum water-level declines, averaged over each model block, were about 35 ft after 300 yr of pumping, which is about a sixth of what declines were when pumpage was maintained at the higher rate for the entire 300-yr pumping period.

All six development scenarios resulted in water-level declines and reduced natural evapotranspiration during pumping. All six showed water-level recoveries and increased natural discharge approaching pre-pumping rates after pumping ceased. However, the extent of water-level declines and the rate of reduction of natural discharge depended on the areal distribution of wells and rate of pumping. The subsequent recovery depended on the location of wells, the extent of water-level declines in the discharge area, and the cumulative amount of water removed from storage. Pumping at greater rates, or in more concentrated areas, or in areas away from major sources of recharge produced greater drawdowns. Recovery of water levels and resumption of natural discharge were slower and not as complete for development scenarios that (1) produced large areas of water-level declines away from major recharge sources, (2) caused greater water-level declines in the natural-discharge area, or (3) removed greater volumes of water from storage.

Simulations for Smith Creek Valley may be representative of conditions in other hydrologically closed single-valley flow systems throughout the Great Basin. Therefore, most hydrologically closed valleys in the Great Basin probably can be evaluated as basins filled with saturated fine- to coarse-grained deposits surrounded and underlain by consolidated rocks of low permeability, with water recharging the basin-fill aquifer around its perimeter and discharging in the topographically low points. An additional characteristic of hydrologically closed valleys is a general decrease in grain size away from the mountain blocks toward the topographically lowest part of the valley, where the sediments commonly are fine-grained playa deposits. Water levels and hydraulic gradients in these aquifers are controlled by:

(1) Water levels in the discharge area, because the valley is hydrologically closed and all water recharging aquifers in the valley is discharged in the low-lying part;
(2) Major recharge areas, because large volumes of water recharging the aquifer would produce proportionately higher water levels in the recharge areas of the valley; and
(3) Grain-size distributions, because finer grained sediments outside the discharge area would produce steeper hydraulic gradients than coarse-grained sediments. Furthermore, pumping of these aquifers should produce generally flat and extensive water-level declines in a manner similar to that shown in figures 20*A*, 21*A*, 22*A*, and 23*A*.

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