GROUND-WATER FLOW AND SIMULATED EFFECTS OF DEVELOPMENT IN STAGECOACH VALLEY, A SMALL, PARTLY DRAINED BASIN IN LYON AND STOREY COUNTIES, WESTERN NEVADA

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1409-H

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Ground-Water Flow and Simulated Effects of Development in Stagecoach Valley, a Small, Partly Drained Basin in Lyon and Storey Counties, Western Nevada

By JAMES R. HARRILL *and* ALAN M. PREISSLER

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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.

Gordon P. Eaton

Gordon P. Eaton
Director

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Multiply	By	To obtain
acre	4,047	square meter
acre-foot (acre-ft)	1,233	cubic meter
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
cubic foot per second (ft^3/s)	28.32	liter per second
foot (ft)	0.3048	meter
foot per second (ft/s)	0.3048	meter per second
square foot per day (ft^2/d)	0.0929	square meter per day
gallon (gal)	3.785	liter
inch (in.)	25.4	millimeter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.59	square kilometer

For temperature, degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) by using the formula °C=0.5556(°F–32).

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

GROUND-WATER FLOW AND SIMULATED EFFECTS OF DEVELOPMENT IN STAGECOACH VALLEY, A SMALL, PARTLY DRAINED BASIN IN LYON AND STOREY COUNTIES, WESTERN NEVADA

By JAMES R. HARRILL and ALAN M. PREISSLER

ABSTRACT

Stagecoach Valley is a small, topographically closed basin in western Nevada with a total area of about 70 square miles. Local hydraulic continuity exists between Stagecoach Valley and the adjacent Carson River to the south and beneath parts of the northeastern area adjacent to Churchill Valley. Most of the locally derived runoff and recharge is generated in the Flowery Range on the north side of Stagecoach Valley. The basin fill is at least 500 feet thick throughout most of Stagecoach Valley and has a maximum thickness of about 3,000 feet.

A ground-water flow model was used to simulate the hydrology of Stagecoach Valley. It is estimated that about 1 million acre-feet of water is stored in the basin-fill aquifer of the Stagecoach Valley and that under predevelopment conditions the annual flow through the basin-fill aquifer underlying the valley was about 920 acre-feet. On the basis of the simulation, the flow components are (1) total inflow that included about 550 acre-feet per year of recharge from local precipitation, about 280 acre-feet per year of subsurface inflow from the upstream reach of the Carson River, and about 90 acre-feet per year of subsurface inflow from the downstream reach of the Carson River and (2) total outflow that included about 630 acre-feet per year by evapotranspiration, about 170 acre-feet per year by subsurface flow to Churchill Valley, and about 120 acre-feet per year by subsurface outflow to the downstream reach of the Carson River.

During the 11 pumping seasons 1971 through 1981, slightly more than 11,000 acre-feet of water was pumped from the basin-fill aquifer. Of this, slightly less than 1,000 acre-feet of the water pumped for irrigation was recirculated back to the aquifer, resulting in a net pumping draft of about 10,000 acre-feet. About 3,000 acre-feet was supplied by reductions in evapotranspiration and by changes in subsurface inflow and outflow; the remaining 7,000 acre-feet of the pumpage was derived from ground-water storage. Water-level declines throughout the basin-fill aquifer ranged from 1 foot or less near the Carson River to more than 15 feet in the developed area. Changes in subsurface inflow and outflow primarily involved inducing additional inflow from the Carson River.

The probable response to long-term pumping stress was evaluated by simulating nine hypothetical development scenarios. The results suggest that the sustained-yield concept of managing basin development is viable for Stagecoach Valley. This concept involves regulating pumpage so that over a long term withdrawals do not exceed the amount of natural discharge plus the additional induced recharge. In all scenarios, considerable additional subsurface inflow was induced from the Carson River. This induced flow in turn suggests that the basin-fill aquifer of the valley is capable of attaining a new equilibrium in response to pumping rates far in excess of the natural (predevelopment) inflow rate of 920 acre-feet per year. Consequently, the predevelopment or natural flow through the aquifer is not considered the best criterion to be used in determining a sustained pumping rate in Stagecoach Valley. The limiting factor probably is the degree to which decreases of Carson River flows can be tolerated. The ground-water flow system is sensitive to variations in pumping rates and to the location of pumping; pumping the central and northern parts of the valley has the least effect on Carson River flows. Water-level changes alone are not adequate criteria for evaluating the aquifer's response; changes in subsurface inflow and outflow must also be considered. Because of the area's small size, the proximity of the aquifer boundaries strongly affects the response to any pumping stress.

INTRODUCTION

The hydrologic study of Stagecoach Valley was a part of the Great Basin Regional Aquifer-System Analysis (RASA) conducted by the U.S. Geological Survey. As discussed in the "Foreword," the RASA program is a study of ground-water systems at a large scale and is designed to systematically evaluate the major aquifer systems in the United States. The Great Basin area of Nevada, Utah, and adjacent States is considered to contain a regional aquifer system because the numerous individual basins within the area share many common characteristics and can be studied collectively. Currently about 240 hydrographic areas (valleys that contain one or more structural basins) have been recognized within the study area of the Great Basin RASA (Harrill and others, 1983, p. 5). Detailed studies of all 240 areas were precluded because of limitations in time and resources. Consequently, a major problem in planning the study was allocating the available resources in a manner most likely to produce information with significant transfer value. The approach taken was to study areas that have conditions typical of other areas that would not be studied. Eight basins, which collectively represent most hydrologic conditions present in the Great Basin, were selected for study by use of ground-water flow modeling techniques.

Stagecoach Valley was selected because it is a small arid basin that is topographically closed to surface drainage yet at the same time is partly drained by subsurface flow. The boundary conditions are complex and appear to have a strong influence on the hydrologic regime of the area. Knowledge developed about the influence of the boundary conditions on the hydrologic regime and on the general response to pumping stresses should be applicable to other small arid basins. Also, detailed information about the boundary conditions of the basin-fill aquifer may have significant transfer value for parts of larger basins.

PURPOSE

The primary purpose of this study was to gain insight into processes affecting ground-water flow in small, arid alluvial basins. The specific objectives of the study were to describe the basin-fill aquifers in Stagecoach Valley quantitatively, configure and calibrate a ground-water flow model to simulate nine pumping scenarios, and present the results of the model simulations in general terms that may be compared with other areas modeled as a part of the Great Basin RASA study. This report presents the results of the study and also evaluates the applicability of the sustained-yield concept of management to this type of area. This concept involves regulating pumpage so that, over a long term, pumping rates do not exceed the amount of natural discharge that can be captured by pumping plus any additional recharge that is induced as a result of pumping.

LOCATION AND GENERAL FEATURES OF THE STUDY AREA

Stagecoach Valley is in western Nevada about 20 mi east of Carson City (fig. 1). The general study area is bounded on the north by the Flowery Range, on the east by Churchill Valley and Churchill Butte, on the west by the Carson Plains part of Dayton Valley, and on the south by the Carson River. Stagecoach Valley is topographically closed and has a drainage area of about 70 mi^2, about 33 mi^2 of which is underlain by basin-fill deposits. The study area includes Stagecoach Valley and adjacent parts of both Churchill Valley and the Carson River, which have some degree of hydrologic continuity with Stagecoach Valley. The highest mountains in the study area are in the Flowery Range and have a maximum altitude of 7,095 ft above sea level. Mountains bordering the east side of the area have altitudes of 5,812 ft or less, and those bordering the southwest margin of the valley have altitudes of 5,221 ft or less. Most surficial drainage is to a playa, Misfits Flat (altitude about 4,260 ft), in the southeastern part of the valley (fig. 1). Sand dunes west of Misfits Flat cause intermittent ponding to occur where a few minor streams drain to a small area of alkali soil and sparse vegetation on the southwestern part of the valley floor. Vegetation in the remainder of the valley is sparse, especially on the valley floor where sage and shadscale predominate; piñon pine and juniper are present at higher altitudes at the north end of the area. Stagecoach Valley was virtually undeveloped before 1971.

APPROACH AND METHODS

Fieldwork began in the spring of 1982 and was completed by fall 1983. It consisted primarily of cataloging and measuring water levels in about 60 wells; surveying altitudes of most of these wells; assembling and interpreting existing hydrologic and geologic information; mapping the geology and hydrologic features of selected areas; inventorying pumpage based on house counts and areas of lawns and irrigated cropland; collecting 24 water samples for chemical analysis; and performing geophysical surveys, including gravity readings at 100 stations and two seismic profiles. Quantitative estimates of aquifer geometry were based on the analysis of geophysical information and data reported by Schaefer and others (1986) and Schaefer (1988). Estimates of other aquifer properties were based on analysis of geologic and hydrologic information reported in well logs and on field observations. Harrill and others (1984) did a preliminary analysis of the geologic controls on ground-water flow and later analyzed water samples to obtain geochemical information to verify patterns of ground-water flow (Harrill and others, 1993).

A multilayered ground-water flow model was formulated on the basis of available information and an analysis of the hydraulic properties of aquifer materials. One pumping test and 11 specific-capacity values from drillers' logs were used to

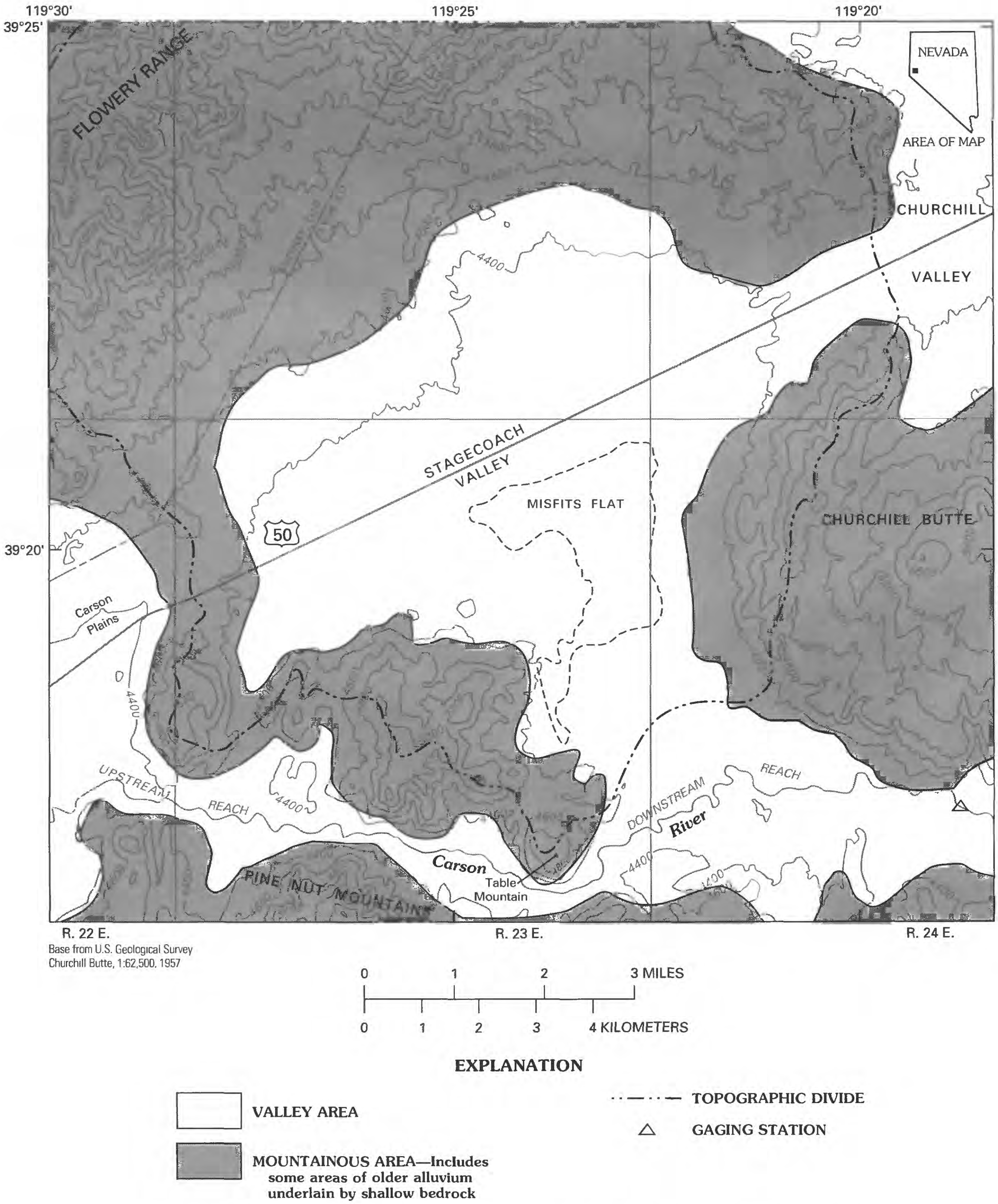


FIGURE 1.—Location and general features of Stagecoach Valley. Base contour interval 200 feet.

estimate the hydraulic properties of the aquifer materials. The model was calibrated against predevelopment heads and observed changes in head during the 11 pumping seasons from 1971 to 1981. The calibrated model was used to simulate the generalized long-term response to nine selected development scenarios.

The water-level data and well information are from the data base of the ground-water site-inventory files of the U.S. Geological Survey's National Water Data Storage and Retrieval System (WATSTORE). This information may be obtained through the U.S. Geological Survey office in Carson City, Nev., or through any designated National Water Data Exchange (NAWDEX) assistance center (Edwards, 1987).

ACKNOWLEDGMENTS

We thank the residents of Stagecoach Valley for their cooperation in supplying data and permitting access to their wells. Information on the Stagecoach public water-supply system was furnished by Oley Lawson of Stagecoach Utilities, Inc. The Utah Mining and Construction Company provided detailed information pertaining to the drilling and testing of several exploratory wells. Drillers' logs were obtained from the Nevada Division of Water Resources.

GEOLOGIC SETTING

GEOLOGIC UNITS

The 11 principal geologic units in Stagecoach Valley are grouped into two general types: (1) unconsolidated and partly consolidated basin-fill clastic deposits and (2) consolidated rock that composes the mountains and underlies the basin fill. The first group consists of mostly highly porous materials such as sand and gravel, which in general readily transmit water, whereas the consolidated rock commonly has low porosity and permeability and does not readily transmit water except where highly fractured. The 11 units are described in terms of their geologic characteristics and water-bearing properties in table 1, and the generalized geology of the study area is shown in figure 2.

STRUCTURAL FEATURES

The structural features that appear to have most strongly influenced the geometry of the basin-fill aquifer in Stagecoach Valley are the Carson lineament and basin-and-range-type extensional faulting.

The Carson lineament is one of several northeast-trending structural zones with left-lateral strike-slip movement that are present in the western Basin and Range province. They were described by Slemmons and others (1979) as occurring in that part of the Basin and Range province between the eastern fault escarpment of the Sierra Nevada, about 25 mi to the west of the study area, and the Walker Lane, about 20 mi to the east. The Carson lineament extends northeastward from Carson City for about 94 mi (Rogers, 1975) and passes through Stagecoach Valley (fig. 2). Shawe (1965) described this feature as a left-lateral complement to the Walker Lane, which is a northwest-trending structural feature with a right-lateral component of movement. The low, broad divides separating Stagecoach Valley from the adjacent basins to the east and west are geomorphic expressions of the Carson lineament in the study area. No evidence of recent movement has been observed in Stagecoach Valley, but near Carson City, about 20 mi to the southwest, Rogers (1975) showed Holocene left-lateral, oblique-slip movement on faults cutting the Quaternary basin fill. The structural basin that underlies Stagecoach Valley appears to be elongate parallel to the Carson lineament, and fracturing possibly resulting from movement along the lineament might provide conduits for leakage of water from within the topographic boundaries of the area.

Basin-and-range faulting oriented generally north-south also has had a significant influence on the geometry of the basin underlying Stagecoach Valley. Churchill Butte, which is on the east side of the area, was described by Moore (1969, p. 22) as the exposed part of a westward-tilted fault block. The western part of this tilted block appears to extend beneath Stagecoach Valley and to form a north-south-trending structural trough beneath Misfits Flat and that part of the area northeast of Table Mountain. Basin-fill deposits in this trough allow for hydraulic continuity between the flood-plain deposits of the Carson River and Stagecoach Valley. Relatively intensive normal faulting along the southwest boundary of Stagecoach Valley may have also provided for hydraulic continuity between flood-plain deposits of the Carson River and Stagecoach Valley by producing fractures in andesitic rocks and by tilting parts of a relatively permeable basalt flow to a position below the water table (fig. 2, cross section).

GEOLOGIC HISTORY

The pre-Tertiary event that most significantly influenced the present hydrologic regime was the

TABLE 1.—*Geologic units, their general hydrologic properties, and representative parts of ground-water flow model*

[Composite descriptions are based on Thompson (1956), Bonham (1969), Moore (1969), and Rose (1969) and on drillers' logs and field observations. —, locally unknown]

Geologic unit	Age	Thickness (feet)	Geologic description	Occurrence, general hydrologic properties, and representative parts of ground-water flow model
			Cenozoic basin-fill deposits	
Playa deposits	Holocene and Pleistocene.	0–40	Unconsolidated clay, silt, and fine sand.	Occurs on and adjacent to Misfits Flat. Deposits have high porosity and low hydraulic conductivity; generally function as confining beds and do not readily yield water to wells. Represented in ground-water flow model as active cells in top layer (layer 1) with low hydraulic conductivity.
Flood-plain deposits.	.....do	0–30	Unconsolidated sand, gravel, and some silt.	Occurs as terrace and channel deposits along Carson River flood plain. Contains highly permeable sand and gravel deposits that are aquifers and readily yield water to wells. Represented in ground-water flow model as active cells in top layer (layer 1) that have high hydraulic conductivity and high specific yield. Flow to and from Carson River is simulated from flood-plain deposits in model.
Younger alluvium.	.....do	0–250	Unconsolidated sand, gravel, silt, and clay. Alluvial-fan and colluvial basin-fill deposits. Sand and gravel deposits interfingered with silt and clay toward center of basin.	Occurs on valley floor and as alluvial fans at margins of valley. Lenses of sand and gravel and reworked shoreline lacustrine deposits yield water readily to wells and are most productive aquifers in valley. Saturated deposits represented as active cells in top and sometimes underlying layers of ground-water flow model; cells have high hydraulic conductivity and high specific yield.
Lacustrine (Lake Lahontan) deposits.	.....do	0–50	Unconsolidated silt, clay, sand, and some gravel.	Occurs below altitudes of 4,400 feet as prominent levee delta between Misfits Flat and Carson River. Deposits on much of valley floor covered by younger alluvium. Generally fine-grained materials with high porosity and low hydraulic conductivity; however, beach deposits, beds of sand, and some gravel lenses may yield water readily to wells. Simulated in ground-water flow model as active cells in either top or underlying layer of model. Hydraulic conductivity may range from low to high depending on deposit type.
Older alluvium	Pleistocene or Pliocene.	0–3,000	Unconsolidated and partly consolidated sand, gravel, silt, and clay.	Exposed in limited areas along north margin of valley. Also occurs at depth. Lenses of sand or gravel may yield moderate to large volumes of water to wells. Hydraulic conductivity thought to decrease with depth. Simulated as active cell in lower and middle layers of ground-water flow model; hydraulic conductivity of cells in middle layer is moderate to high and in lower layer is generally low.
			Cenozoic and Mesozoic rocks	
Basaltic rocks	Pleistocene or Pliocene.	—	Predominantly thin lava flows with interbeds of scoriaceous basalt breccia and diatomaceous sedimentary rocks.	Comprises most of Churchill Butte and occurs as scattered outcrops along south border of basin. Hydraulic conductivity may be high to moderate in scoriaceous zones and along numerous cooling joints and fractures. These generally saturated and fractured rocks are simulated as active cells in top or middle layer of ground-water flow model.
Andesitic rocks	Pliocene and Miocene.	—	Flow breccias, lava flows, and agglomerates with interbedded sedimentary rocks. Locally includes basaltic and rhyolitic rocks.	Crops out extensively around valley margins. Includes the Alta and Kate Peak Formations and the Chloropagus Formation of Axelrod (1956). Has virtually no interstitial permeability but locally may have zones of moderate to high permeability due to cooling fractures and joints. Flow breccias may have localized zones of moderate to high hydraulic conductivity. Generally simulated in ground-water flow model as inactive cells except in localized areas that are characterized by highly fractured rock.
Sedimentary rocks.	.....do	—	Tuffaceous sandstone, siltstone, diatomaceous shale, and rhyolitic tuff.	Crops out along northeast boundary of area. Includes the Coal Valley Formation of Axelrod (1956). Generally low permeability; simulated in ground-water flow model by inactive cells.
Rhyolite tuff	.....do	—	Devitrified, slightly to strongly welded, crystal-rich ash-flow tuff.	Crops out in some minor exposures in northwestern part of basin. Includes the Hartford Hill Rhyolite Tuff of former usage. Simulated in ground-water flow model by inactive cells.
Granitic rocks	Cretaceous	—	Nonporphyritic quartz monzonite, granodiorite, and mafic rocks, undivided.	Represents two intrusive bodies whose outcrops total several square miles in surface exposure in western part of area. Has very low hydraulic conductivity; locally may transmit some water if highly fractured. Simulated in ground-water flow model by inactive cells.
Metasedimentary rocks.	Jurassic or Triassic.	—	Mainly shale, slate, tuffaceous siltstone, sandstone, and graywacke; largely derived from volcanic rocks. Minor interbeds of conglomerate, limy shale, limestone, dolomite, and gypsum.	Minor exposures in western part and at northeast margin of valley. Low hydraulic conductivity; transmits water only where fractured. Simulated in ground-water flow model by inactive cells.

emplacement of poorly permeable granitic rocks during Late Cretaceous to early Tertiary time (96 to 63 Ma). This event formed the mass of poorly permeable granitic rocks on the west border of the area and altered existing rocks of Triassic or Jurassic age to form the metamorphic rocks exposed on the east flank of the Flowery Range.

A major tectonic change occurred during Miocene time (24 to 5 Ma) with the onset of extensional faulting at about 17 Ma. This faulting formed the major basins and ranges that characterize the present-day physiography. The initial extension was accompanied by volcanic activity that continued into late Tertiary time and produced widespread sequences of andesitic rocks throughout the area. In late Tertiary time, a structural basin began to form, and the sediments mapped by Rose (1969) as the Coal Valley Formation were deposited in the north-

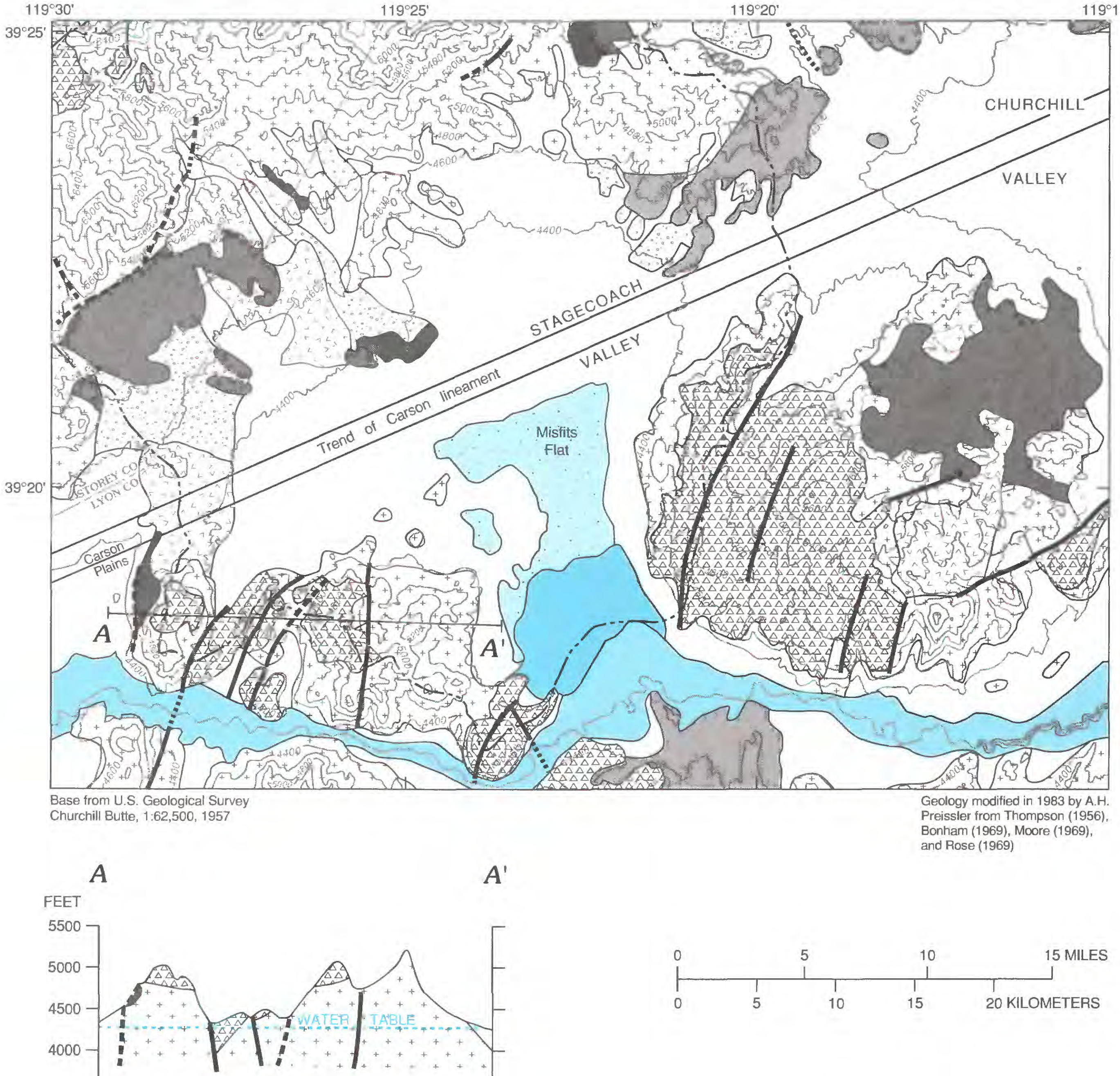


FIGURE 2.—Generalized geology of Stagecoach Valley area. See table 1 for description of lithologic units. "WATER TABLE" in section refers to *projection* of water table from beneath flood plain of Carson River.

ern part of the area. The boundaries of the structural basin associated with these deposits probably did not coincide well with the present-day boundaries of Stagecoach Valley. High-angle extensional faulting probably was occurring contemporaneously, and during this same general time a structural block that underlies Churchill Butte and extends about 2 mi west beneath Stagecoach Valley began tilting to the west. This tilting initiated the deformation that accounts for the main structural depression underlying Stagecoach Valley.

During Pliocene or early Pleistocene time, renewed volcanic activity covered a significant part of the area with basalt flows. Structural deformation, primarily high-angle normal faulting and possibly some left-lateral movement along the Carson lineament, continued during this time, completing the structural depression that now underlies Stagecoach Valley.

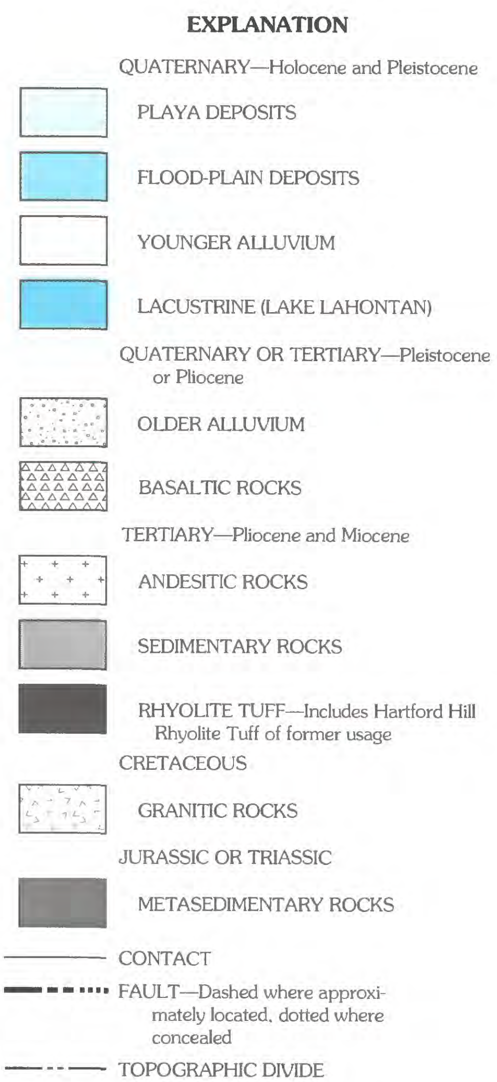


FIGURE 2.—Continued.

In Pleistocene time, before the development of ancient Lake Lahontan (which, during part of the Pleistocene, covered much of western Nevada), the major topographic features of the study area were similar to those of today, except that the area was not a topographically closed basin. Runoff from the Flowery Range flowed southeastward across Stagecoach Valley and contributed to the pre-Lahontan flow of the Carson River. When Lake Lahontan formed and reached its highest levels (altitude of about 4,366 ft in the vicinity of Stagecoach Valley), most of the valley floor was submerged to a depth of at least 100 ft. During interlake stages, sediment carried by the Carson River was deposited across the mouth of the pre-Lahontan drainage from Stagecoach Valley, forming a pronounced natural levee that blocked surface discharge from Stagecoach Valley and caused the playa, Misfits Flat, to form. Hydrologic processes, such as subsurface outflow and ground-water evapotranspiration, also were affected. The exact time (or times) at which blockage occurred is not known; however, radiocarbon dating indicates that Lake Lahontan rose high enough to flood the basin and then receded at least three times during the Lake Lahontan pluvial period: at least once before 40,000 yr B.P. (radiocarbon years before 1950), from 25,000 to 21,500 yr B.P., and from 13,600 to 11,100 yr B.P. (Benson, 1978, p. 312-315). For the past 11,000 yr the area has been a topographically closed basin, and a relatively thin sequence of playa deposits has accumulated behind the topographic divide formed by the levee. Lacustrine beaches and nearshore deposits associated with the rise and recession of Lake Lahontan form relatively permeable units in the upper basin fill.

HYDROLOGIC SETTING

CLIMATE AND PRECIPITATION

The climate in Stagecoach Valley and the surrounding mountains varies from arid to semiarid, depending principally on altitude. There are no precipitation stations in the area, but average annual precipitation at three stations within 10 to 15 mi of Stagecoach Valley is 9 in. at Virginia City, altitude 6,002 ft; 4.4 in. at Lahontan Dam, altitude 4,158 ft; and 6.6 in. at Fernley, altitude 4,160 ft (Glancy and Katzer, 1975, p. 18). Precipitation for most of the valley is within this range except for those areas above 6,000 ft altitude, which generally receive more precipitation. Most of the precipitation falls as rain and snow in the winter months; minor amounts of rain fall during thunderstorms in the summer. Summers are hot and dry, with daytime temperatures occasionally exceeding 100°F, and winters are cool, with temperatures sometimes falling below 0°F. The average growing season is about 120 days.

SURFACE WATER

The streams in Stagecoach Valley flow only during periods of rapid snowmelt or intense precipitation. Their channels are poorly defined downstream from the mountains because flow usually spreads out over the porous Quaternary basin-fill sediments and infiltrates rapidly. Consequently, significant flow reaches Misfits Flat only during exceptionally wet seasons or after periods of intense rainfall. Flow that reaches Misfits Flat ponds and evaporates. There is no surface outflow to the Carson River. Because of the intermittent flow and the small size of the drainage areas, the streams probably contribute relatively small amounts of recharge to the ground-water system after leaving the mountains.

The Carson River is one of the major rivers draining the east flank of the Sierra Nevada. Its headwaters are in the Sierra Nevada, about 60 mi southwest of Stagecoach Valley, and the river terminates in the Carson Sink, about 60 mi northeast of Stagecoach Valley. Based on 71 yr of record, the average annual flow at a gaging station on the Carson River near Fort Churchill, about 2 mi downstream from Stagecoach Valley, was 264,000 acre-ft as of 1982 (Frisbie and others, 1983, p. 141). An additional 16,000 acre-ft/yr of water is diverted upstream of the gaging station through the Buckland ditch (Glancy and Katzer, 1975, p. 66), so the total flow averages about 280,000 acre-ft/yr.

Much of the annual flow occurs during snowmelt periods in May or June. Lowest flows generally occur during August, September, or October. There have been some periods of no flow at the gaging station each year since 1923 (Frisbie and others, 1983, p. 141); however, low-flow conditions in the Buckland ditch (fig. 1) are not well documented, so firm flow-duration estimates of periods of low flow cannot be made. Schroer and Moosburner (1978, p. 211) indicated that flow at the gaging station equaled or exceeded 1 ft^3/s 85 percent of the time and that flow equaled or exceeded 10 ft^3/s about 70 percent of the time.

Reaches of the Carson River are in hydraulic continuity with the aquifers in Stagecoach Valley. The river may be either a source of subsurface inflow or a drain for subsurface outflow, depending on the local hydraulic gradients. The annual flow of the river is much greater than the annual ground-water flow through Stagecoach Valley; consequently, changes in inflow to or outflow from Stagecoach Valley aquifers would probably not be significant in relation to overall annual volume of river flow. However, the changes may be significant during periods of very low flow.

SOURCE, OCCURRENCE, AND MOVEMENT OF GROUND WATER

Virtually all the ground water in Stagecoach Valley is derived from two sources: precipitation that falls within the drainage basin and subsurface inflow from beneath the flood plain of the Carson River. Most recharge occurs in or adjacent to the mountain fronts (where precipitation infiltrates a thin mantle of unconsolidated deposits), through weathered or fractured consolidated rock, or where intermittent runoff infiltrates channel deposits and percolates down to the zone of saturation.

Ground water occurs in saturated basin fill under both water-table and confined conditions. The general depth to ground water in 1971, prior to development, is shown in figure 3. The general configuration of the water table in Stagecoach Valley prior to development in 1971 is shown in figure 4. The water-level contours were constructed using the earliest data available, mostly measurements in the early 1970's. Contours along the flood plain of the Carson River are based on controls from a few wells and on controls at points where contours on topographic maps cross the river. In the north and central parts of the valley, flow is generally from areas adjacent to the Flowery Range toward Misfits Flat and the flood plain of the Carson River southeast of Misfits Flat. The water table between Misfits Flat and the Carson River is nearly flat. However, the flood plain of the Carson River has an eastward gradient of about 7 ft/mi. Consequently, along the downstream reach of the river some water may flow northward into Stagecoach Valley in the vicinity of Table Mountain, whereas a short distance downstream water flows southeastward from Stagecoach Valley toward the flood plains of the river. There is probably a small net outflow from Stagecoach Valley to the Carson River.

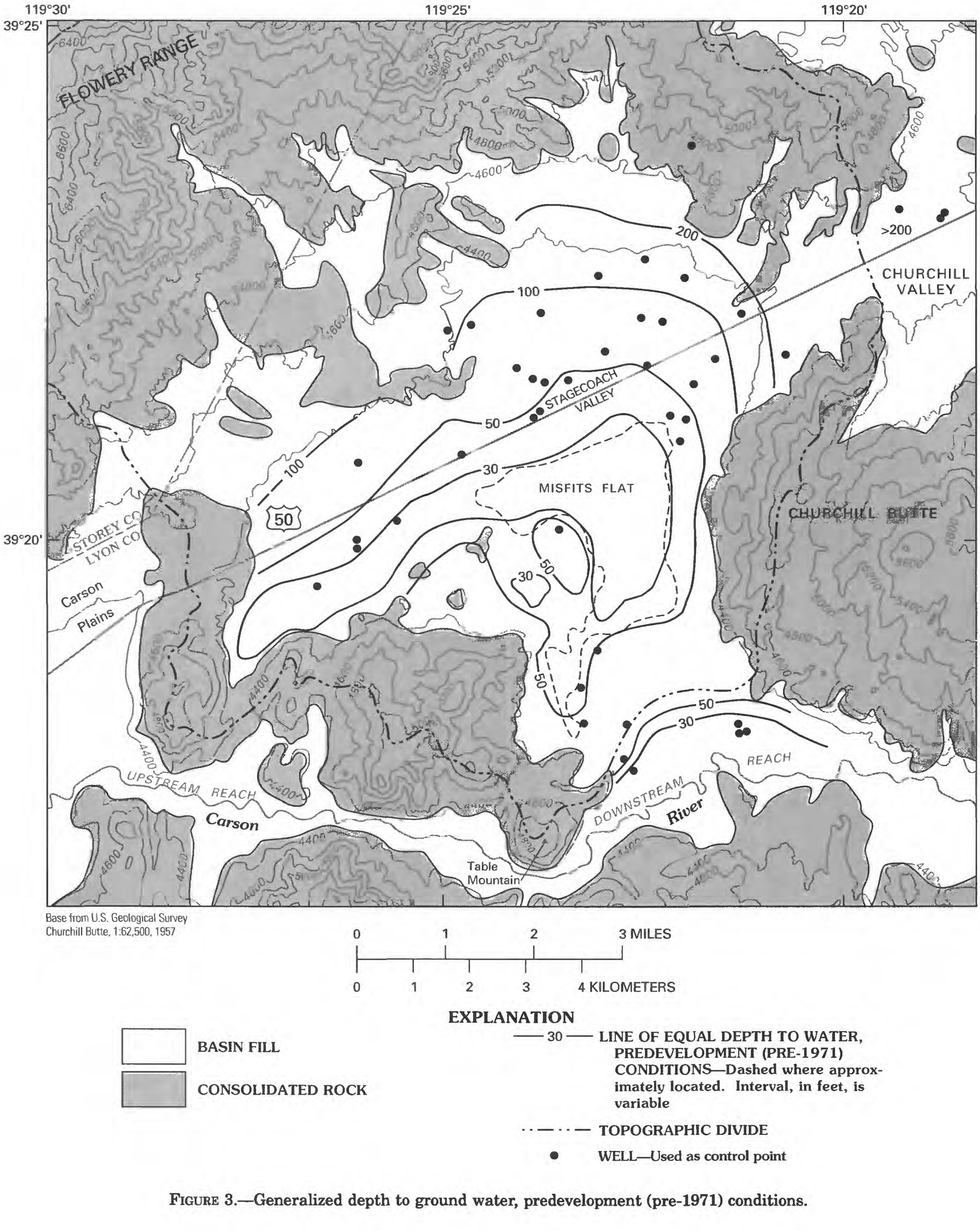
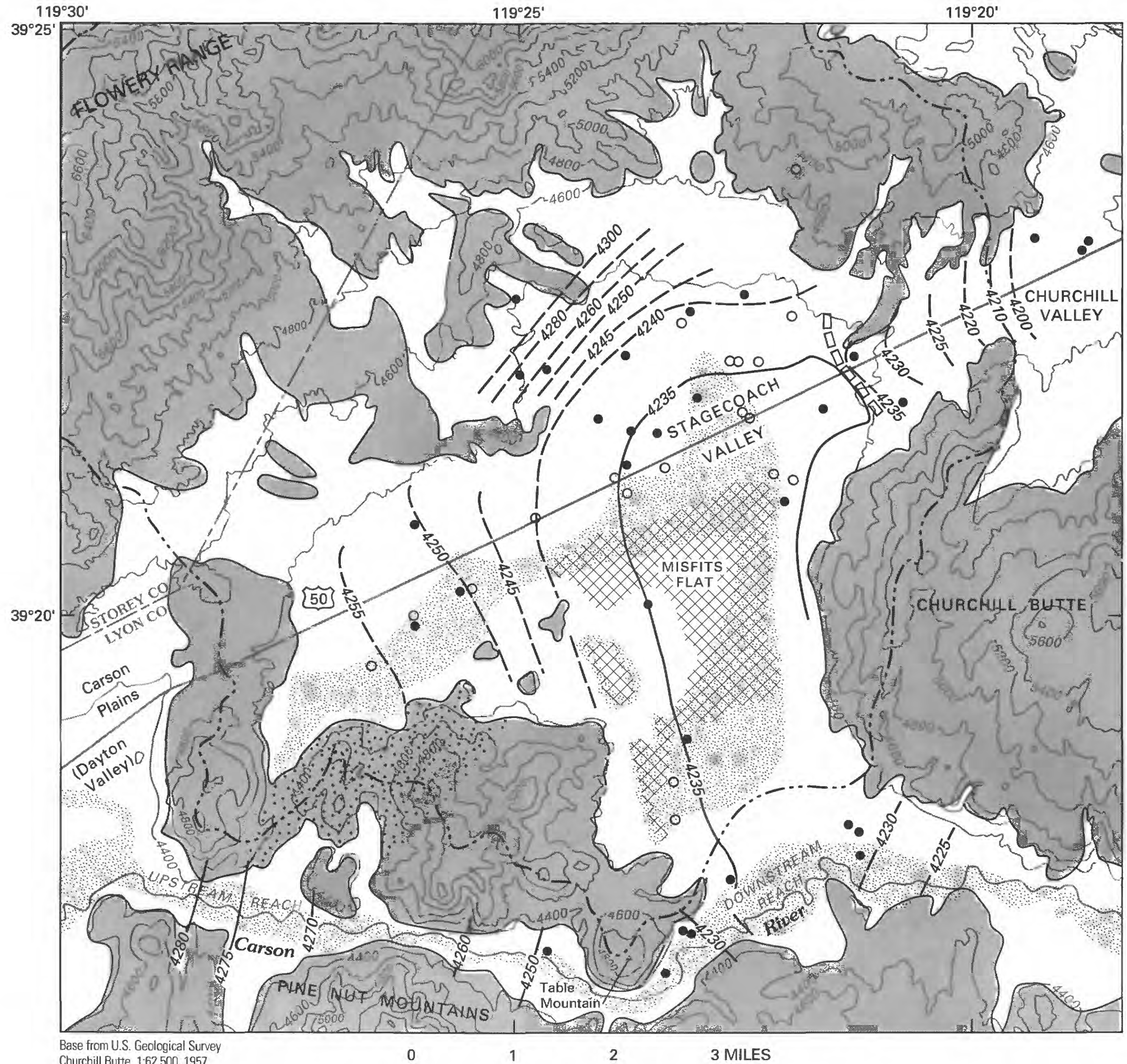


FIGURE 3.—Generalized depth to ground water, predevelopment (pre-1971) conditions.



EXPLANATION

BASIN FILL—Cross hachures indicate area of playa

CONSOLIDATED ROCK—Stipple indicates rock permeable enough to allow subsurface inflow to Stagecoach Valley

AREA OF NATURAL GROUND-WATER DISCHARGE—Vegetated primarily by greasewood (P.A. Glancy, U.S. Geological Survey, written commun., 1982)

—*4240*— WATER-LEVEL CONTOUR, PREDEVELOPMENT (PRE-1971) CONDITIONS—Dashed where approximately located. Contour interval, in feet, is variable. Datum is sea level

GROUND-WATER DIVIDE

TOPOGRAPHIC DIVIDE

● WELL—Used as primary control point

○ WELL—Where data were extrapolated back to estimate predevelopment (pre-1971) conditions. Used as supplemental control point

FIGURE 4.—Generalized water-level contours in basin-fill deposits, predevelopment (pre-1971) conditions.

In sec. 1, T. 17 N., R. 23 E. (fig. 4) the ground-water divide between Stagecoach Valley and Churchill Valley is located about 1½ mi west of the topographic divide. The steep hydraulic gradient between Stagecoach Valley and Churchill Valley could indicate either low permeability or a narrowing of the flow section.

In the western part of the valley, the water-level contours indicate that water moves northeastward toward Misfits Flat (fig. 4). Harrill and others (1993) use an average hydraulic gradient of about 5 ft/mi, a width of section of about 1 mi, and an estimated transmissivity of about 4,000 ft^2/d to estimate an annual flow of 130 to 170 acre-ft/yr moving northeastward from the area of postulated subsurface inflow (fig. 4).

Because most local recharge is generated in areas of higher altitude in the Flowery Range, the expected direction of flow is generally eastward or southeastward toward the playa. The 4,250-ft contour (fig. 4) was based on data from three wells that all had measurements in the early 1970's, were completed in the same interval of saturated basin fill, and had altitudes determined by leveling; consequently, the northeast flow direction indicated by this contour is considered valid. This suggests that inflow occurs in the southwestern part of the valley. The most probable explanation is that ground water beneath the flood plain of the Carson River, with a head about 25 ft higher than ground water in adjacent Stagecoach Valley, flows northeastward into Stagecoach Valley through fractured consolidated rock. Harrill and others (1993) conclude that this hypothesis is valid by evaluating the geologic feasibility of flow through the andesitic and basaltic rocks at the south end of the valley and by showing that the stable isotopic composition of water downgradient from the area of inferred inflow is similar to that of water in the Carson River.

BASIN-FILL AQUIFER

Stagecoach Valley occupies a structural depression that is bounded and underlain by consolidated rock. The depression is filled in part by deposits of gravel, sand, silt, and clay derived primarily from the adjacent mountains. These deposits form the basin-fill aquifer in Stagecoach Valley; the sand and gravel in particular yield a large amount of water to wells.

AREAL EXTENT

The approximate areal extent of the basin-fill aquifer of Stagecoach Valley is shown by the extent of younger alluvium in figure 2. Total surface area is about 33 mi^2, or almost 50 percent of the total drainage area of Stagecoach Valley (fig. 1). The basin-fill aquifer is bounded on the north and east by consolidated rock of the Flowery Range and Churchill Butte, respectively. On the west an alluvium-covered topographic divide exists between Stagecoach Valley and the Carson Plains part of Dayton Valley and is underlain by consolidated rock at relatively shallow depth. The southwestern part of the basin-fill aquifer is bounded by exposures of volcanic rock, but in the southeastern part the aquifer is continuous with the flood-plain deposits of the Carson River.

THICKNESS

Wells drilled in Stagecoach Valley range in depth from several tens of feet to 820 ft, and most are completed in basin fill. Gravity surveys were used to obtain estimates of the thickness of basin-fill deposits. The vertical acceleration of gravity is affected, in part, by the density of underlying materials. Unconsolidated basin-fill deposits typically have substantially lower densities than adjacent and underlying consolidated rock, therefore gravity anomalies in the valley are roughly proportional to the thickness of basin fill.

Schaefer (1988) constructed a detailed Bouguer gravity anomaly map on the basis of about 100 measurements made in the Stagecoach Valley area. These data were processed using techniques similar to those used in producing maps of gravity anomalies in several valleys in west-central Nevada (Schaefer, 1983, p. 6–9). Thickness of basin fill was estimated semiquantitatively using a computer program that makes a three-dimensional analysis of gravity anomalies based on a specified density contrast between basin-fill deposits and consolidated rock (Cordell and Henderson, 1968). A density contrast of 0.5 g/cm^3 was used in this analysis. Two seismic profiles were made to verify the thicknesses computed from the gravity data.

The structural depression that underlies the valley is complex. The thickest deposits of basin fill occur along a zone that extends from the north-central part of the valley beneath Misfits Flat to the area northeast of Table Mountain (fig. 5). Moore (1969, p. 22) stated that Churchill Butte was the exposed part of a westward-tilted fault block. The north-south-trending area of thickest fill shown on figure 5 may indicate the downdropped west edge of this fault block. There also are indications of a shallow structural depression oriented along the general trend of the Carson lineament. The thickest fill (3,000 ft) occurs slightly southwest of Misfits Flat. However, most of the fill is between 500 and 2,000 ft thick. In the southeastern part of the valley, the basin fill appears to be continuous with flood-plain deposits of the Carson River, and the depth to consolidated rock northeast of Table Mountain is estimated to range from 500 to more than 1,000 ft (see the area of downstream reach shown in fig. 5). Data from a seismic profile at Misfits Flat show an increase in density at a depth of about 530 ft. The denser material may be partly consolidated fill and probably does not transmit water as readily as the overlying deposits.

HYDRAULIC PROPERTIES

HYDRAULIC CONDUCTIVITY

Most alluvial deposits in Stagecoach Valley are nearly flat lying; thus, horizontal hydraulic conductivity is usually much greater than vertical hydraulic conductivity. Approximate values of the horizontal hydraulic conductivity of deposits typical of those in the basin-fill aquifers are listed in table 2; these values agree closely with those shown for similar materials by Chow (1964, fig. 13-8).

The general distribution of horizontal hydraulic-conductivity values in Stagecoach Valley was mapped using lithologic descriptions from drillers' logs, the results of one pumping test, and 11 specific capacities. The estimated average horizontal hydraulic-conductivity values of the upper 150 ft of saturated basin fill are shown in figure 6.

The average vertical hydraulic conductivity of a sequence of deposits is estimated as a geometric mean that is influenced strongly by low values typical of intervals of silt and clay. The ratio between the average vertical and average horizontal hydraulic conductivities varies according to the types of material in the interval evaluated. Sequences of well-sorted sand and gravel commonly have higher vertical hydraulic conductivities than sequences that contain significant amounts of clay, silt, or cemented materials, for which the average vertical hydraulic-conductivity values can be as small as 0.1 to 1 percent of the horizontal hydraulic conductivity. The distribution of estimated vertical hydraulic-conductivity values in the upper 150 ft of saturated basin fill (fig. 7) was based on the distribution of geologic materials and estimated ratios between the average vertical and horizontal conductivity. The range of values for average vertical hydraulic conductivity shown in figure 7 is less than the range of values shown in figure 6 for horizontal hydraulic conductivity because the average values of vertical hydraulic conductivity are weighted toward values typical of silt and clay.

TABLE 2.—*Hyraulic conductivities of basin-fill deposits*

[Geologic units are described more fully in table 1]

Geologic unit	Typical materials	Estimated range of horizontal hydraulic conductivity (feet per second)
Playa deposits	Clay and silt	1.0×10^{-6} to 3.0×10^{-6}
	Very fine sand	1.0×10^{-6} to 1.8×10^{-5}
Flood-plain deposits	Sand	4.6×10^{-5} to 3.0×10^{-4}
	Gravel	2.3×10^{-4} to 1.7×10^{-3}
Younger alluvium	Mostly silt, some sand and gravel	1.0×10^{-6} to 4.6×10^{-5}
	Sand and gravel[1]	4.6×10^{-5} to 1.7×10^{-3}
Lacustrine (Lake Lahontan) deposits.	Silt and clay	1.0×10^{-6} to 6.0×10^{-5}
	Fine sand	1.0×10^{-5} to 4.6×10^{-5}
Older alluvium[2]	Mostly silt, some sand and gravel[1]	1.0×10^{-6} to 4.6×10^{-5}
	Sand and gravel	4.6×10^{-5} to 1.7×10^{-3}

[1]Poorly sorted mixtures that occur primarily in fanglomerate deposits in younger and older alluvium.

[2]Same typical materials as younger alluvium, but materials are more likely to be semiconsolidated. Same range of estimated horizontal hydraulic conductivity is shown for both younger and older alluvium, but materials in older alluvium are more likely to have values near lower end of range because of greater consolidation and cementation.

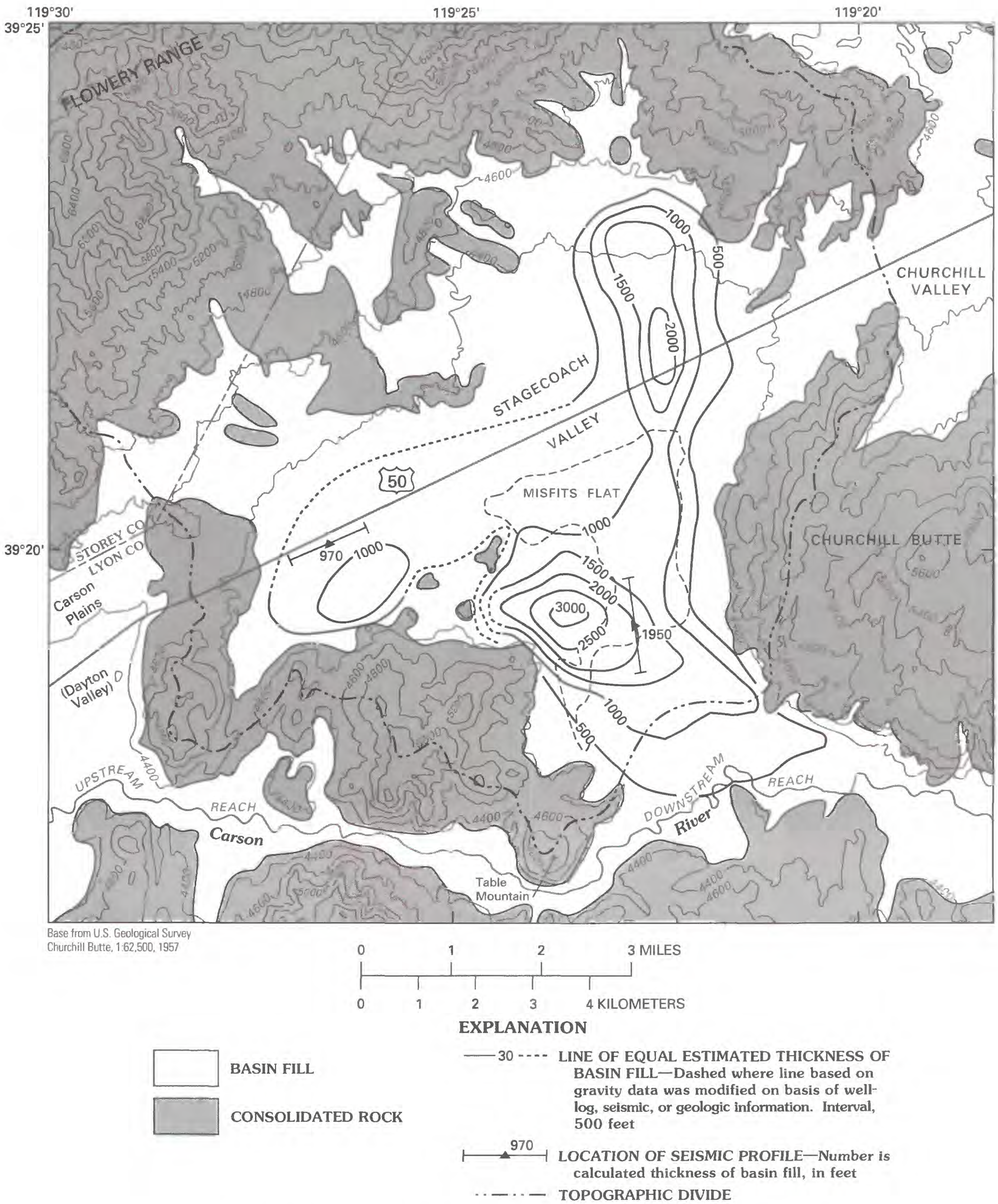
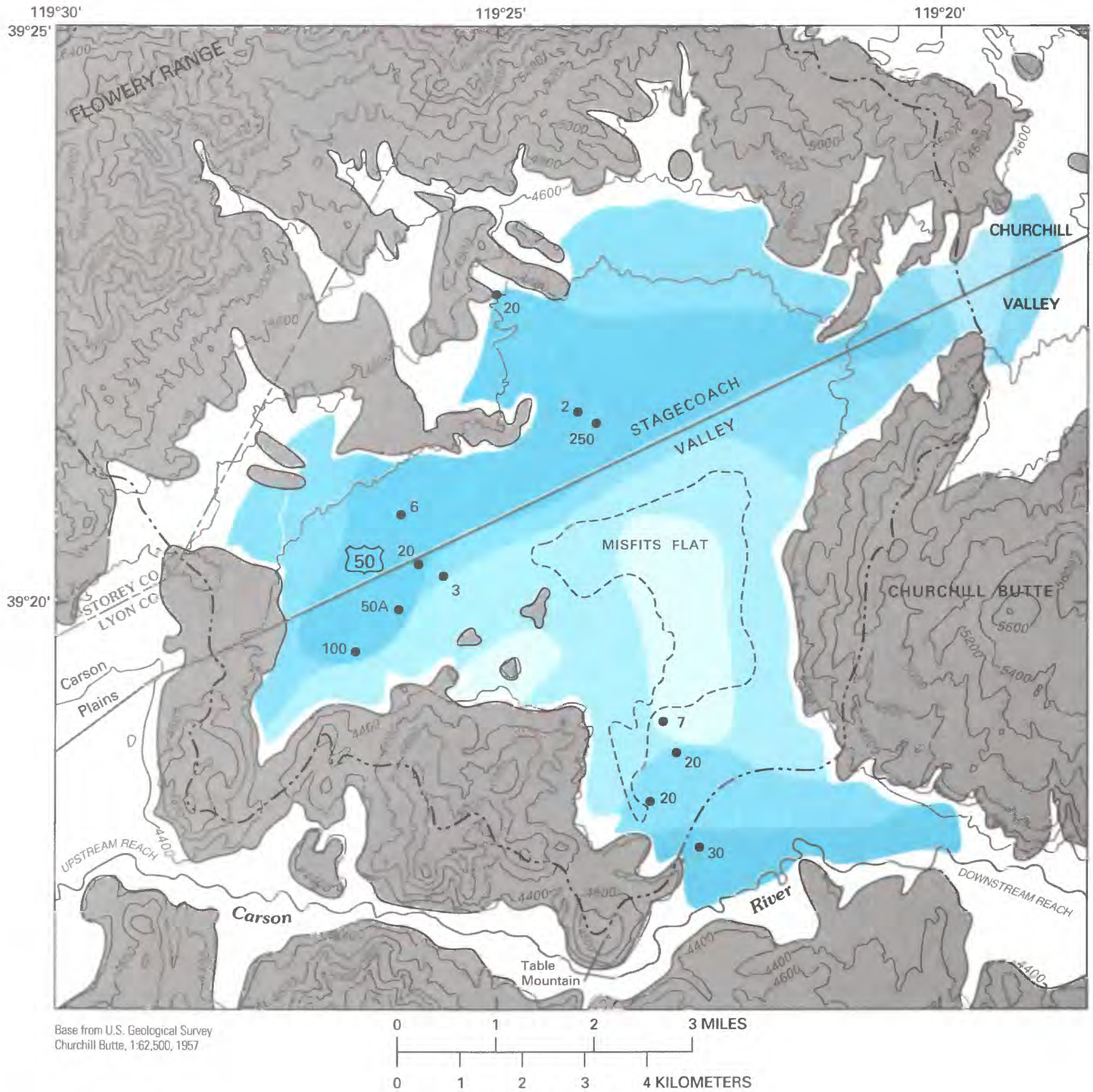


FIGURE 5.—Estimated thickness of basin-fill deposits. Adapted from Schaefer (1988).



EXPLANATION

CONSOLIDATED ROCK

ESTIMATED HORIZONTAL HYDRAULIC CONDUCTIVITY OF BASIN FILL, IN FEET PER SECOND

- Less than 1.0×10^{-5}
- 1.0×10^{-5} to 7.5×10^{-5}
- 7.5×10^{-5} to 12.5×10^{-5}
- Greater than 12.5×10^{-5}
- Not estimated

— · · — TOPOGRAPHIC DIVIDE

● 20 WELL—Where hydraulic conductivity was estimated on basis of specific capacity or aquifer test. Estimated hydraulic conductivity is given in 1×10^{-5} feet per second. "A" following value signifies value based on aquifer test

FIGURE 6.—Distribution of horizontal hydraulic-conductivity values, upper 150 feet of saturated basin fill.

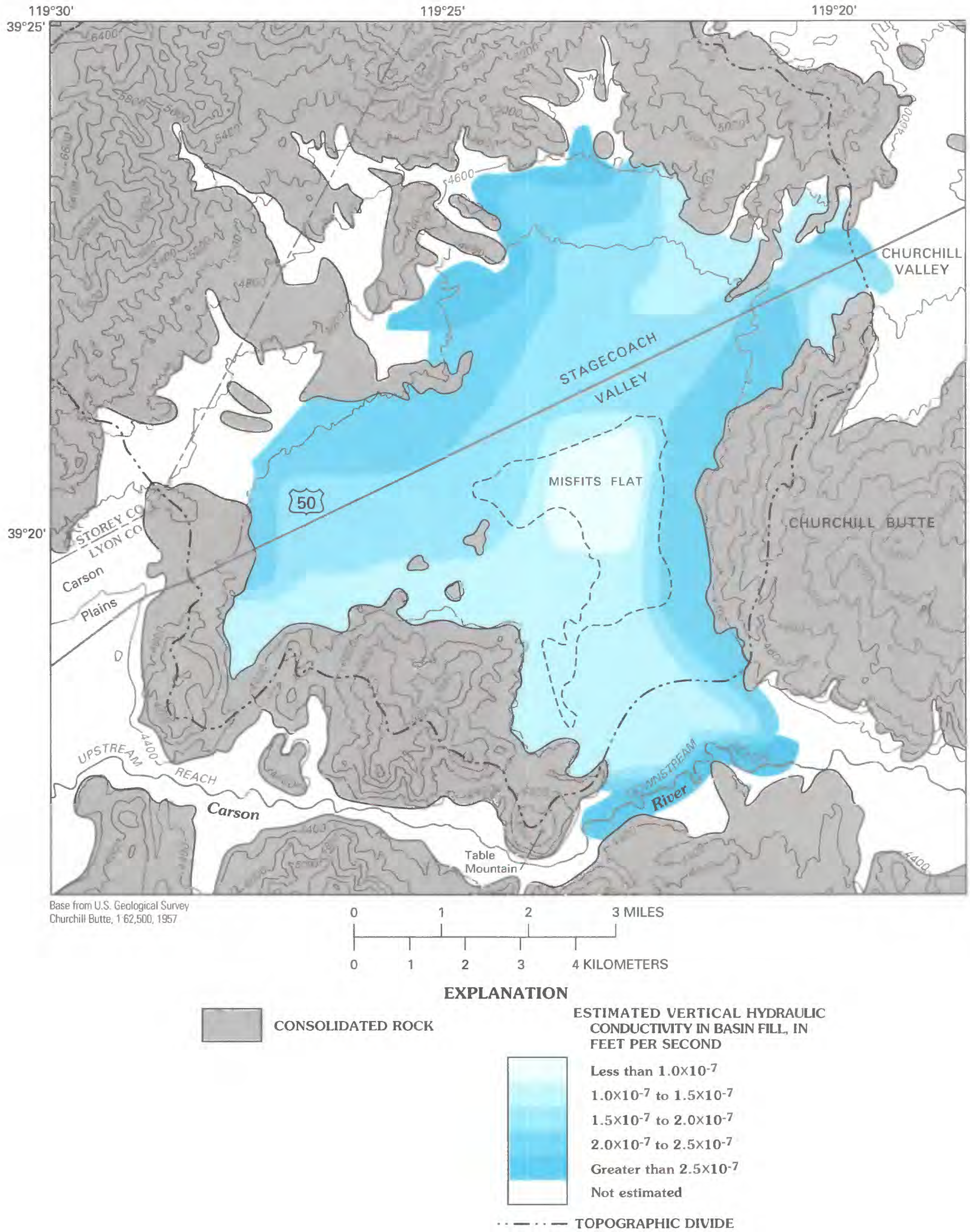


FIGURE 7.—Estimated vertical hydraulic conductivity, upper 150 feet of saturated basin fill.

STORAGE PROPERTIES

Specific yield of alluvial deposits ranges from less than 5 percent in compacted clay or deposits with extensive cementation or caliche development to about 30 percent in well-sorted sand or gravel (fig. 8). Estimates of specific yield were developed by evaluating lithologic descriptions in drillers' logs and assigning values based on specific-yield data listed by Morris and Johnson (1967) for similar deposits. An area of predominantly playa deposits beneath Misfits Flat was assigned an average specific yield of less than 5 percent. Areas on the valley floor underlain predominantly by fine-grained lacustrine and other deposits were assigned average specific-yield values of 5 to 10 percent. The alluvial-fan and associated coarse-gravel deposits around the margins of Stagecoach Valley were assigned specific yields of greater than 10 percent. Assigned values for individual deposits ranged from 10 to 30 percent, with the higher values representing well-sorted sand and gravel (primarily reworked beach and nearshore lacustrine deposits) and the lower values representing poorly sorted deposits such as mudflows. Average specific yield for these deposits was estimated to be about 13 percent.

Coarse-grained deposits, such as sand and gravel, are characterized by a relatively rigid framework that is supported by grain-to-grain contact; aquifer compression in these deposits is typically elastic. Fine-grained deposits, such as clay or silt, generally do not possess a rigid framework. In a confined aquifer, if the head is high enough to keep framework stresses below the level of preconsolidation stress of the sediments, then water yielded would be from elastic expansion of water. However, if the head falls to the level where framework stresses exceed the preconsolidation stress of the sediments but is still above the altitude of the confining bed, then water yielded would be from inelastic compression of the sediments as well as from expansion of water. Studies in other areas (Holzer, 1981) have shown that water levels had to be lowered about 50 ft, to as much as 200 ft below prepumping levels, before the preconsolidation stress of the sediments was exceeded and inelastic depletion of storage began. Consequently, for this study, the ground-water system is considered to be an elastic system. In several model scenarios developed in a later section of this report, localized drawdowns exceed 100 ft; however, errors induced by not including inelastic storage are considered relatively small and probably do not significantly affect the overall results.

For elastic conditions, Lohman (1972, p. 53) stated that the rule-of-thumb relation to roughly estimate a storage coefficient of a confined aquifer is to multiply the thickness of the aquifer by 1×10^{-6}. In Stagecoach Valley, saturated basin fill ranges from 0 to 3,000 ft in thickness. If no fine-grained deposits were present and yielded water was entirely from the expansion of stored water in the confined part of the aquifer, the storage coefficient of confined deposits for thicknesses of 100 to 3,000 ft would be 1×10^{-4} to 3×10^{-3}.

GROUND WATER IN STORAGE

The potentially recoverable water stored in the basin-fill aquifer can be estimated as the product of an area, a thickness, and a specific yield. The following procedures were used to compile estimates of water stored in the basin-fill aquifer of Stagecoach Valley: (1) The area underlain by basin fill was divided into 176 rectangular cells identical to those used in the ground-water model discussed later in this report. (2) For each cell the thickness of saturated basin fill was determined using total thickness of fill shown in figure 5 and depths to water shown in figure 3; the specific yield was determined from figure 8. (3) Two intervals were used, the upper 200 ft of saturation and the underlying remainder of the saturated basin fill. The water stored in the interval below 200 ft was further reduced by 5 to 15 percent of the estimated amount to account for the expectation of lower specific yields in the older, more deeply buried deposits. (4) Values for cells were summed to give estimates of total stored ground water (table 3). Computations were made for pre-1971 conditions for two general areas, the area that was developed as of 1982 and the remainder of the basin-fill aquifer (fig. 8).

The total amount of water stored in the basin-fill aquifer of Stagecoach Valley prior to development (before 1971), estimated to be about 1 million acre-ft (table 3), illustrates the large volume of water that has accumulated over many centuries in a comparatively small basin. Much of this water cannot be pumped because of economic factors (such as high pumping lifts, which mean high costs using existing technology), legal issues (such as probable depletion of flow in the Carson River), and environmental factors (such as land subsidence and drying of habitat along the flood plain of the Carson River). The quality of the deeper water is not known; however, in the upper 200 ft of saturated

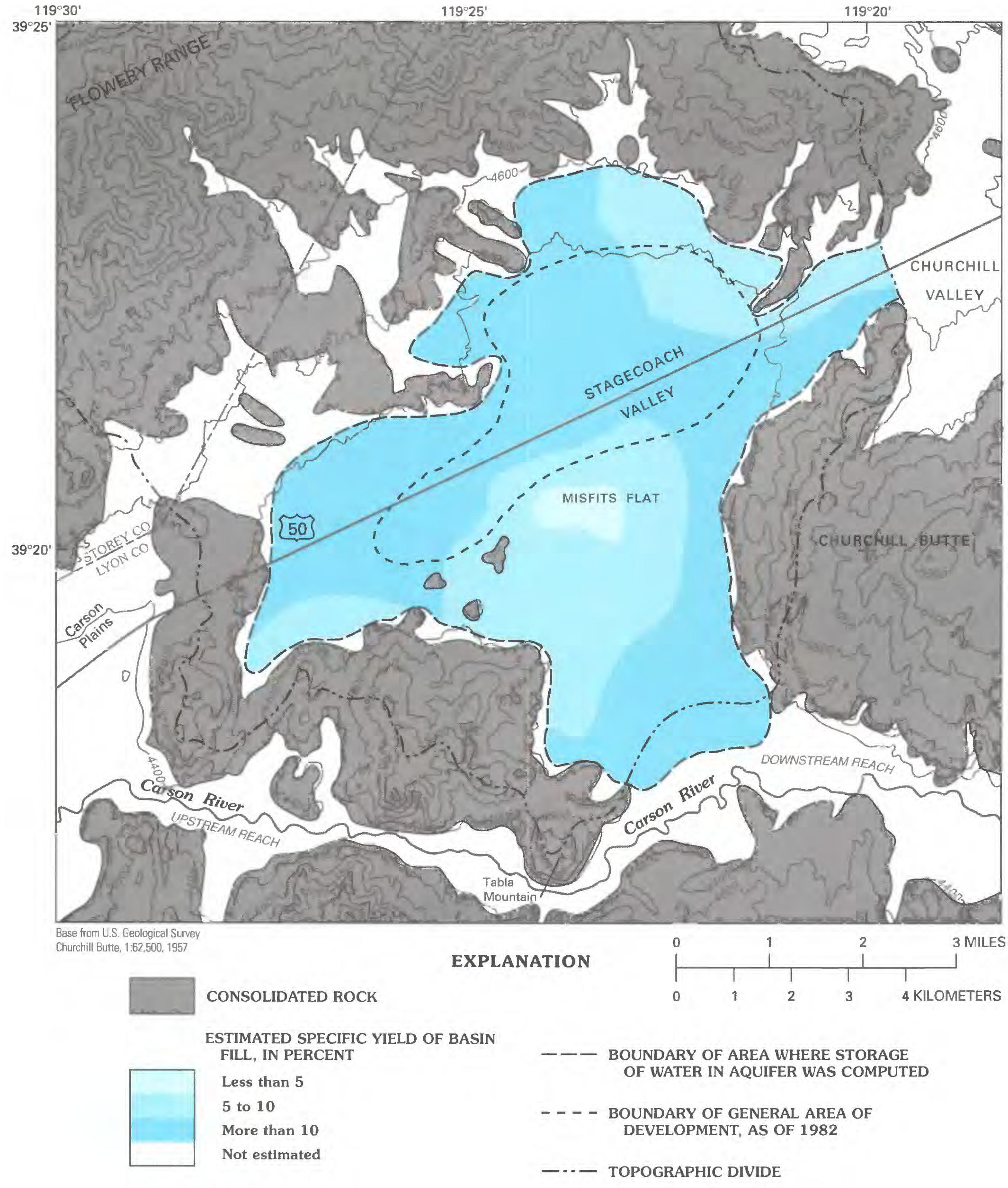


FIGURE 8.—Distribution of estimated specific-yield values, upper 150 feet of saturated basin fill.

TABLE 3.—*Estimates of water stored in basin-fill aquifer of Stagecoach Valley, predevelopment (pre-1971) conditions*

[—, not applicable. All values are rounded to two significant figures]

Depth of saturated basin fill (feet)	Ground water in storage (acre-feet)		
	Inside area developed as of 1982	Outside area developed as of 1982	Total
Stagecoach Valley			
0–200	89,000	180,000	270,000
>200	220,000	470,000	690,000
Subtotal	310,000	650,000	960,000
Area between Carson River and topographic divide of Stagecoach Valley[1]			
0–200	—	26,000	26,000
>200	—	45,000	45,000
Subtotal	—	71,000	71,000
Combined areas			
Total	310,000	720,000	1,000,000

[1]Area that is expected to be affected under simulated development scenarios.

basin fill, the estimated 89,000 acre-ft of water inside the 1982 area of development and some of the 180,000 acre-ft outside that area probably can be pumped with economic pumping lifts, using the 1982 distribution of pumping. Even so, depletion of this amount of storage might reduce heads significantly in places and would probably induce significant inflow from the Carson River.

RECHARGE FROM PRECIPITATION

MAXEY-EAKIN METHOD

Ground-water recharge from precipitation was estimated using the method outlined by Eakin and others (1951, p. 79–81). The average precipitation for altitude zones within the basin (fig. 1) was estimated using data from local stations and values from a map of Nevada showing annual precipitation (Hardman and Mason, 1949, p. 10). Then for each altitude zone the percentage of the total precipitation that would potentially become ground-water recharge was estimated using coefficients developed by Maxey and Eakin (1949, p. 40–41). The highest estimated percentages of recharge occur in the mountain areas, where most of the annual precipitation occurs. Over most of the valley floor, the estimated percentage of recharge is negligible, although some recharge may be generated in the valley floor area during extremely localized storms. The estimated average annual recharge to the basin-fill aquifer of Stagecoach Valley from precipitation falling within the drainage area is about 580 acre-ft/yr (table 4).

CHLORIDE-BALANCE METHOD

A technique to estimate recharge on the basis of chloride balance between precipitation and ground water was developed and applied to selected areas in the Great Basin by Dettinger (1989). This technique assumes that chloride is conserved, that areas of recharge are known, that chloride concentrations in precipitation (C_p) and in ground water (C_g) in and near recharge areas are known, and that the volume of precipitation that falls on the recharge areas (V_p) can be estimated. When these assumptions are valid, the volume of recharge (V_r) can be approximated as follows:

$$V_r = C_p {}^* V_p / C_g.$$

Using 0.4 mg/L as the average concentration of chloride in precipitation, 12,200 acre-ft/yr as the average volume of precipitation at altitudes of 5,000 ft or higher (see table 4), and 11.0 mg/L as the average concentration of chloride in ground water in the recharge area based on 10 samples from 4 springs and wells in or near recharge areas in Stagecoach Valley (Dettinger, 1989, p. 69), the recharge to the basin-fill aquifer from precipitation is estimated to be about 440 acre-ft/yr. Although this estimate is somewhat less than the 580 acre-ft/yr estimate (table 4), both indicate relatively low volumes of annual recharge and are considered to be reasonable.

DISCHARGE BY EVAPOTRANSPIRATION

Of about 4,000 acres of phreatophyte vegetation mapped in the valley (fig. 4), about 3,000 acres consists primarily of low- to moderate-density greasewood (generally less than 10 percent areal cover) and about 1,000 acres (Misfits Flat) consists primarily of bare soil and low-density greasewood mounds. Average annual ground-water consumption in these areas was estimated to be 0.2 and 0.1 ft/yr, respectively, on the basis of rates used by Glancy and Katzer (1975, p. 62–64). Using these rates and areas, the total estimated evapotranspiration of ground water in Stagecoach Valley is about 700 acre-ft/yr.

TABLE 4.—*Estimated average annual recharge to basin-fill aquifer from precipitation, Stagecoach Valley*

[Average annual precipitation for altitude zones and percentage recharge are same as those used for Dayton Valley by Glancy and Katzer (1975, p. 48). —, negligible or not applicable. Totals are rounded to two significant figures]

Altitude zone (feet above sea level)	Area (acres)	Average annual precipitation		Average annual recharge	
		(feet, rounded)	(acre-feet, rounded)	(percent)	(acre-feet, rounded)
7,000–8,000	30	1.5	45	15	7
6,000–7,000	4,630	1.1	5,090	7	360
5,000–6,000	8,890	.8	7,110	3	210
<5,000	31,600	.5	15,800	—	—
Total	45,000	—	28,000	—	[1]580

[1]Additional 30 acre-feet per year recharge, mostly generated from precipitation that falls on southwest flank of Churchill Butte, is estimated to occur in area between topographic divide of Stagecoach Valley and Carson River.

GROUND-WATER DEVELOPMENT

Ground-water withdrawals for crop irrigation and for public water supply began in 1971, and about 320 acre-ft of water was pumped that year. Pumpage increased steadily until 1977, when the total annual pumpage was about 1,600 acre-ft. Pumpage has declined slightly since then and in 1982 was about 1,200 acre-ft. Changes in irrigated land, population, and estimated pumpage during the period 1971–82 are shown in table 5. The distribution of irrigated land and areas of domestic and public supply by ground water in 1982 are shown in figure 9. Similar maps prepared for the 11 pumping seasons 1971 to 1982 were used to help estimate the distribution of pumpage values for the ground-water flow model.

During the 11 yr from spring 1971 to spring 1982 (excluding most of 1982 pumpage shown in table 5), a total of about 11,000 acre-ft of water was pumped from the basin-fill aquifer, mostly for irrigation. Most of the water was consumed by evapotranspiration; however, some of the water pumped for irrigation infiltrated deep enough to escape consumption. Initially, part of this infiltrated water was retained as soil moisture, and the remainder returned to the water table. Generalized contours of water levels in the aquifer as of spring 1982 are shown in figure 10. The net decline in ground-water levels during the period spring 1971 to spring 1982, determined on the basis of the difference between pre-1971 water levels (fig. 4) and the spring 1982 water levels (fig. 10), is shown in figure 11. Additional control was provided by nine wells with measured changes between 1972 (or earlier) and 1982. Declines of 5 to 15 ft occurred throughout most of the area of ground-water withdrawals, and the maximum measured decline was about 19.5 ft.

SIMULATION OF GROUND-WATER FLOW

The preceding sections of this report have presented information that describes ground-water flow in the basin-fill aquifer of Stagecoach Valley. The remainder of this report describes a digital computer model of the ground-water flow in the aquifer. The model was calibrated using (1) predevelopment conditions and (2) estimated pumpage from 1971 to 1982 and the resultant water-level declines. The calibrated model was used to simulate long-term trends that describe the probable future response to selected developmental scenarios.

CONCEPTUALIZATION OF GROUND-WATER FLOW IN BASIN-FILL AQUIFER

Ground-water flow in the basin-fill aquifer of Stagecoach Valley was conceptualized as a three-layer flow system (fig. 12). The top layer (layer 1, fig. 12*C*) represents the water table and 50 ft of underlying saturated material. The zone represented by layer 1 is where the processes of recharge from precipitation, irrigation-return flow, evapotranspiration, interaction with the Carson River, and depletion of water by draining of aquifer materials occur. The middle layer (layer 2, fig. 12*C*) represents the zone from 50 to 300 ft below the water table. The

TABLE 5.—*Estimated irrigated acreage, population, and ground-water pumpage, Stagecoach Valley, 1971–82*

[Totals are rounded to two significant figures]

Calendar year	Irrigated land[3] (acres)	Population[1]			Pumpage[2] (acre-feet per year)			
		Served by domestic wells	Served by public systems	Total	Irrigation	Domestic and stock	Public supply	Total
1971	100	20	30	50	300	>10	>10	>320
1972	110	20	60	80	330	10	10	350
1973	140	30	110	140	420	10	20	450
1974	315	40	140	180	940	10	30	980
1975	315	50	170	220	940	10	40	990
1976	390	60	150	210	1,170	10	30	1,200
1977	520	70	150	220	1,560	10	30	1,600
1978	420	80	370	450	1,260	20	80	1,400
1979	440	90	460	550	1,320	20	100	1,400
1980	280	100	510	610	840	20	110	970
1981	310	110	560	670	930	30	130	1,100
1982	330	110	570	680	990	30	130	1,200
Total					11,000	190	720	12,000

[1]Estimates based on field inventory of houses and lots in 1982, house count from photographs taken in 1974, dates when domestic wells reportedly were drilled, number of hookups on Stagecoach Utilities water system, and supplemental information from photographs taken in 1980. Population was estimated using average occupancy of 2.75 persons per household (U.S. Department of Commerce, 1983, p. 61, 63).

[2]Irrigation use based on estimated water use of 3 feet per acre. Alfalfa is principal crop, although some garlic was grown in early 1980's. Domestic and public-supply water use based on estimated use per household. Each household was assumed to contain, on average, 2.75 persons who use 100 gallons per day (gpd) each for household purposes. Overall use was estimated as sum of household use plus water used for lawn and garden irrigation. Estimates beyond field inventory of lawn and garden irrigation ranged from 0 to as much as 3 acre-feet per household in 1982. Thus, overall estimates based on 1982 inventory ranged from about 0.3 to about 3.3 acre-feet per household; average use was estimated to be about 0.65 acre-feet per household or about 210 gpd per person.

[3]Estimates based on field inventories, aerial photographs, orthophoto maps, and Landsat images.

zone represented by layer 2 is where virtually all pumpage occurs and is the conduit for much of the subsurface inflow and outflow. The bottom layer (layer 3, fig. 12*C*) represents the zone that extends from 300 ft below the water table to consolidated rock. This zone is the conduit for deep flow in the basin fill and a source of stored water. Vertical flow between the layers is restricted by the combined effect of numerous discontinuous, fine-grained deposits at various depths within the basin fill that collectively act as confining beds.

DIGITAL COMPUTER MODEL

A three-layer digital computer model was used to simulate the predevelopment conditions and the response to development of the basin-fill aquifer of Stagecoach Valley. The model is based on a partial differential equation that describes the three-dimensional movement of ground water of constant density through porous earth material. This equation and its approximation as a finite-difference expression were described by McDonald and Harbaugh (1988, p. 2-1 to 2-20). They also described a modular computer program that uses the finite-difference approximation of the differential equation of ground-water flow. This approximation is capable of simulating ground-water flow in a multilayer heterogeneous aquifer system with irregular borders and a variety of stresses such as pumping wells, evapotranspiration, and head-dependent flow to and from a river. Flow between layers is treated as vertical flow through confining beds; horizontal movement and storage of water in the confining beds is assumed to be insignificant. The basin-fill aquifer of Stagecoach Valley is represented by a three-layered finite-difference network composed of 25 columns and 22 rows with nodes spaced 2,000 ft apart at the center of each cell. The network has 210 active cells in the top layer (layer 1), 209 active cells in the middle layer (layer 2), and 170 active cells in the bottom layer (layer 3), as shown in figure 13. The perimeter of the model, outside the model cells, is surrounded by no-flow boundaries as required by the computer code. The strongly implicit procedure was used to solve the finite-difference equations simultaneously for flow between cells. (For detailed information about the model, see McDonald and Harbaugh, 1988.)

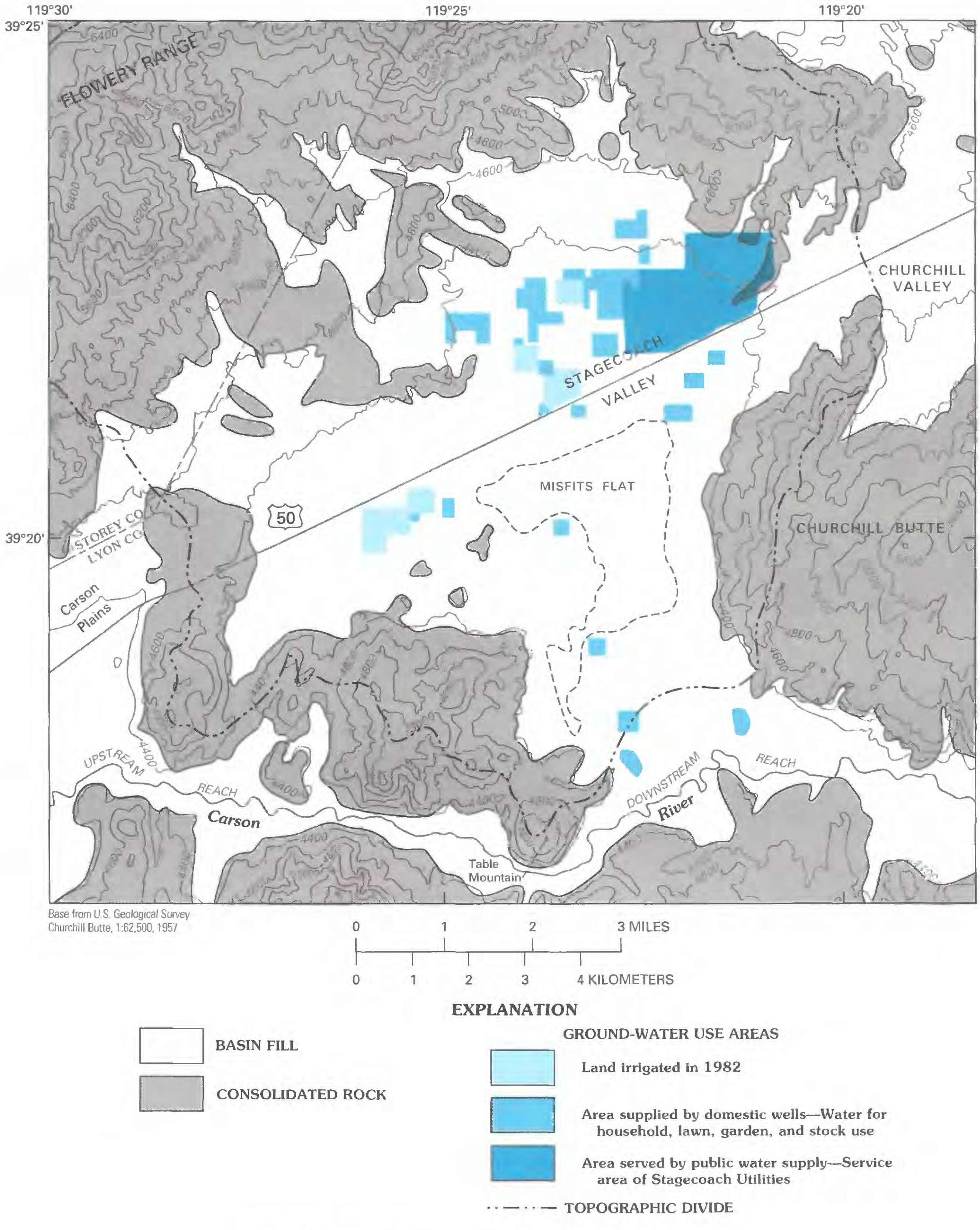
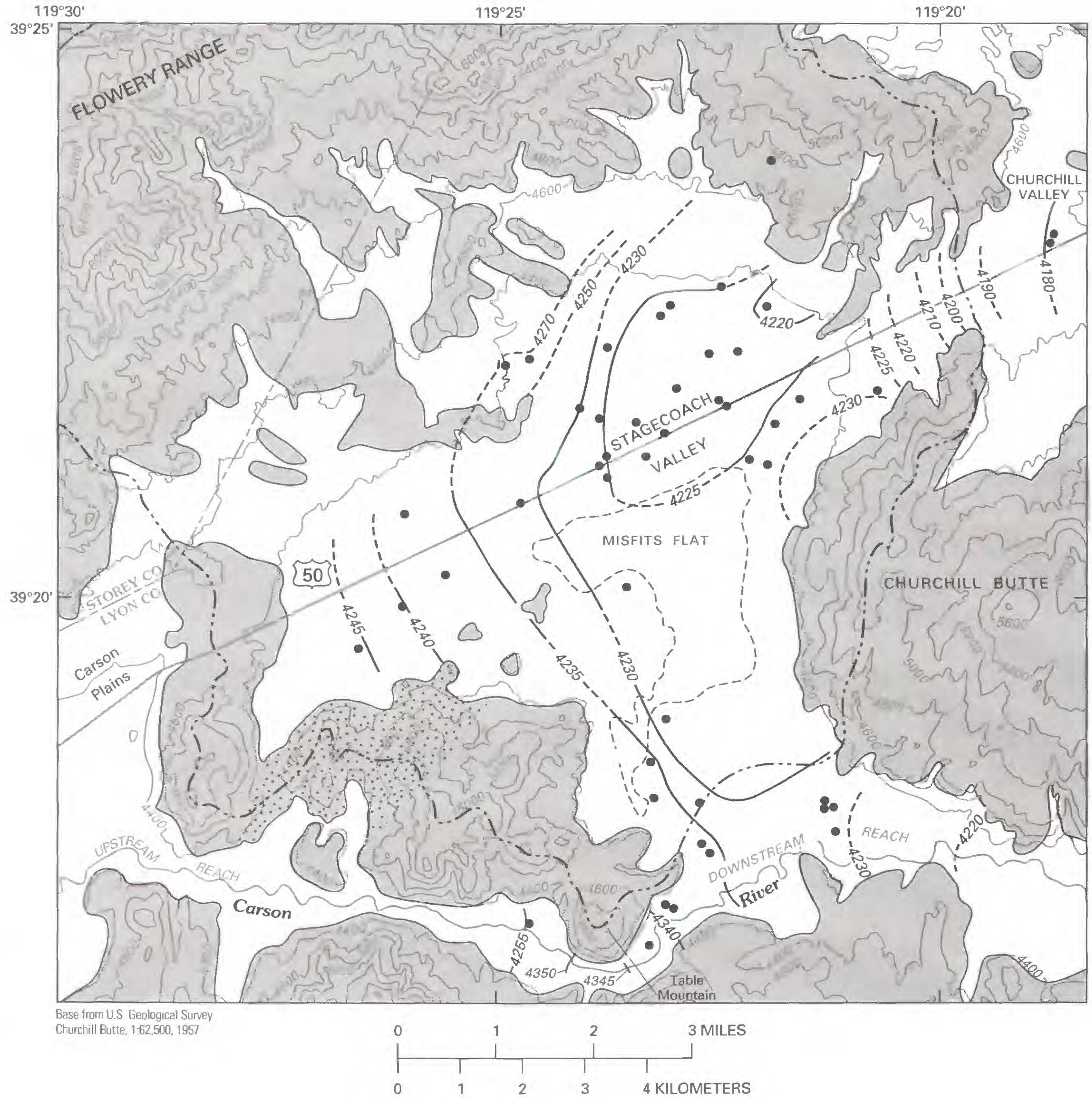


FIGURE 9.—Ground-water use areas in Stagecoach Valley, 1982.



EXPLANATION

BASIN FILL

CONSOLIDATED ROCK—Stipple indicates rock permeable enough to transmit subsurface flow to Stagecoach Valley

—4230— WATER-LEVEL CONTOUR, SPRING 1982—Shows altitude of water level for wells that penetrate most heavily pumped zone of basin fill. Dashed where approximately located. Contour interval, in feet, is variable. Datum is sea level

TOPOGRAPHIC DIVIDE

● WELL—Used as control point

FIGURE 10.—Ground-water-level contours, spring 1982.

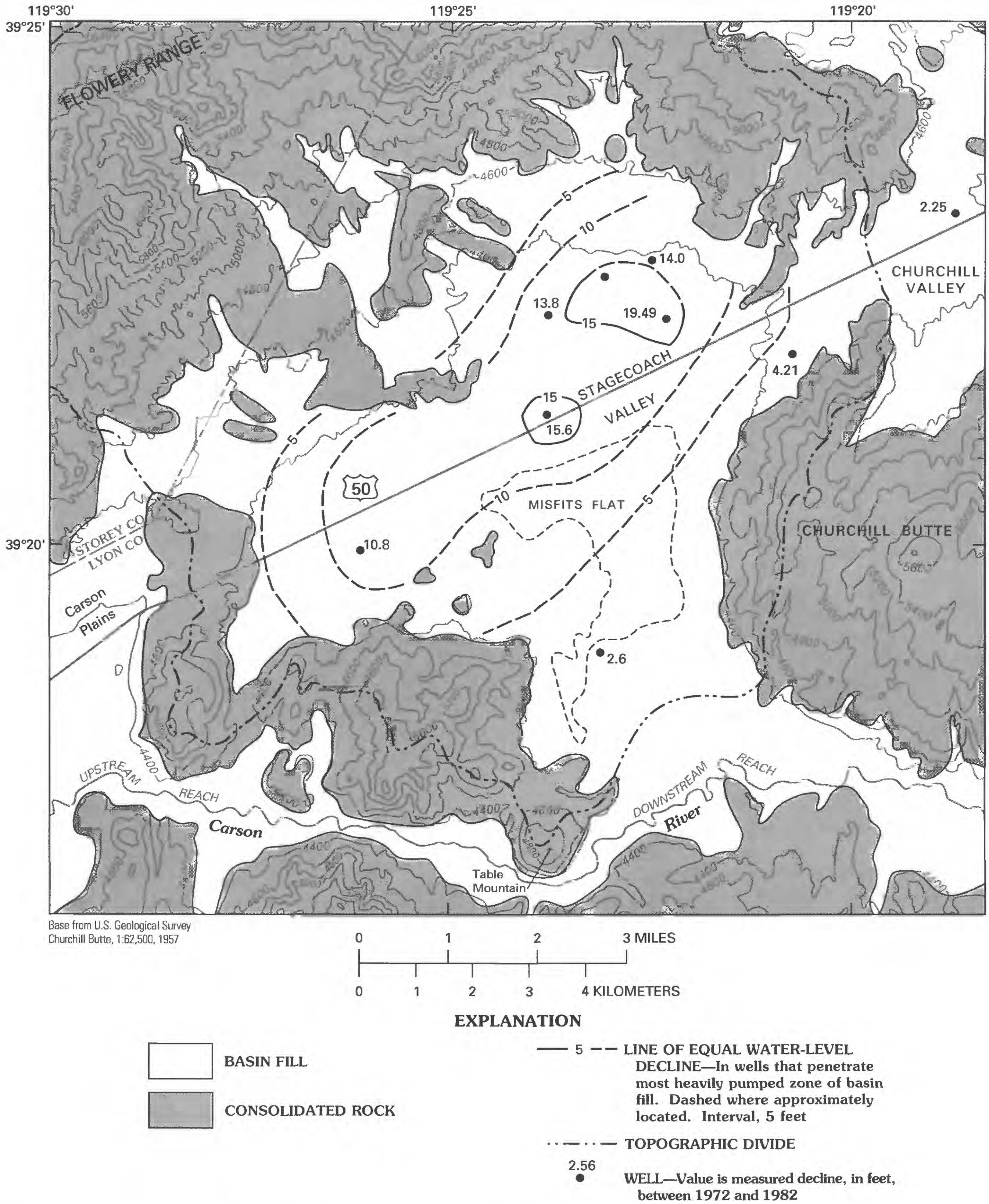


FIGURE 11.—Ground-water-level declines, spring 1971 to spring 1982. Most wells were completed in depth interval corresponding to model layer 2.

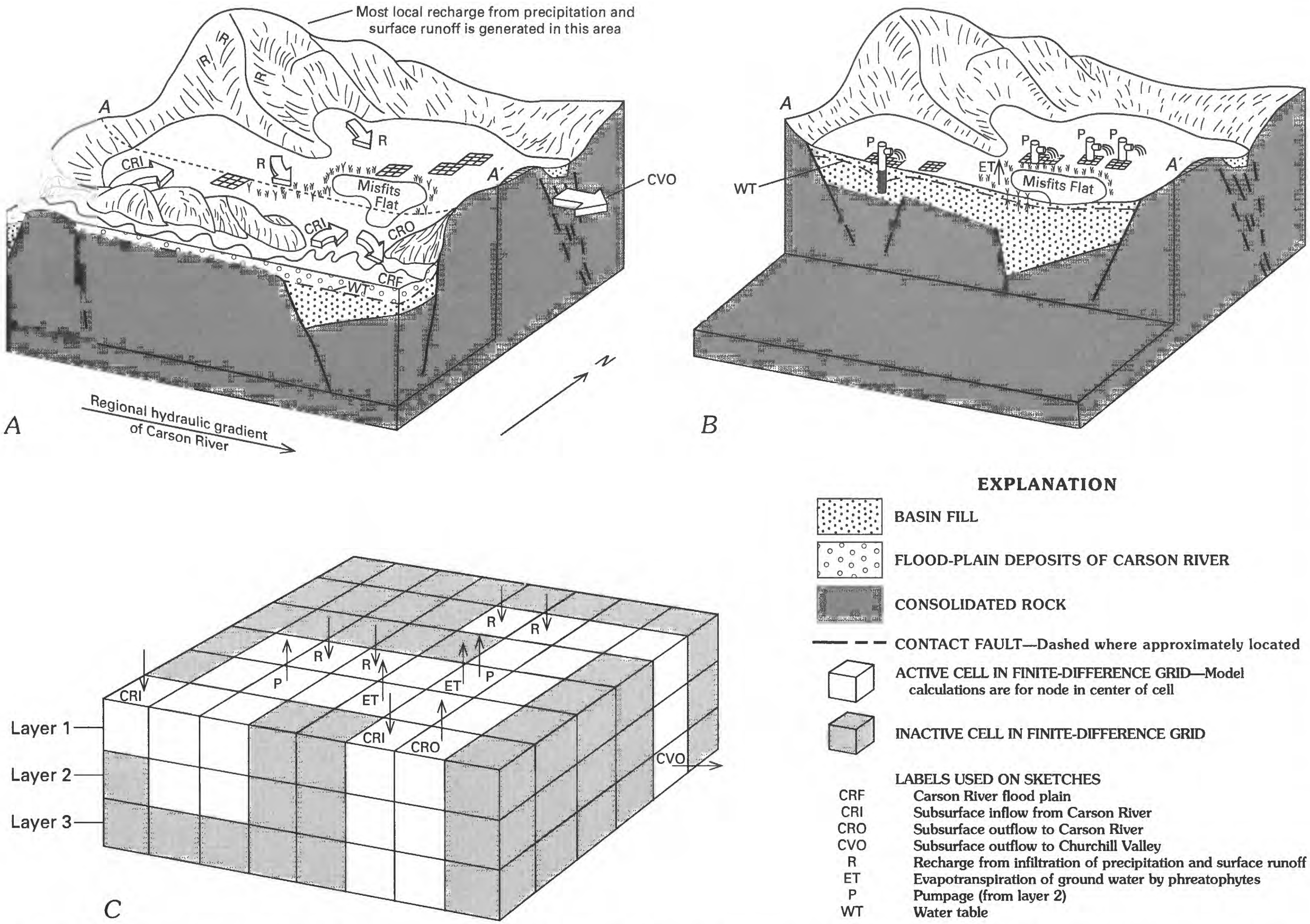
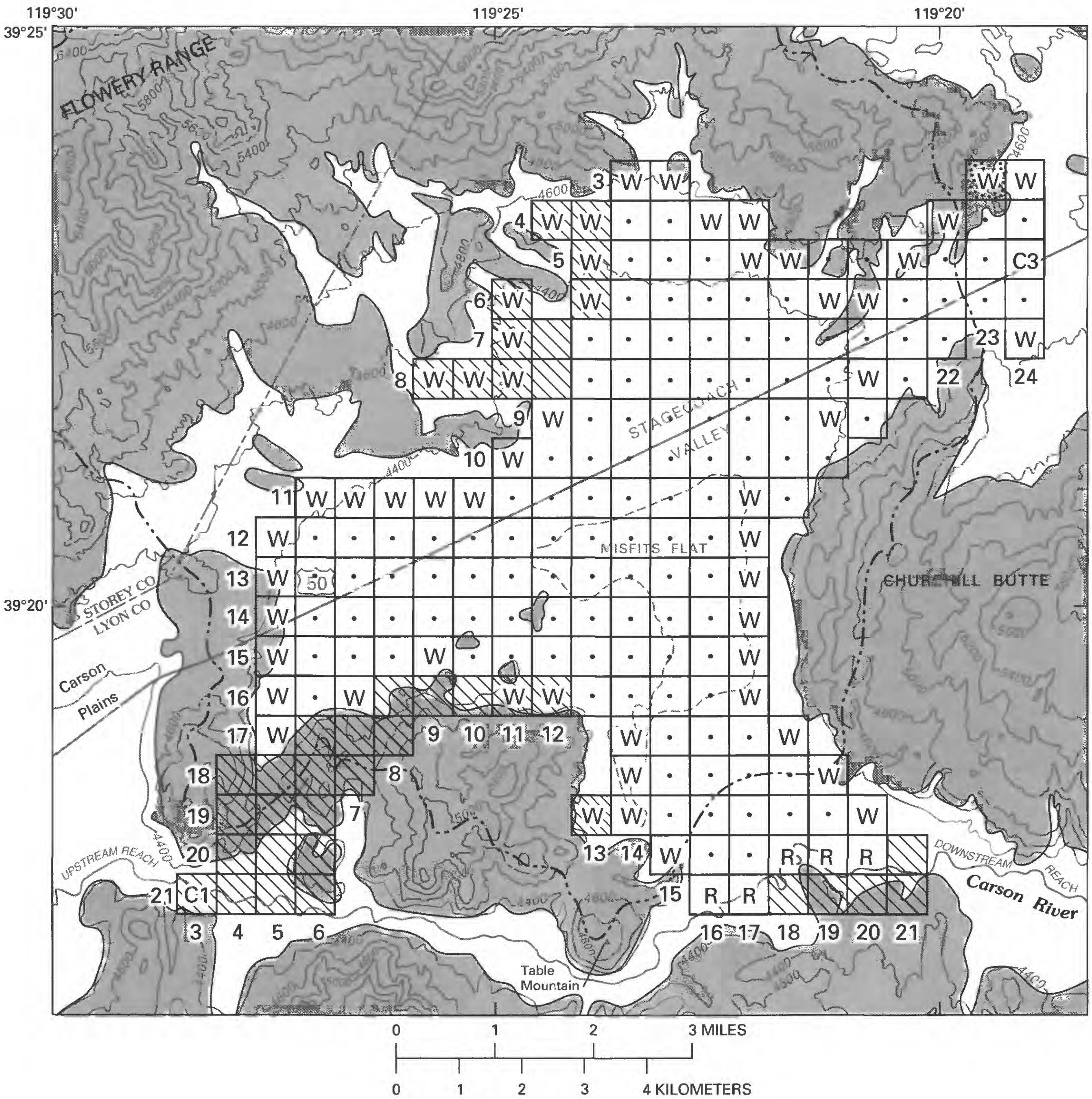


FIGURE 12.—Schematic block diagrams. *A*, Selected inflow-outflow relations. *B*, Generalized hydrogeologic section. *C*, Representation of basin-fill aquifer derived using three-dimensional finite-difference grid of ground-water flow model.



EXPLANATION

- BASIN FILL
- CONSOLIDATED ROCK
- TOPOGRAPHIC DIVIDE
- GRID CELL—Where only layer 1 is active
- GRID CELL—Where layers 1 and 2 are active
- GRID CELL—Where layers 1, 2, and 3 are active
- C1 GRID CELL—Where constant-head conditions in layer 1 are used to simulate inflow from Carson River in upstream part of Stagecoach Valley
- C3 GRID CELL—Where constant-head conditions in layer 3 are used to simulate outflow to Churchill Valley
- R GRID CELL—Where "river leakage package" is used to simulate inflow from or outflow to Carson River in downstream part of Stagecoach Valley
- W GRID CELL—Where constant flux of recharge is introduced into layer 1

FIGURE 13.—Finite-difference grid used to model ground-water flow system in Stagecoach Valley. Row number is shown at left end of row; column number is shown at bottom of column.

BOUNDARY CONDITIONS

The model solution is strongly influenced by the boundary conditions specified; consequently, evaluating and specifying the boundary conditions constituted an important part of this study. Cells in the finite-difference grid were designated either active or inactive depending on whether ground-water flow through the aquifer sediments could be represented by that cell (fig. 13). Generally, basin-fill deposits were considered permeable, whereas most areas of consolidated rock were considered to have very low hydraulic conductivity except for localized zones of fractured and weathered rock. Thus, basin-fill sediments were represented as active cells, and consolidated rock as inactive cells. Local areas of basalt and fractured andesite in the southwestern part of Stagecoach Valley (fig. 2) were considered permeable and were represented by active cells. Various flux conditions at the model boundaries were treated by assigning appropriate flow conditions to active cells adjacent to the no-flow boundary outside the model cells.

The northwest boundary of the study area is recharged by inflow from the Flowery Range and was treated as a constant-recharge boundary in the model. Cells along this boundary in layer 1 of the model were characterized by recharge at a constant rate (fig. 13). The west, north, east, and parts of the south boundaries of the area were modeled in a similar way, except that not all cells receive recharge (or rates of recharge to the cells were negligible).

Ground water in the northeastern part of the area flows into Churchill Valley in the subsurface. This outflow condition was simulated in the model by extending the area of active cells the equivalent of about 1 mi east into Churchill Valley and then representing the lowest water level in that area as a constant-head cell in layer 3 (cell C3 in fig. 13). This approximation is adequate to represent steady-state conditions. However, the assumption of constant-head cells may have introduced error in the simulation results because pumping changes flow rates to or from the constant-head cell but does not remove water from storage and does not change water levels in the constant-head cells. However, if the distance of the constant-head boundary is far away from the pumping centers, then the approximation probably is adequate for the intended uses of the model.

Two reaches of the Carson River are in hydraulic continuity with the basin-fill aquifer of Stagecoach Valley. Boundary conditions at the upstream reach were simulated by a constant-head cell in the top layer of the model, representing the Carson River (cell C1 in fig. 13). The head was set at the altitude of the average river stage. The constant-head cell provides a source of water to adjacent cells representing permeable basaltic and andesitic rocks that transmit flow into the basin-fill aquifer of Stagecoach Valley. Boundary conditions along the downstream reach of the Carson River were simulated by using the model's river package, whereby areas traversed by the river are represented by five active cells that are characterized by head-dependent fluxes to and from the river (McDonald and Harbaugh, 1988, p. 6-1 to 6-30). Average altitudes estimated from topographic maps were used to represent the average river stage. A riverbed conductance value of 0.10 ft^2/s was used to simulate continuity between the river and layer 1 of the model. This riverbed conductance was estimated using a river length of 2,000 ft in each cell, an average river width of 50 ft, a thickness of streambed material of 50 ft, and a vertical hydraulic conductivity of 5×10^{-5} ft/s. The high value of vertical hydraulic conductivity was used because the sand and gravel deposits that underlie the Carson River contain hardly any silt and clay and are highly permeable. Fluctuations in stage due to variations in river flow were ignored. The river was treated primarily as either a source of or sink for water.

INITIAL CONDITIONS

Prior to development (before 1971), the hydrologic system was in a state of dynamic equilibrium where, over the long term, the average recharge equaled the average discharge and there was no appreciable change in storage or decline in water level. Pumping of water from wells produces a nonequilibrium condition where some of the water is removed from aquifer storage and ground-water levels decline with time.

The pre-1971 ground-water levels shown in figure 4 were assumed to represent equilibrium conditions and are considered to be best represented by conditions in layer 2 of the model. Similar maps were prepared for layers 1 and 3 and were used as the initial head conditions to represent the undeveloped flow system in the model.

TRANSMISSIVITY

The transmissivity of each of the three layers in the model was initially estimated as the product of the thickness of that layer and a hydraulic-conductivity value. The values used for thickness of saturated material in layers 1 and 2 were the average thickness values assigned to these layers in the model (50 and 250 ft, respectively); the thickness value used for layer 3 was assumed to be the difference between the estimated total thickness of saturated basin-fill material (fig. 5) and the 300 ft already assigned to layers 1 and 2.

The horizontal hydraulic-conductivity values used for layers 1 and 2 were estimated from specific capacity or from aquifer tests (fig. 6). Horizontal-conductivity values used for layer 3 were derived by reducing these values 5 to 15 percent to account for the anticipated decrease in conductivity with depth. All these adjustments to initial estimates were made during the calibration process in model simulations. The distribution of transmissivity values used for the three layers of the model is shown in figures 14, 15, and 16, respectively. The adjustments made during the calibration process imply that values of hydraulic conductivity in layers 2 and 3 should be lower than the values shown in figure 6. These lower values worked best in the model and might be attributed to pre-Lake Lahontan deposits having lower values of hydraulic conductivity. The transmissivity of layer 1 (fig. 14) was derived from the calibrated model of steady-state conditions. For transient pumping simulation, the altitude of the bottom of layer 1 was entered into the model so that when simulated pumping removed water from storage and the water table was drawn down, the transmissivity would be adjusted to represent the newly reduced thickness of saturated material in layer 1. Transmissivities of layers 2 and 3 were considered to be constant because no water-table changes below layer 1 would occur under the pumping conditions used to calibrate the model for transient conditions for the period 1971–82.

LEAKANCE

Flow between the three layers of the model was simulated as vertical flow through two confining units, representing the cumulative effect of numerous discontinuous, lenticular, fine-grained deposits at various depths within the basin fill. The ability of a confining unit to transmit vertical flow is expressed as leakance, which is the ratio K'/b', in which K' and b' are the vertical hydraulic conductivity and thickness, respectively, of the confining unit (Lohman, 1972, p. 30). For the purposes of this model, the thickness of the confining unit was represented as the distance between the midpoints of two adjacent layers, and the vertical hydraulic conductivity used was the average vertical hydraulic conductivity of the basin-fill materials in the interval of interest.

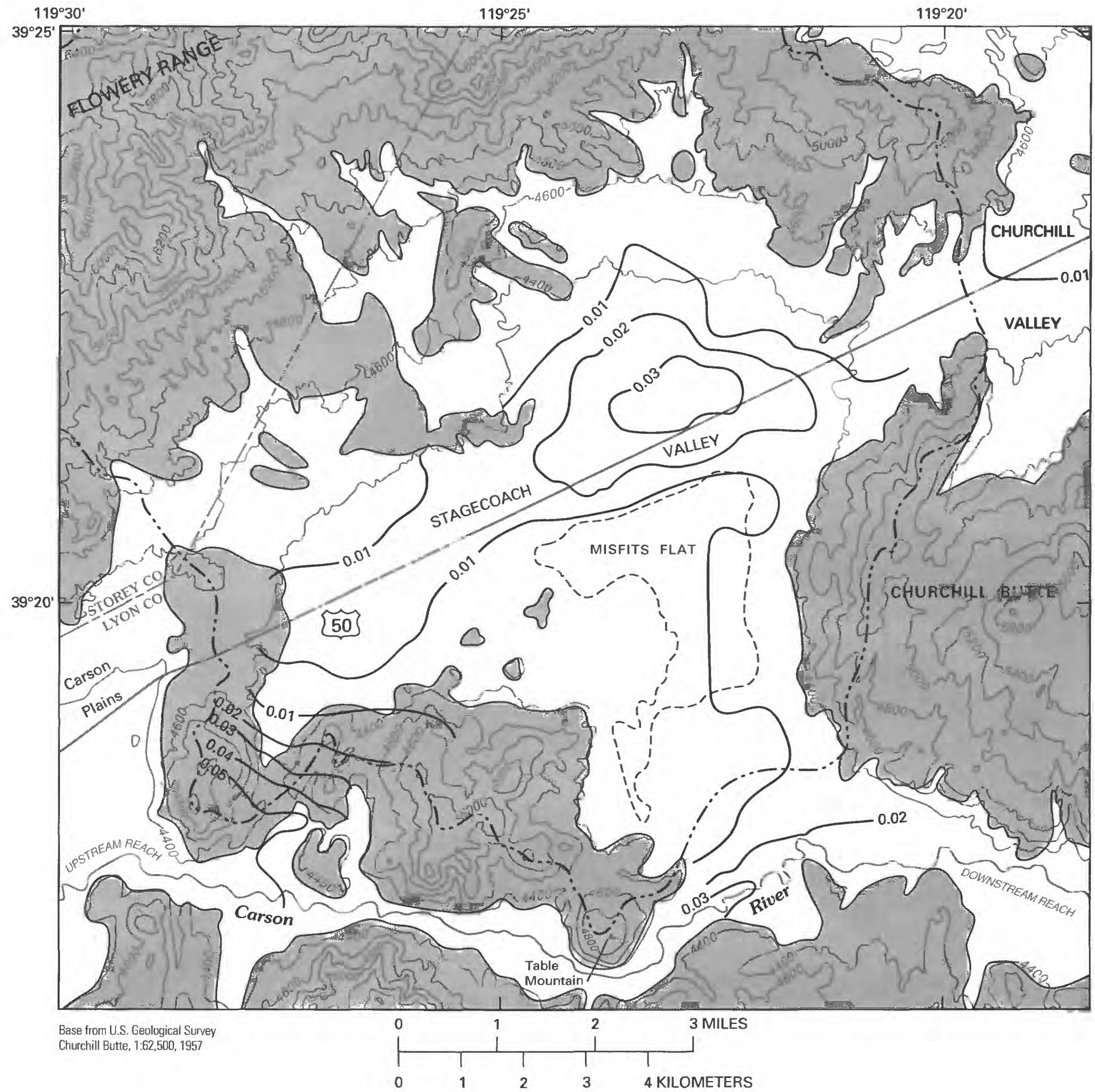
The distribution of leakance values between layers 1 and 2 (fig. 17) and between layers 2 and 3 (fig. 18) was simulated using the vertical hydraulic conductivities shown in figure 7. Values of vertical hydraulic conductivity for layer 3 were reduced by 5 to 15 percent to reflect the anticipated lower values of vertical hydraulic conductivity at depth. Distance between the midpoints of layers 1 and 2 was 150 ft, whereas distance between midpoints of layers 2 and 3 varied according to the thickness of layer 3. For two localized areas adjacent to Churchill Valley (figs. 17, 18), unrealistically high values of leakance were used to allow the outflow from a single constant-head cell to be integrated over a larger area.

STORAGE COEFFICIENT

The specific-yield values shown in figure 8 were used as the storage coefficients for layer 1. A uniform storage coefficient of 0.0003 was used for layer 2; rounded to one significant figure, it was estimated using the approximate relation given by Lohman (1972, p. 53) and a thickness for layer 2 of 250 ft. A uniform storage coefficient of 0.0007 was used for layer 3; it was estimated from the compressibility of water and an assumed average thickness of layer 3 of 700 ft. (The thickness of layer 3 varies from 0 to 3,000 ft; however, 700 ft probably is an acceptable average.)

EVAPOTRANSPIRATION OF GROUND WATER

Evapotranspiration of shallow ground water is approximated assuming a linear decrease from a maximum value, when the water table is at the land surface, to 0, when the water table is at the extinction depth (the depth at which significant evapotranspiration ceases). The value used for maximum evapotranspiration, 3.5 ft/yr, is about equal to the average annual lake evaporation in the area (Kohler and others, 1959, pl. 2). The extinction depth was set at 12 ft beneath areas of bare soil (such as most of Misfits Flat) and at 35 ft beneath stands of phreatophytes (on the basis of observed depth to water at the outer edges of such stands). A transition zone with extinction depths between 12 and 35 ft was used to represent a band of sparse phreatophytes around the margins of Misfits Flat.



EXPLANATION

BASIN FILL

CONSOLIDATED ROCK

— 0.02 — LINE OF EQUAL TRANSMISSIVITY OF LAYER 1—Interval, 0.01 feet squared per second

-··—··— TOPOGRAPHIC DIVIDE

FIGURE 14.—Model-derived transmissivity of model layer 1 (corresponding to upper 50 feet of saturated basin fill).

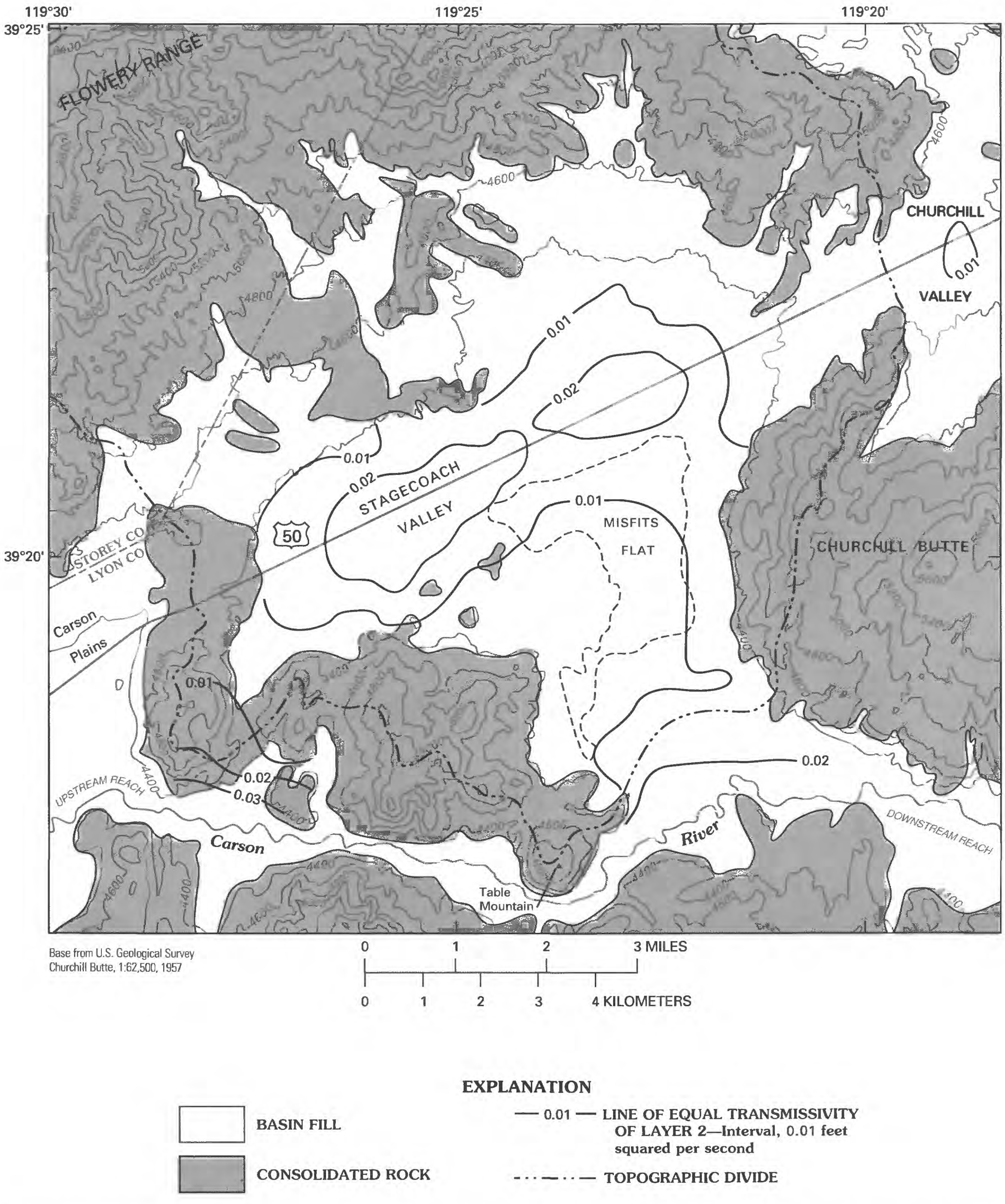


FIGURE 15.—Model-derived transmissivity of model layer 2 (corresponding to interval 50–300 feet below steady-state water table).

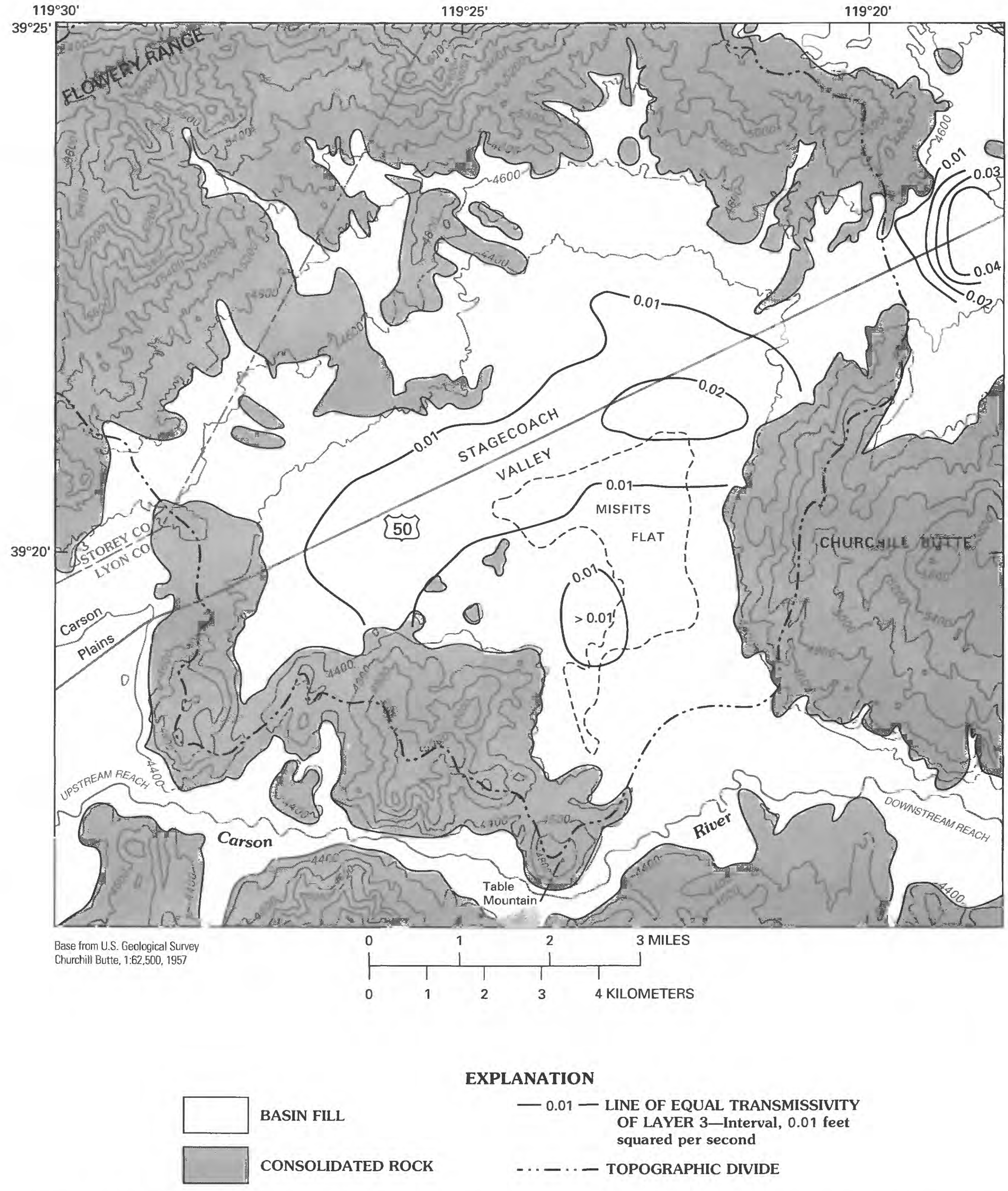
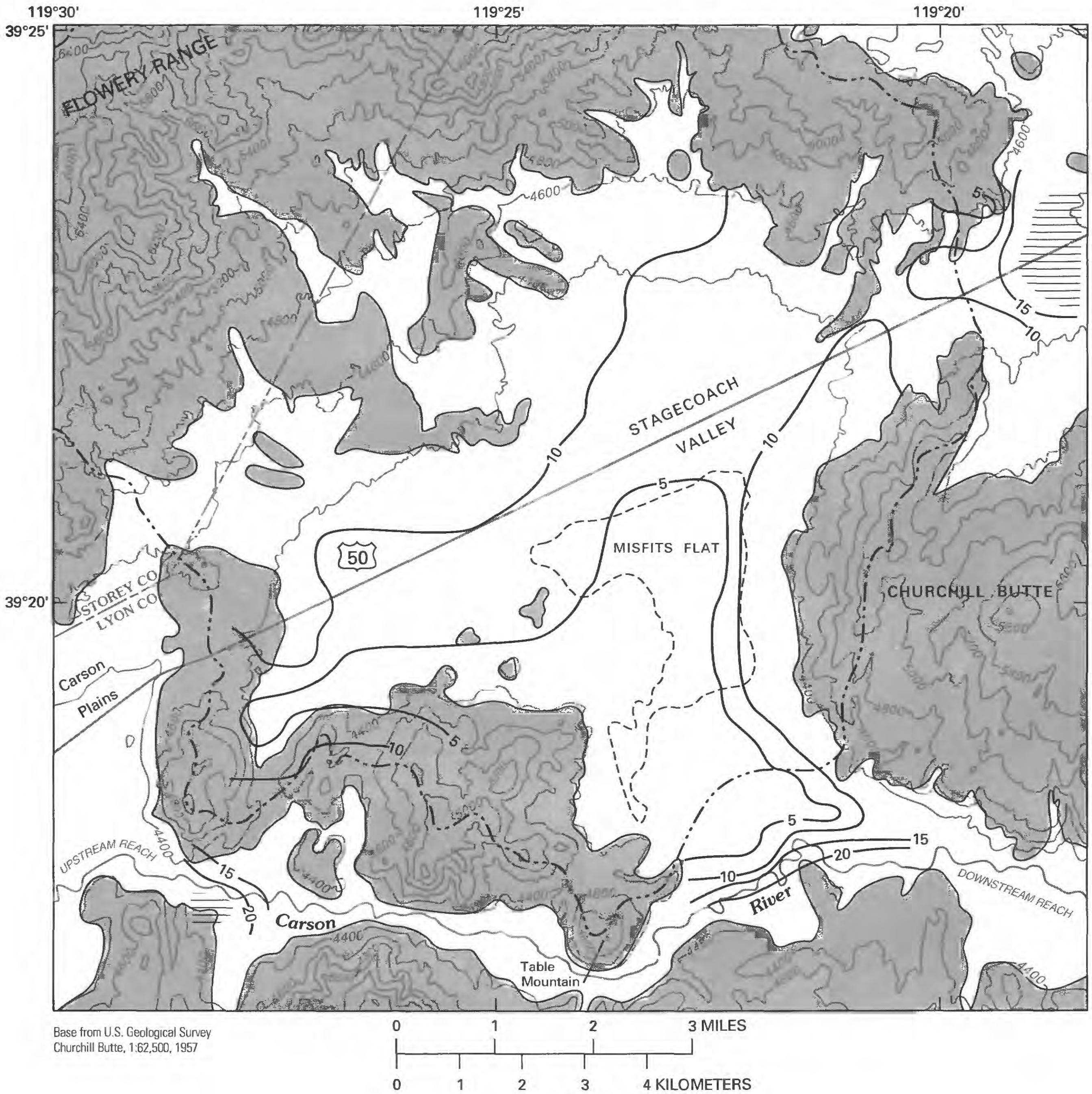


FIGURE 16.—Model-derived transmissivity of model layer 3 (corresponding to saturated deposits 300 feet below steady-state water table to bottom of basin fill).



EXPLANATION

BASIN FILL

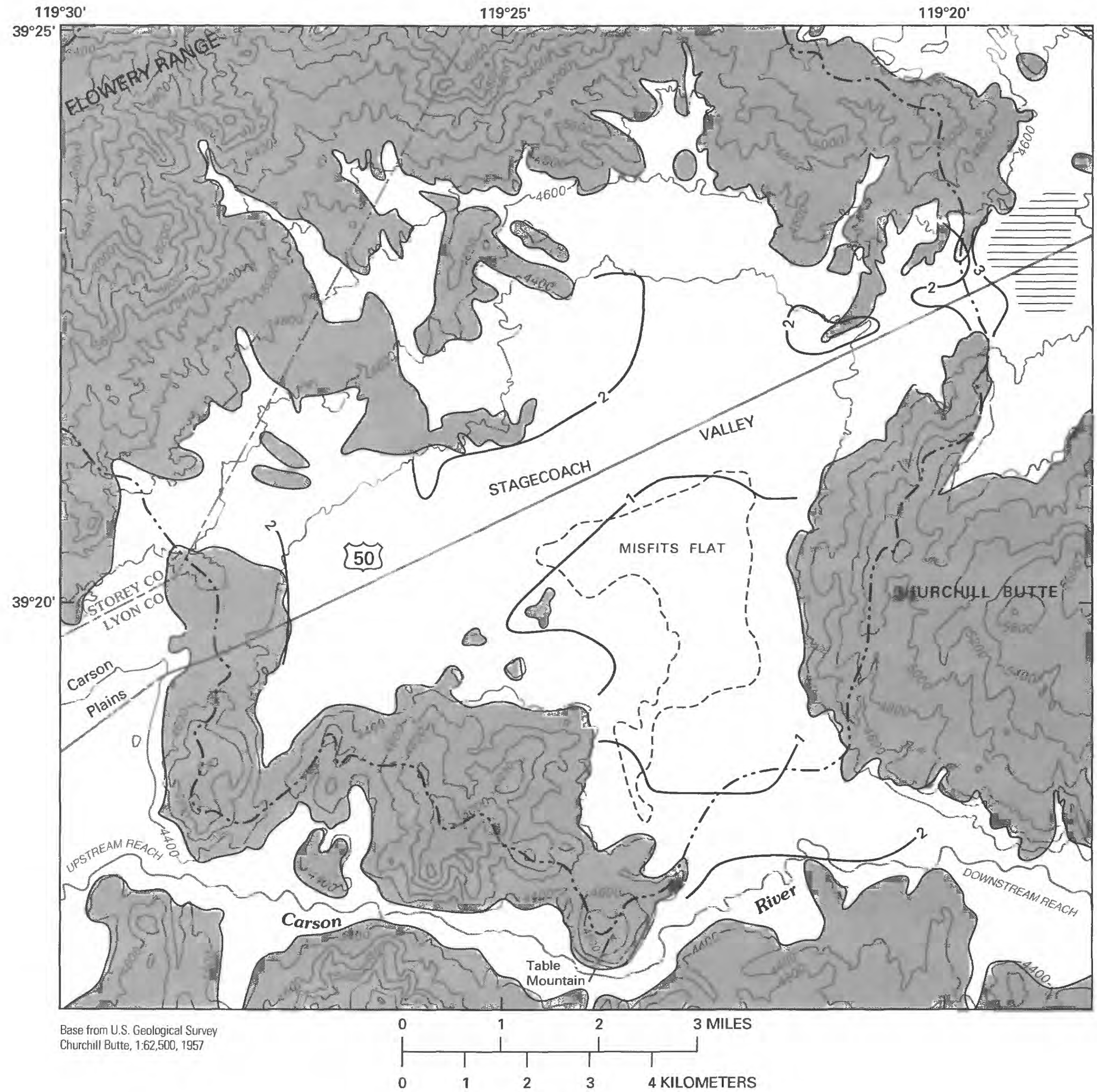
CONSOLIDATED ROCK

AREA WHERE UNREALISTICALLY HIGH LEAKANCE VALUES WERE USED—To allow subsurface inflow or outflow to be represented by single constant-head node

5 LINE OF EQUAL LEAKANCE BETWEEN LAYERS 1 AND 2—Unit is 1×10^{-10} feet per second. Interval, 5×10^{-10} feet per second

TOPOGRAPHIC DIVIDE

FIGURE 17.—Model-derived leakance between model layers 1 and 2.



EXPLANATION

BASIN FILL

CONSOLIDATED ROCK

AREA WHERE UNREALISTICALLY HIGH LEAKANCE VALUES WERE USED—To allow subsurface outflow to be represented by single constant-head cell

— 2 — LINE OF EQUAL LEAKANCE BETWEEN LAYERS 2 AND 3—Unit is 1×10^{-10} feet per second. Interval, 1×10^{-10} feet per second

· · — · · — TOPOGRAPHIC DIVIDE

FIGURE 18.—Model-derived leakance between model layers 2 and 3.

METHOD OF ANALYSIS OF SIMULATION RESULTS

If a reasonable match between observed and simulated conditions can be obtained, then within the limitations inherent in the simulation (Harrill, 1986, p. 26–28) the flow model is expected to provide information on the availability of ground water and probable hydrologic responses to future development. The basin-fill aquifer and its response to development were simulated in three phases: (1) the undeveloped condition, before 1971 (steady state), (2) the response to conditions from 1971 through 1981 (transient), and (3) probable long-term responses to selected future-development scenarios (transient). Analysis of the first phase provided insight about predevelopment basin-fill aquifer characteristics and was a means of evaluating parameters used in the model. The steady-state and transient simulations were calibrated so that computed heads, water-level changes, and distribution of evapotranspiration values agreed reasonably well with observed values. Transmissivity, vertical hydraulic conductivity, recharge, and subsurface inflow and outflow from the valley were adjusted to be within ranges compatible with field measurements or other available data mostly during the first-phase calibration. Storage coefficients and rate of recirculation from irrigation-return flow were adjusted mostly during the second-phase calibration. Once the fit between observed and simulated values was reasonable, we ran a series of simulations varying selected parameters to evaluate how sensitive the model was to errors in the values of the parameters.

Finally, during the third phase, a series of simulations were made to evaluate the probable long-term response to hypothetical pumping stresses associated with various development scenarios. The purpose was not to make specific predictions but rather to develop insight regarding the general response of the basin-fill aquifer to sustained (long-term) stresses.

RESULTS OF SIMULATION

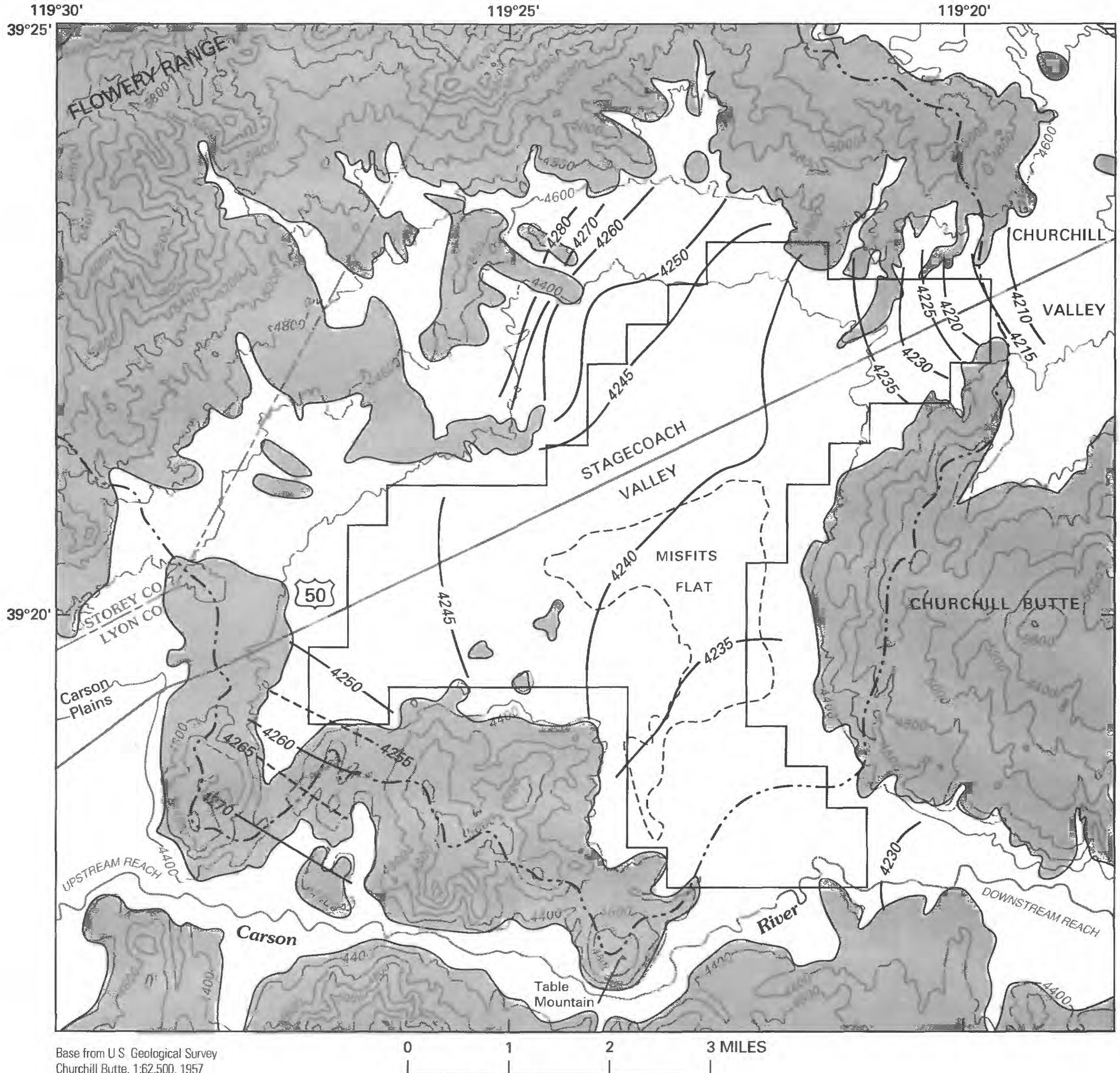
PREDEVELOPMENT CONDITIONS

Simulation of predevelopment conditions provides a means of analyzing the system under conditions of natural steady-state equilibrium. The simulated steady-state potentiometric surface for layer 2 of the model (fig. 19) is in reasonable agreement with the pre-1971 potentiometric surface based on field data (fig. 4). The potentiometric surfaces simulated for layers 1 and 3 are similar to the surface simulated for layer 2. The simulation allowed for downward flow in recharge areas and upward flow in discharge areas by incorporating small head differences between layers.

The closeness of fit between observed and simulated water levels was evaluated using two methods. The first method was to compare heads measured at 16 wells (in 1972 or earlier and in 1982) with model-derived heads computed for the cells that correspond to the areas in which the 16 wells are located (fig. 20). About half of the model-computed heads were within 5 ft of the measured heads. Part of this error probably is due to the fact that measured heads for wells located anywhere within a 2,000-ft by 2,000-ft cell were compared to the model-computed head at the center of the cell. In most cases this could cause a difference of 1 to 2 ft given the prevailing head gradient of the basin-fill aquifer. The second method was to select those model cells where there was adequate water-level information to make a reasonable estimate of the predevelopment head. In addition to the 16 wells that had measurements in 1972 or earlier, wells that had measurements in the mid-1970's provided supplemental control based on extrapolations of water-level trends back to 1971. In the model grid, a group of 120 contiguous cells was selected for comparison. About 70 percent of the 120 cells in layer 1 and about 60 percent of the 120 cells in layer 2 had differences between estimated (predevelopment) and simulated heads of 5 ft or less (fig. 20). Differences of this magnitude were considered reasonable, especially since land-surface altitudes at some wells were determined from topographic maps. The areas of poorest fit were along the northeast boundary of the model, where measurements indicated a steep hydraulic gradient toward Churchill Valley.

Values of recharge and discharge computed by the model are in good agreement with those estimated using empirical techniques. However, the model results indicate significant additional subsurface inflow and outflow.

Comparison of the modeled distribution of evapotranspiration values with the mapped extent of phreatophytes in Stagecoach Valley (excluding riparian areas shown in fig. 4) provides an additional check on the ability of the model to simulate the basin-fill aquifer of Stagecoach Valley (fig. 21). Areas where phreatophyte vegetation was mapped but where no evapotranspiration was calculated by the model occur where the depth to water calculated by the model exceeds 35 ft. Adjustments for a closer fit were not attempted because the error is



EXPLANATION

BASIN FILL

CONSOLIDATED ROCK

—4240— SIMULATED STEADY-STATE WATER-LEVEL CONTOUR IN LAYER 2 OF MODEL—Dashed where approximately located. Contour interval, in feet, is variable. Datum is sea level

——— BOUNDARY OF 120-CELL AREA—Area used to compare simulated heads to observed heads. Results of comparison are shown in figure 20

-··—··— TOPOGRAPHIC DIVIDE

FIGURE 19.—Simulated steady-state potentiometric surface for model layer 2.

probably a combined effect of land-surface altitudes determined at the centers of the cells, assigned extinction depths for evapotranspiration, and aquifer properties assigned to layer 1.

RESPONSE TO DEVELOPMENT

From spring 1971 to spring 1982, more than 11,000 acre-ft of water was pumped from the basin-fill aquifer of Stagecoach Valley. Simulation of this 11-yr period of pumping has indicated a response that includes water-level declines, reductions in natural evapotranspiration, changes in subsurface inflow and outflow, and depletion of ground water in storage.

WATER-LEVEL DECLINES

Water-level data from shallow wells are not adequate for a detailed comparison of measured and simulated water-table fluctuations in layer 1; however, short-term records available at several wells indicate net declines of the same general magnitude as the simulated declines shown in figure 22. Hydrographs of three shallow wells for the period 1977–82 are plotted in figure 23 for comparison with simulated water-level altitudes in the model cell that contains each well. The fluctuations in the observed data reflect seasonal variation in pumping rates that were not simulated in the model; consequently, only the trends are comparable.

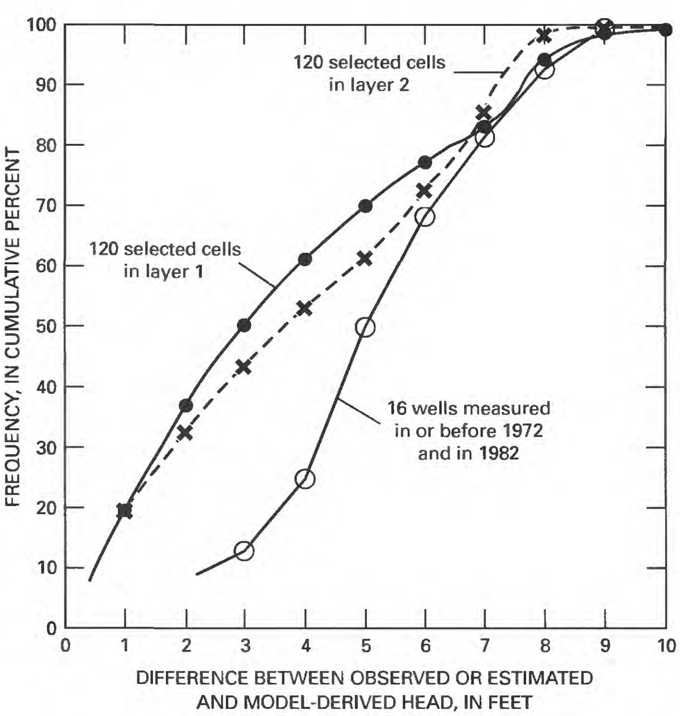


FIGURE 20.—Cumulative-frequency distributions showing closeness of fit between observed or estimated heads and model-derived heads.

For the period spring 1971 to spring 1982, the simulated water-level declines for layer 2 of the model (fig. 24) are in general agreement with the observed declines (fig. 11). Any differences probably are due to the observed declines being measured at individual wells, whereas the simulated declines represent values averaged over a 2,000-ft by 2,000-ft finite-difference cell. Hydrographs of five wells that penetrated more than 50 ft of saturated basin fill (into the equivalent of layer 2), based on periodic measurements during 197182, are plotted in figure 25; also shown are simulated water-level changes in the cells that contain each well so that trends can be compared.

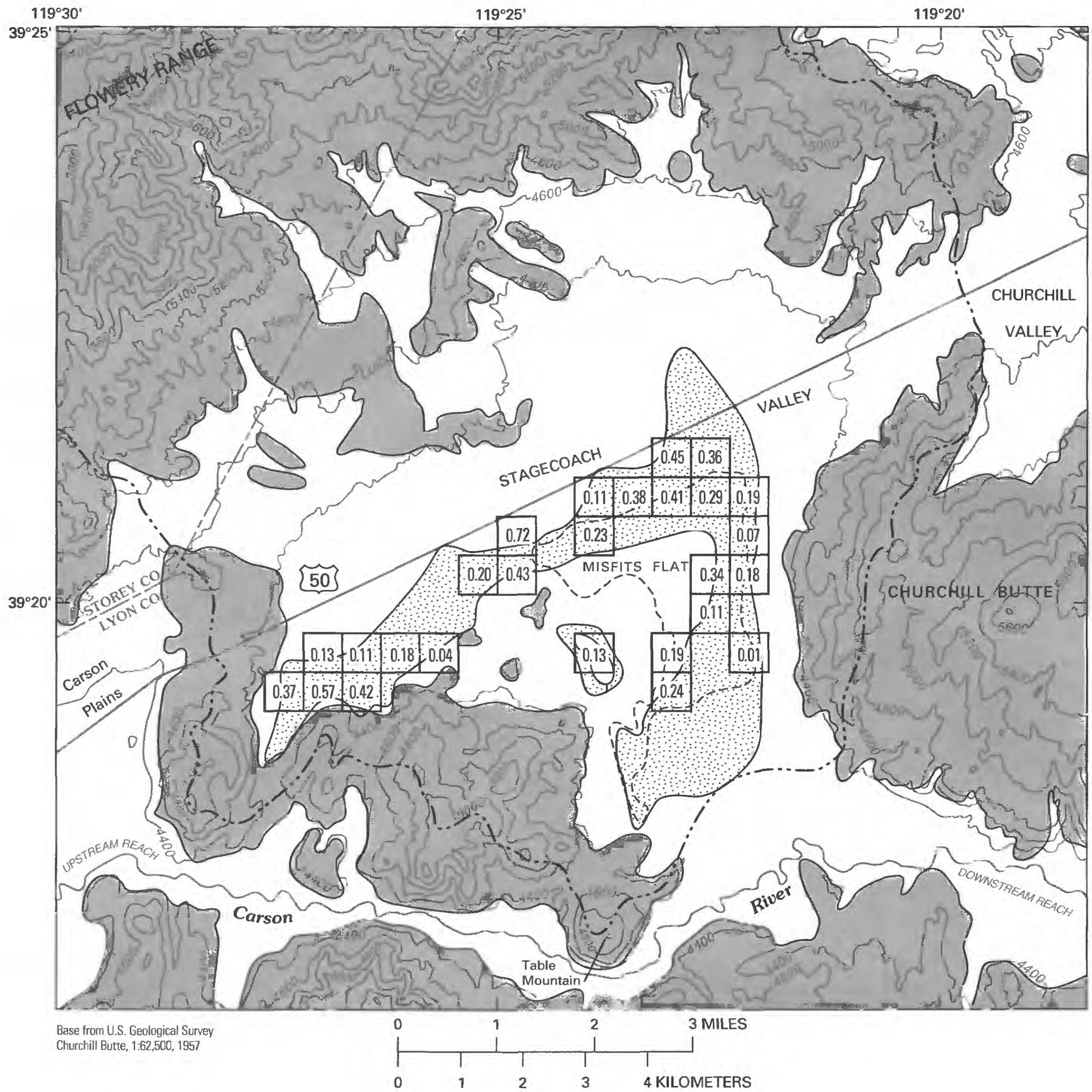
Simulated water-level declines in layer 3 for the same period were less than those in layer 2 and generally were distributed over larger areas. There are no observation wells that penetrate the basin fill to depths represented by layer 3 of the model.

REDUCTIONS IN EVAPOTRANSPIRATION

The simulated annual evapotranspiration of ground water decreased from about 640 to about 180 acre-ft/yr from 1971 to 1982. This reflects the decline of the water table by pumping. Field observations in 1982 and 1983 indicated that the areal extent of phreatophytes was about the same as had been mapped in the early 1970's. However, the general vigor and areal density of the plants appeared slightly less than before 1971. The simulated change could not be precisely verified because decreased consumption of ground water may have been partly offset by increased consumption of soil moisture as the plants responded to stresses caused by declining water tables.

CHANGES IN INFLOW AND OUTFLOW

The 11 pumping seasons provided sufficient time to affect all head-dependent fluxes at the margins of the area simulated in the ground-water flow model. The simulated changes are listed in table 6. As with evapotranspiration, these changes in subsurface flow represent water captured by pumping in Stagecoach Valley. They could not be verified by field observations, but they are comparable to the general magnitude of observed water-level declines



EXPLANATION

BASIN FILL

CONSOLIDATED ROCK

AREA OF PHREATOPHYTE VEGETATION — Primarily greasewood (P.A. Glancy, U.S. Geological Survey, written commun., 1982). Predevelopment (pre-1971) conditions

0.29 CELL IN LAYER 1 WHERE EVAPOTRANSPIRATION WAS CALCULATED BY MODEL—Value is rate, in feet per year

TOPOGRAPHIC DIVIDE

FIGURE 21.—Observed and simulated areas of evapotranspiration.

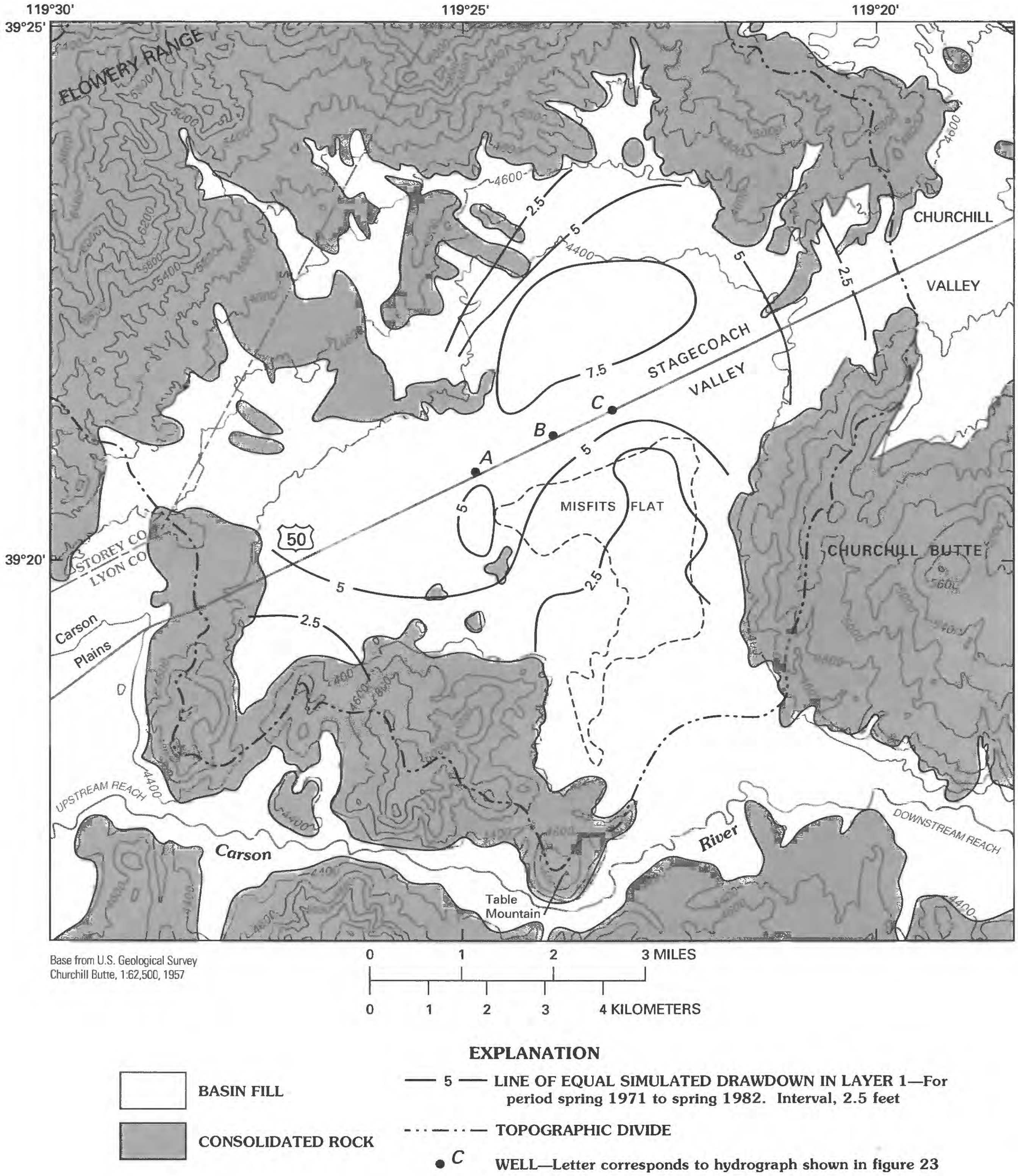


FIGURE 22.—Simulated water-table decline in model layer 1, spring 1971 to spring 1982.

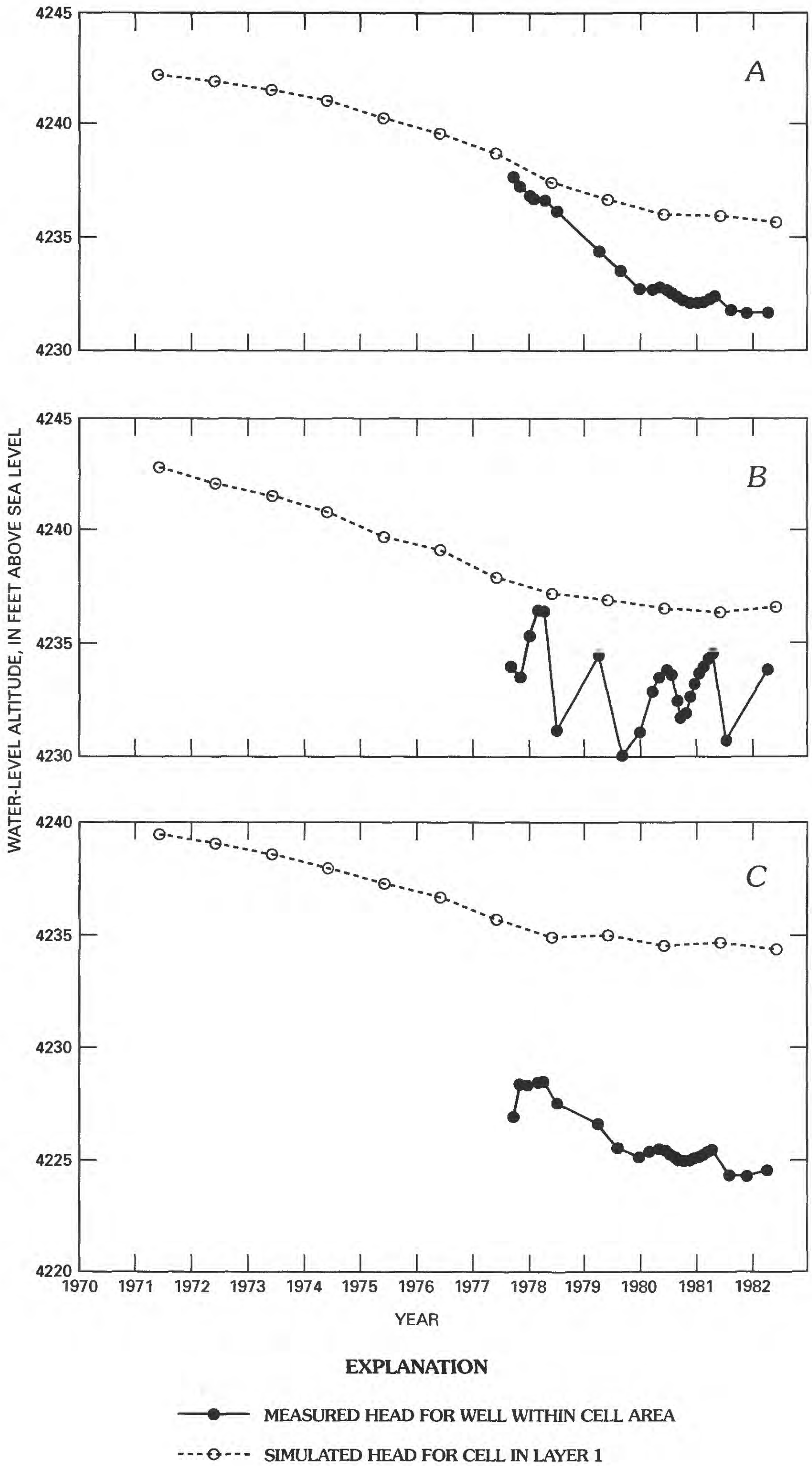


FIGURE 23.—Comparison of simulated and observed hydrographs for three wells (fig. 22) completed at depths corresponding to model layer 1. (See fig. 13 for layer 1 cell locations.) *A*, Well in SW¼ sec. 9, T. 17 N., R. 23 E.; well depth 82 feet and top of open interval at 52 feet; cell (11,11) in layer 1 used for simulation. *B*, Well in SE¼ sec. 9, T. 17 N., R. 23 E.; well depth 84 feet and top of open interval at 82 feet; cell (10,13) in layer 1 used for simulation. *C*, Well in NE¼ sec. 10, T. 17 N., R. 23 E.; well depth 88 feet and top of open interval at 86 feet; cell (10,15) in layer 1 used for simulation.

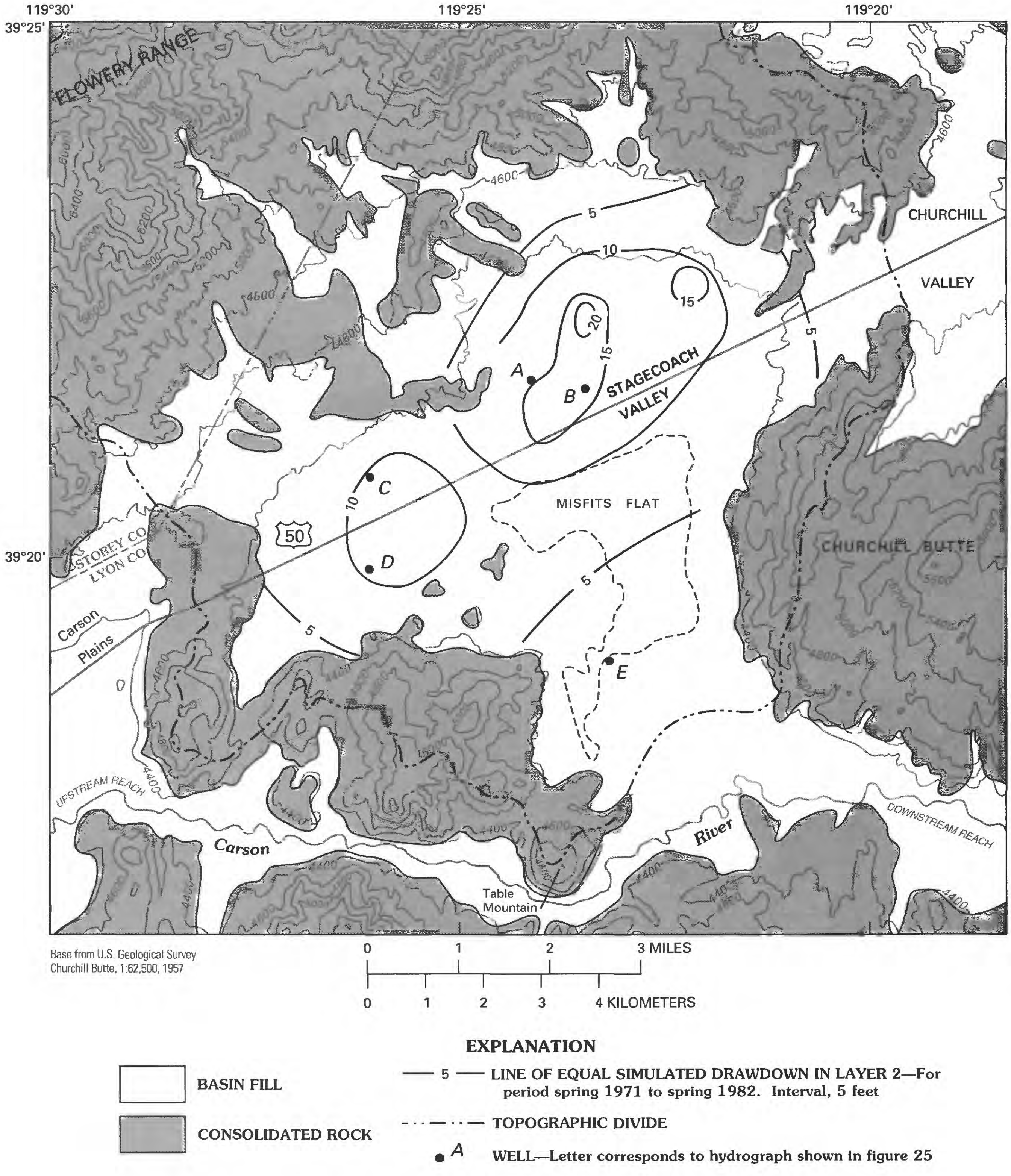


Figure 24.—Simulated water-level decline in layer 2, spring 1971 to spring 1982.

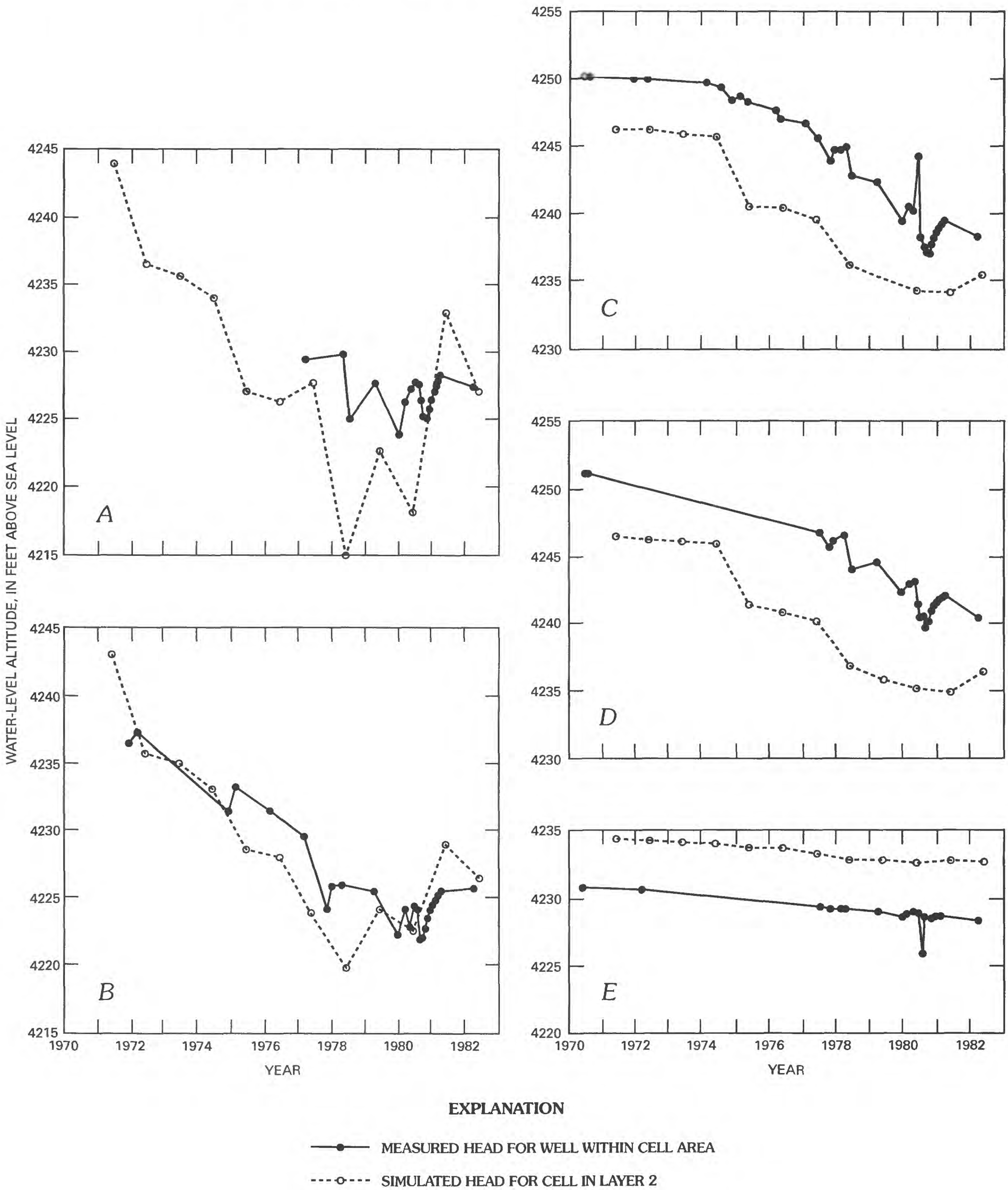

A
B
C
D
E
WATER-LEVEL ALTITUDE, IN FEET ABOVE SEA LEVEL
YEAR
1970
1972
1974
1976
1978
1980
1982
4215
4220
4225
4230
4235
4240
4245
4250
4255
EXPLANATION
MEASURED HEAD FOR WELL WITHIN CELL AREA
SIMULATED HEAD FOR CELL IN LAYER 2

TABLE 6.—*Simulated net changes in inflow and outflow from spring 1971 to spring 1982*

[Values that are in acre-feet per year are rounded]

	Rate (acre-feet per year)		Change in rate, 1971-82	
	Pre-1971	Spring 1982	(acre-feet [percent])	
Inflow				
From upstream reach, Carson River.	280	300	20	[7]
From downstream reach, Carson River.	90	100	10	[11]
Outflow				
To Churchill Valley	170	160	−10	[−6]
To downstream reach, Carson River.	120	110	−10	[−8]

in observation wells along the margins of the area, as modeled. These changes are small in relation to the flow of the Carson River, whose mean annual flow is about 380 ft^3/s and mean flow in September, the month with the lowest mean discharge, about 10 ft^3/s (Schroer and Moosburner, 1978, p. 212). The total capture of water from the river (increase in inflow plus decrease in outflow, as indicated in table 6) is equivalent to a flow in the river of only about 0.05 ft^3/s.

GROUND-WATER BUDGETS

Ground-water budgets listed in table 7 summarize the inflow to and outflow from the basin-fill aquifer of Stagecoach Valley for predevelopment (pre-1971) conditions and for conditions during the period spring 1981 to spring 1982. Two sets of figures were used for long-term natural conditions; one was estimated using field observations and analysis of empirical data, and the other was calculated using the calibrated steady-state model. Most of the model results are at least one-third larger than the empirically estimated values. Most of this difference is in values of subsurface inflow and outflow, which in several cases were considered negligible in the budget on the basis of field observations and analysis of empirical data.

The 11 pumping seasons from spring 1971 through spring 1982 had discernible effects on the water budget for the basin-fill aquifer. During the 1981 pumping season about 1,100 acre-ft/yr was being pumped, of which about 150 acre-ft was being recirculated back to the basin-fill aquifer as irrigation-return flow (table 7). This resulted in a net pumpage of about 950 acre-ft/yr. About 48 percent of the net pumpage was supplied by reductions in evapotranspiration, about 3 percent by increased subsurface inflow from the Carson River, and about 2 percent by reduced subsurface outflow to Churchill Valley and Carson River (downstream reach). The remaining 47 percent of the net pumpage was supplied from ground-water storage, causing the declines in ground-water levels shown in figure 11.

SENSITIVITY ANALYSIS

Twenty model simulations were made to determine the sensitivity of model results to the uncertainty in the determination of five parameters. Consecutively each of the five parameters was varied by 20 percent above and below its final calibrated value while the other four parameters were kept at the calibrated values (table 8). The model then was run for steady-state conditions, and the average of the absolute differences between simulated and measured values of head was determined. The average of the absolute differences obtained with the varying parameter then was compared with the average of the absolute differences that existed in the calibrated steady-state model. The extent to which the average difference of head changed in response to the varying parameter was used as an index of sensitivity. The head-dependent fluxes at the boundaries were evaluated the same way and compared with fluxes computed by the

FIGURE 25.—Comparison of simulated and observed hydrographs for five wells (fig. 24) completed at depths corresponding to model layer 2. (See fig. 13 for layer 2 cell locations.) *A*, Well in SE¼ sec. 4, T. 17 N., R. 23 E.; well depth 339 feet and top of open interval at 287 feet; cell (9,13) in layer 2 used for simulation. *B*, Well in NW¼ sec. 10, T. 17 N., R. 23 E.; well depth 300 feet and top of open interval at 234 feet; cell (9,14) in layer 2 used for simulation. *C*, Well in SE¼ sec. 7, T. 17 N., R. 23 E.; well depth 386 feet and top of open interval at 12 feet; cell (12,8) in layer 2 used in simulation. *D*, Well in SE¼ sec. 18, T. 17 N., R. 23 E.; well depth 273 feet and top of open interval at 137 feet; cell (14,8) in layer 2 used for simulation. *E*, Well in NE¼ sec. 27, T. 17 N., R. 23 E.; well depth 220 feet and top of open interval at 180 feet; cell (17,15) in layer 2 used for simulation.

TABLE 7.—*Predevelopment (pre-1971) and 1981–82 ground-water budgets based on field observations and empirical analysis and on steady-state and transient simulations*

[All data given in acre-feet per year. Negative values indicate that water is being added to storage. All values are rounded to two significant figures. —, negligible or not applicable]

	Predevelopment conditions		Spring 1981 to spring 1982 conditions, based on transient simulation
	Based on field observaions and empirical analysis	Based on steady-state simulation	
Inflow			
Recharge from precipitation	440–580	550	550
Inflow from Carson River (upstream reach)	130–170	280	300
Inflow from Carson River below Table Mountain (downstream reach).	—	86	98
Recirculated from irrigation return	—	—	150
Total	570–750	920	1,100
Outflow			
Evapotranspiration	700	630	180
Pumpage	—	—	1,100
Outflow to Churchill Valley	—	170	160
Outflow to Carson River (downstream reach)	—	120	110
Total	700	920	1,600
Net results			
Net outflow-inflow	–50 to 130	—	450
Storage depletion (simulated by model)	—	—	460

calibrated model. The change in flux was also used as an index of sensitivity. These changes are summarized in table 8.

A general conclusion that can be drawn from this sensitivity analysis is that the head distribution simulated by the model is not highly sensitive to moderate uncertainties in values of hydrologic parameters except the evapotranspiration extinction depth. Changes of ±20 percent in the other four hydrologic parameters generally produced a change in head of less than 15 percent. The high sensitivity to evapotranspiration extinction depth is probably because small changes in head in the center of the valley, caused by changes in extinction depth, also cause changes in inflow to and outflow from the Carson River. Moreover, the initial quantities of inflow and outflow are small, so minor changes in flow rates could cause large percentage changes. Also, boundary fluxes calculated by the model are generally more sensitive than calculated heads to uncertainties in values of the model parameters. Consequently, a close fit between observed and simulated heads does not guarantee accuracy. For instance, estimates of flow to and from the Carson River and flow to Churchill Valley may be somewhat in error because of moderate errors in determining key hydrologic parameters such as transmissivity.

LIMITATIONS OF THE MODEL

The digital computer model is a simplified representation of the basin-fill aquifer of Stagecoach Valley. It was calibrated for predevelopment steady-state conditions on the basis of sparse data available in the early 1970's and for transient pumping conditions during a relatively short period of 11 pumping seasons. Limited hydrologic data are available for the area, so hydrologic conditions had to be inferred for some parts of the aquifer, especially beneath the playa, along the south and east margins of the valley, and at depths greater than about 350 ft in the basin fill.

The model-simulated head distributions and fluxes were similar to those measured during the period spring 1971 to spring 1982. If the well distribution and associated pumping rates remained about the same as during the calibration period, then simulations for relatively short periods (5–10 yr) of future pumping would indicate water-level declines and flux changes with the same degree of accuracy as calculated for the 1971 to 1982 period.

TABLE 8.—*Summary of model-sensitivity runs*

[ET, evapotranspiration rate; —, none]

Parameter varied	Part of model affected	Change applied to parameter (percent)	Difference between observed and simulated head[1] (feet (percent change)): Layer 1	Difference between observed and simulated head[1] (feet (percent change)): Layer 2	Flux at head-dependent boundaries (cubic feet per second (percent change)): Upstream reach Carson River, net flux	Downstream reach Carson River, net flux	Outflow to Churchill Valley	Evapotranspiration
Steady-state calibration								
—	All layers	—	3.63 (—)	3.84 (—)	0.39165 (—)	−0.04200 (—)	0.22869 (—)	0.87756 (—)
Sensitivity analysis								
Transmissivity:	All layers	+20	3.43 (5.5)	3.72 (3.1)	0.46271 (18.1)	−0.04888 (16.4)	0.25303 (10.6)	0.91683 (4.5)
		−20	3.93 (8.3)	4.10 (6.8)	.31889 (18.6)	−.03568 (15.0)	.20115 (12.0)	.83809 (4.5)
	Layer 1	+20	3.45 (5.0)	3.66 (4.7)	.44668 (14.1)	−.04105 (2.3)	.23436 (2.5)	.91075 (3.8)
		−20	3.84 (5.8)	4.08 (6.2)	.33568 (14.3)	−.04293 (2.2)	.21544 (5.8)	.83430 (4.9)
	Layer 2	+20	3.60 (.8)	3.87 (.8)	.40757 (4.1)	−.05036 (19.9)	.23293 (1.8)	.88083 (.4)
		−20	3.68 (1.4)	3.83 (.3)	.37493 (4.3)	−.03390 (19.3)	.22437 (1.9)	.87351 (.5)
	Layer 3	+20	3.62 (.3)	3.82 (.5)	.39191 (.1)	−.04086 (2.7)	.23761 (3.9)	.87351 (.5)
		−20	3.65 (.6)	3.87 (.8)	.39138 (.1)	−.04307 (2.5)	.21925 (4.1)	.88517 (.9)
Vertical hydraulic conductivity.	All layers	+20	3.51 (3.3)	3.71 (3.4)	.39520 (.9)	−.04136 (1.5)	.23990 (4.9)	.87018 (.8)
		−20	3.84 (5.8)	4.09 (6.5)	.38727 (1.1)	−.04302 (2.4)	.21352 (6.6)	.88766 (1.2)
	Interface between layers 1 and 2.	+20	3.61 (.6)	3.78 (1.6)	.39494 (.8)	−.04035 (3.9)	.22900 (.1)	.88219 (.5)
		−20	3.67 (1.1)	3.93 (2.3)	.38756 (1.0)	−.04400 (4.8)	.22817 (2.3)	.87213 (.6)
	Interface between layers 2 and 3.	+20	3.53 (2.8)	3.76 (2.1)	.39191 (.1)	−.04300 (2.4)	.23596 (3.2)	.86566 (1.4)
		−20	3.80 (4.7)	4.00 (4.2)	.39137 (.1)	−.04099 (2.4)	.21397 (6.4)	.89338 (1.8)
Recharge rate:	Layer 1	+20	4.06 (11.8)	4.34 (13.0)	.38869 (.8)	−.04817 (14.7)	.23768 (3.9)	1.01090 (15.2)
		−20	3.31 (8.8)	3.48 (9.4)	.39485 (.8)	−.03554 (15.4)	.21986 (3.9)	.74851 (14.7)
Maximum ET rate.	Layer 1	+20	3.61 (.6)	3.76 (2.1)	.39660 (1.3)	−.03898 (7.2)	.22682 (.8)	.88750 (1.1)
		−20	3.67 (1.1)	3.96 (3.1)	.38504 (1.7)	−.04608 (9.7)	.23116 (1.1)	.86443 (1.5)
Maximum ET depth.	Corresponding depth in layer 1.	+20	5.13 (41.3)	4.48 (16.7)	.44411 (13.4)	.02927 (169.7)	.20500 (10.4)	1.02440 (16.7)
		−20	4.97 (36.9)	5.56 (44.8)	.33272 (15.0)	−.09627 (129.2)	.25222 (10.3)	.73898 (15.8)

[1]Average of absolute difference between observed-head distribution and simulated-head distribution in 120-cell area of layers 1 and 2 where there were enough data to make reliable estimates of predevelopment water levels (fig. 17).

Longer periods of pumping and different well distributions and pumping rates can be evaluated, but the results should be considered reliable only in terms of the general response of the basin-fill aquifer to the pumping. Actual pumping for long periods of time may produce aquifer responses different from those simulated and discussed in this report; however, the general changes and trends should be similar.

SIMULATED RESPONSE TO HYPOTHETICAL DEVELOPMENT SCENARIOS

The general response of the basin-fill aquifer of Stagecoach Valley to nine hypothetical pumping scenarios was simulated and then evaluated using the calibrated flow model. Model simulations were made for an arbitrary period of 600 yr, 300 yr of pumping and 300 yr of recovery to allow the flow system to respond to sustained stresses and to approach a new equilibrium by the end of the 600 yr of pumping and recovery. In the model, transmissivity of the top layer was arbitrarily held constant instead of being allowed to vary in response to changing water levels. This was done to avoid difficulties with model cells going dry (during some of the scenarios that involved heavy pumping) and then becoming saturated again during subsequent periods of recovery. To test the effect of assuming a constant transmissivity, the 11 pumping seasons, from 1971 to 1982, were also simulated without varying transmissivity. Drawdowns in the most heavily pumped cells were within 1 to 2 ft of the calibrated drawdowns (figs. 22, 24). Drawdown in areas distant from the heavily pumped wells agreed closely with drawdown obtained during calibration. Consequently, for moderate drawdown, the assumption of constant transmissivity is valid for the purpose of this report. However, some pumping scenarios involved great pumping rates, and large drawdowns were simulated for some areas near pumping cells. In two scenarios, the localized drawdown exceeded 100 ft but decreased rapidly away from pumping centers. In these scenarios, the simulated drawdown in the vicinity of pumping centers may be erroneously small. However, these errors would be restricted to localized areas and the assumption of constant transmissivity for layer 1 probably is acceptable, as a first approximation, for evaluation of valleywide response to the pumping scenarios.

Hypothetical pumping scenarios were constrained using the following limitations:

1. Pumping wells were not located where fine-grained deposits have low transmissivity, where the predevelopment depth to water exceeded 200 ft, where the thickness of saturated basin fill is less than 200 ft, where model cells are bounded on two or more sides by consolidated rock, or where land-surface slopes are greater than 200 ft/mi. Stagecoach Valley is partly drained by subsurface outflow, and no significant areas of saline water are known to be present; consequently, water quality was not used as a constraint in siting wells.
2. Pumping cells were distributed according to two general strategies. One is to distribute pumping cells strategically in or adjacent to areas of large estimated evapotranspiration to facilitate capture of the water consumed by natural evapotranspiration. This strategy is termed "strategically distributed." The other strategy is to form a single pumping center where pumping is concentrated in a few specific cells that are adjacent to each other. This strategy is termed "concentrated."
3. To avoid the necessity of quantifying and simulating the rate of irrigation-return flow from pumped water, all model simulations were based on net pumpage.
4. The range of pumping rates assigned to wells in an individual model cell was constrained. The maximum allowable rate prevented massive overpumping in individual cells, and the minimum rate avoided assignment of unrealistically low pumpage in numerous individual cells. The maximum pumping rate for each pumping cell was estimated on the assumption that a draft of about 3 ft/yr of water is evenly applied to the entire cell area. This rate of 3 ft/yr is the average rate of consumption by irrigation in the valley. Four pumping cells (three pumping at the maximum rate and one pumping at about half the maximum rate) are required to withdraw water at a total rate equal to the estimated predevelopment inflow and outflow (920 acre-ft/yr). The minimum rate for each pumping cell was about half of the maximum rate; therefore, seven pumping cells are required to withdraw water at a total rate equal to the estimated predevelopment inflow and outflow. In the model, all pumping was from layer 2.

The nine simulated development scenarios (A through I), grouped in table 9 by strategic distribution of pumping cells, were selected to illustrate how the general response of the basin-fill aquifer of Stagecoach Valley might change under a variety of imposed pumping stresses. Scenarios A, C, D, F, and H also provide a general test of the feasibility of the concept of sustained yield, in which pumping

TABLE 9.—*Simulated development scenarios*

Distribution of pumping cells	Scenario	Pumping rate (acre-feet per year)	Drawdown time (years)	Recovery time (years)
Pumping cells are strategically located to efficiently capture evapotranspiration.	A	920	300	300
	B	1,840	300	300
	C[1]	1,840	50	300
		920	250	
Pumping cells are concentrated and centrally located in relation to area of ground-water evapotranspiration.	D	920	300	300
	E	1,840	300	300
Pumping cells are concentrated at north end of Stagecoach Valley.	F	920	300	300
	G	1,840	300	300
Pumping cells are concentrated in south-eastern part of Stagecoach Valley, near Carson River.	H	920	300	300
Pumping cells have same general distribution (and pumping rates) as 1981 pumping wells.	I	970	300	300

[1]Scenario C incorporates two phases of pumping (a 50-year phase succeeded by a 250-year phase) characterized by different pumping rates; 300-year recovery begins after 300-year drawdown.

is maintained at a rate equal to the capture, which is the increase in recharge plus the decrease in discharge (Lohman and others, 1972). The scenarios for Stagecoach Valley test the sustained-yield concept in situations where not only may water be captured from more than one source but also where additional water may actually be induced to flow into the area. The scenarios were also selected to illustrate how the location of pumping wells affects the long-term aquifer response to development.

Scenarios B, E, and G illustrate the aquifer response to pumping rates substantially greater than the average annual recharge. Scenario I was included so that the general long-term response to a development pattern similar to the 1981 pumping pattern could be compared with the simulated scenarios. In all nine scenarios, neither economic and legal consideration nor possible salt buildup in the soil, caused by irrigation, was considered.

SIMULATION RESULTS

Simulation results of the various development scenarios are presented in figures 26 to 34 and in tables 10 to 18. (These figures and tables follow the "References Cited" section of this report.) Simulation of long-term response of the aquifer is only intended to show the nature of the response; therefore, only those factors that indicate general trends are presented in the tables and figures. These factors are areal distribution of pumping cells and associated drawdown after 300 yr of pumping; changes in average drawdown of pumping cells, storage, evapotranspiration, and subsurface inflow and outflow during 300 yr of pumping succeeded by 300 yr of recovery; and the sources of pumped water after specified periods of pumping (ranging from 1.5 to 300 yr).

The areal distribution of cells with associated pumping rates and the distribution of drawdown in layer 2 after 300 yr of pumping (shown in part *A* of figs. 26–34) are used to illustrate the general relation between distributed pumping and areal patterns of drawdown. Drawdowns in layers 1 and 3 have similar patterns, but their magnitude is generally less.

In the graphs showing changes with time (parts *B* through *E* of figs. 26–34), the rate of change of each factor probably is more significant than a particular value at any given time. Generally, rapid changes suggest that the aquifer is strongly out of equilibrium and indicate that most of the pumped water is being derived from storage. On the other hand, small changes with time suggest that the aquifer may be gradually approaching a new equilibrium when little or no change in storage will occur. The potential sources of pumped water are shown in the pie diagrams (part *F*) of figures 26 to 34.

Ground-water budgets (tables 10–18) are used to describe the overall hydrologic effects after 1.5, 25, 50, 100, and 300 yr of pumping. A summary of predevelopment conditions is included in each budget table for comparison.

DISCUSSION OF RESULTS

One of the objectives of the study of the basin-fill aquifer of Stagecoach Valley was to test the general feasibility of sustained yield. A comparison of the aquifer responses to the nine development scenarios is presented in table 19. Results of scenarios A, C, D, F, and H (figs. 26, 28, 29, 31, and 33, respectively) indicate that sustained yield is possible in Stagecoach Valley. In scenarios A, D, F, and H, where net pumping rate was held equal to the simulated rate of predevelopment total inflow to or total outflow from the basin-fill aquifer (920 acre-ft/yr), the model reached or almost reached a new equilibrium condition by the end of the 300-yr simulation. Siting (location and concentration) of pumping wells played a major role in determining the magnitude of the drawdown and what sources of water would be most readily captured; however, none of the scenarios used in this analysis produced adverse effects severe enough to prevent the aquifer from approaching a new equilibrium. A significant assumption in the simulations was that the Carson River added "new" water to the basin-fill aquifer in proportion to the amount of drawdown in the river cells; this balanced inflow and the small size of Stagecoach Valley were significant factors in the aquifer's closely approaching a new equilibrium. If the aquifer had been bounded entirely by no-flow boundaries in the model, the results would have been substantially different. The assumed inflow of water from the boundaries plus the small size of the area facilitated a complete recovery within the 300-yr period after pumping had ceased.

Scenarios B, E, and G illustrate the response of the aquifer to a rate of pumping twice the estimated average annual recharge or discharge (table 19; figs. 27, 30, 32). Drawdowns were from two to six times as great as those due to pumping at the lower rate of 920 acre-ft/yr; however, the aquifer was still nearing a new equilibrium in each of the three scenarios. This was possible only because of "new" water brought into the aquifer, either from induced infiltration through deposits along the Carson River or by capturing ground water that formerly flowed toward Churchill Valley. In this regard the limiting factor on sustained development

TABLE 19.—*Comparison of aquifer responses under nine different development scenarios*

[do. or Do., ditto; ET, areas of known evapotranspiration]

Scenario	Pumping rate (acre-feet per year)	General location of pumping cells	Spacing of pumping cells	Maximum water-level decline in a cell (feet)	Recovery[1] after 300 years of pumping	New equilibrium[2]	Cumulative storage depletion after 300 years of pumping (acre-feet)	Principal source of water after 300 years of pumping
A	920	Near ET	Strategically distributed.	28	Complete	Attained	26,000	Reduction in evapotranspiration.
B	1,840	...do.	...do.	[3]90	...do.	Approaching	99,000	Induced flow from Carson River.
C	1,840 or 920	...do.	...do.	[3,4]53	...do.	Attained	46,000	Reduction in evapotranspiration.
D	920	...do.	Concentrated	42	...do.	...do.	31,000	Do.
E	1,840	...do.	...do.	[3]114	...do.	Approaching	110,000	Induced flow from Carson River.
F	920	North end of Stagecoach Valley.	...do.	[3]82	...do.	...do.	40,000	Reduction in evapotranspiration.
G	1,840	...do.	...do.	[3]186	Almost complete	...do.	122,000	Induced flow from Carson River.
H	920	Near Carson River	...do.	43	Complete	Attained	7,000	Do.
I	970	Same as in 1981	Same as in 1981	35	...do.	Approaching	33,000	Reduction in evapotranspiration.

[1]Water-level recovery is considered complete when maximum residual drawdown is less than 0.5 foot.
[2]A new equilibrium is considered to be attained if less than 3 percent of annual pumpage is derived from depletion of ground-water storage.
[3]Simulated declines would be larger if simulation accounted for decrease in transmissivity due to dewatering aquifer materials.
[4]Maximum decline of 53 feet occurred after 50 years of pumping (just before rate was reduced to 920 acre-feet per year). After 300 years of pumping, decline was only 30 feet.

is not simply the ability to obtain water by pumping; rather, the limiting factor is the extent of allowable adverse effects on other water users. Virtually all water in the Carson River is appropriated, and increasing the subsurface flow from the river into the basin-fill aquifer of Stagecoach Valley would affect downstream users along the Carson River. Similar effects on ground-water users could conceivably occur in Churchill Valley.

Determining the extent to which effects on areas outside Stagecoach Valley can be tolerated is a legal matter and is beyond the scope of this report. Moreover, limiting pumping to the natural (predevelopment) inflow rate of 920 acre-ft/yr does not guarantee that conflicts with surface-water users along the Carson River will be avoided. The natural ground-water budget includes some ground-water outflow to the downstream reach of the river, which, if captured by pumping, could adversely affect the downstream river flows. The quantity would be small in relation to the average flow of the river and probably would not cause a detectable change except during low-flow periods. The location of pumping wells also significantly affects downstream users. For example, from scenario H (fig. 33) the effects on stream flow near the south end of Stagecoach Valley are much greater and occur more rapidly than the effects from either scenarios A or D (figs. 26 and 29, respectively) in the north-central part of the valley, even though the net rate of pumping is the same.

One factor that is common to all the scenarios is that the leaky boundary conditions of the basin-fill aquifer strongly influence the results of the simulation. In order to apply the general findings of this analysis to other small basins within the Great Basin, considerable effort should be expended on defining boundary conditions, which might not be required for the study of a large area.

In summary, the simulated effects of pumping spread quickly throughout the entire basin-fill aquifer in all scenarios, and the capture of discharge probably proceeded more rapidly than it would have in a larger basin. When the aquifer was stressed by pumping, additional inflow induced from the Carson River and to a small extent from Churchill Valley tended to reduce water-level declines. Consequently, if water-level changes alone are used to evaluate the sensitivity of the model to variations in pumping rates, then sensitivity is underestimated; changes in subsurface inflow and outflow also must be included in any evaluation of how the basin-fill aquifer responds to heavy pumping. Moreover, this additional inflow ultimately causes the simulated aquifer to attain a new equilibrium in response to a relatively wide range of pumping rates. The predevelopment flux through the basin-fill aquifer is not necessarily the limiting criterion in determining the maximum sustained pumping rate; the limiting factor probably is the degree to which adverse effects on Carson River flows can be tolerated.

SUMMARY AND CONCLUSIONS

Stagecoach Valley is a small, topographically closed basin in western Nevada with a total area of about 70 mi^2. About 33 mi^2 of the area is underlain by basin-fill deposits; the remainder is underlain by consolidated rock. Surface drainage is from mountains that border most of the area toward Misfits Flat in the south-central part of the valley. Most locally derived runoff and recharge is generated in the Flowery Range, which forms the northwest border and contains the highest altitudes in the area. Churchill Butte and some relatively small unnamed mountains form the east and west boundaries of the area.

The south boundary is more complex. A topographic divide that traverses a series of small mountains and alluvial and lake deposits forms the topographic boundary of the basin. However, local subsurface hydraulic continuity exists between andesitic and basaltic rocks and alluvium beneath the topographic divide. Flood-plain deposits of the Carson River lie immediately south of this topographic divide; thus, the true hydrologic boundary of the basin-fill aquifer of Stagecoach Valley is the Carson River.

The structural depression that underlies Stagecoach Valley is partly filled by deposits of sand, gravel, silt, and clay that form the basin-fill aquifer. The fill is at least 500 ft thick throughout most of the area and attains a maximum thickness of about 3,000 ft. The estimated average hydraulic conductivity of the basin-fill deposits ranges from 1.0×10^{-6} ft/s for the finer grained deposits to as much as 1.7×10^{-3} ft/s for the coarser grained deposits.

The amount of ground water in storage under natural conditions, presumed to correspond to estimated pre-1971 (predevelopment) conditions, is about 1 million acre-ft; however, most of this stored water probably cannot be economically pumped because of excessive lifts and probable effects on the Carson River. Also estimated from predevelopment conditions, 89,000 acre-ft of water is stored in the upper 200 ft of saturated basin fill in the area that was developed as of 1982. This amount of water might be economically pumped. However, constraints related to potential effects on Carson River flow may limit storage depletion to significantly less than this amount.

A numerical model described by McDonald and Harbaugh (1988) was used to simulate the basin-fill aquifer; the aquifer was discretized into three layers. The top model layer (layer 1) approximates the water-table portion of the basin-fill deposits and thus represents the aquifer under unconfined conditions. The processes of recharge from precipitation and irrigation-return flow, evapotranspiration of ground water, interaction with flow in the Carson River, and depletion of storage by ground-water withdrawals all occur in layer 1, which is assumed to be about 50 ft thick. The middle layer (layer 2) represents that part of the basin fill most affected by pumping and is considered to represent the saturated zone at depths of 50 to about 300 ft below the water table. In this study, all pumping is assumed to be from layer 2. The lowest layer (layer 3) consists of saturated basin fill below layer 2, and its thickness varies laterally.

The simulation of predevelopment conditions indicated that the flux through the basin-fill aquifer was about 920 acre-ft/yr. This total inflow to the system included about 550 acre-ft/yr recharge from precipitation, about 280 acre-ft/yr subsurface inflow from flood-plain deposits beneath the upstream reach of the Carson River, and about 90 acre-ft/yr subsurface inflow from flood-plain deposits in the downstream reach of the Carson River. Total outflow from the system included about 640 acre-ft/yr by evapotranspiration, about 170 acre-ft/yr by subsurface flow to Churchill Valley, and about 120 acre-ft/yr by subsurface flow to deposits beneath the downstream reach of the Carson River.

Ground-water development began in 1971, when about 310 acre-ft of water was pumped. Pumpage increased steadily until 1977, when about 1,600 acre-ft was pumped and then decreased slightly in the following years. In 1981 about 1,100 acre-ft of water was pumped. This included about 930 acre-ft for irrigation, about 30 acre-ft for domestic and stock use, and about 130 acre-ft for public supply.

During the 11 pumping seasons from spring 1971 through spring 1982, slightly more than 11,000 acre-ft of ground water was pumped in Stagecoach Valley. Of this total, slightly less than 1,000 acre-ft was recirculated to the aquifer by irrigation-return flow, resulting in a net pumpage of about 10,000 acre-ft. About 3,000 acre-ft of the net pumpage resulted from reductions in evapotranspiration and from changes in subsurface inflow and outflow; the remaining 7,000 acre-ft was removed from aquifer storage. In response to this pumpage, water levels declined by 1 ft or less near the Carson River to more than 15 ft near pumping wells.

Model simulation of the ground-water system for the 11 pumping seasons (spring 1971 to spring 1982) indicates that the rate of evapotranspiration of ground water had decreased by about 460 acre-ft/yr by spring 1982. Subsurface inflow from the upstream reach of the Carson River had increased by about 20 acre-ft/yr, inflow from the downstream reach of the Carson River had increased by about 10 acre-ft/yr, and subsurface outflow to the downstream reach of the Carson River had decreased by about 7 acre-ft/yr. Subsurface outflow to Churchill Valley had decreased by about 8 acre-ft/yr. A total of about 450 acre-ft/yr was estimated to be captured by the pumping.

The potential response to pumping stresses over a 300-yr period was evaluated using nine hypothetical development scenarios. Eight of the scenarios involved pumping at the rate of 920 or 1,840 acre-ft/yr (equal to or twice the predevelopment flux) or pumping at a mixed rate. Assumed pumping strategies for these eight scenarios included distributing pumping wells over a relatively large area, concentrating pumping in one area, strategically locating pumping wells with respect to areas of natural discharge, concentrating pumping at the north end of Stagecoach Valley, and concentrating pumping at the south end of the valley near the Carson River. The ninth scenario used a well distribution and average pumping rate similar to those in use in 1981. A 300-yr pumping period and a subsequent 300-yr recovery period were simulated for each scenario. The following general statements can be made on the basis of the simulation results:

1. The concept of sustained yield appears to be viable for hydrologic conditions like those in Stagecoach Valley. When simulated pumping was held to the predevelopment (natural) inflow or outflow rate of 920 acre-ft/yr, the aquifer essentially attained a new equilibrium after 300 yr of pumping. Moreover, water-level declines in and adjacent to the areas of pumping were generally less than 40 ft, and the greatest declines occurred during the first 50 yr of the simulation. When pumping rates were mixed (50 yr at 1,840 acre-ft/yr and 250 yr at 920 acre-ft/yr), the system also attained a new equilibrium after 300 yr. In the scenarios where the pumping rate was held at 1,840 acre-ft/yr, the aquifer also approached a new equilibrium after 300 yr of pumping even though maximum drawdowns ranged from

90 to 186 ft, depending on the distribution of the pumping wells. In these scenarios additional subsurface inflow was induced from the Carson River and to a small extent from adjacent Churchill Valley; such induced inflow makes achieving a new equilibrium at least theoretically possible. Because the simulated aquifer ultimately attains a new equilibrium in response to a relatively wide range of pumping rates, the predevelopment flux is not necessarily the limiting criterion in determining the sustained pumping rate. The limiting factor probably is the degree to which reduction in Carson River flows can be tolerated.

2. In the model, the basin-fill aquifer was sensitive to pumping rates in that the maximum simulated drawdowns under a pumping rate of 1,840 acre-ft/yr were as much as two to six times as great as those under a pumping rate limited to the predevelopment inflow rate of 920 acre-ft/yr. Substantial additional water inflow that was induced from the Carson River tended to attenuate the rates of water-level decline. If change in water level is the only criterion used to evaluate the sensitivity of the model to variations in pumping rates, then the sensitivity is underestimated. Changes in subsurface inflow and outflow also must be included in any evaluation.

3. In the model simulation, the aquifer also was sensitive to the location of the pumping wells. As previously mentioned, both water-level changes and variations in inflow and outflow must be evaluated to judge the full sensitivity of the aquifer response. Generally, minimal water-level declines occurred in the scenarios where pumping wells were located near areas of natural discharge or near potential sources of induced recharge. Conversely, when pumping wells were located away from these areas, water-level declines were significantly greater.

4. In the model simulation, the effects of pumping spread quickly throughout the entire aquifer, and the capture of discharge probably proceeds more rapidly in a basin the size of Stagecoach Valley than it would in a larger basin. The characteristics of the boundaries of the basin-fill aquifer, as modeled, greatly affected the simulated aquifer response. During pumping, additional inflow induced from the Carson River, and to a small extent from Churchill Valley, tended to reduce water-level declines. If a basin of similar size had no-flow conditions at all boundaries, the boundary effects probably would result in water-level declines much greater than those simulated for Stagecoach Valley. The general conclusion is that in small alluvial basins similar in hydrologic setting to Stagecoach Valley, the boundary conditions of the aquifers probably are much more significant in determining the response to pumping stresses than they are in larger but hydrologically similar basins.

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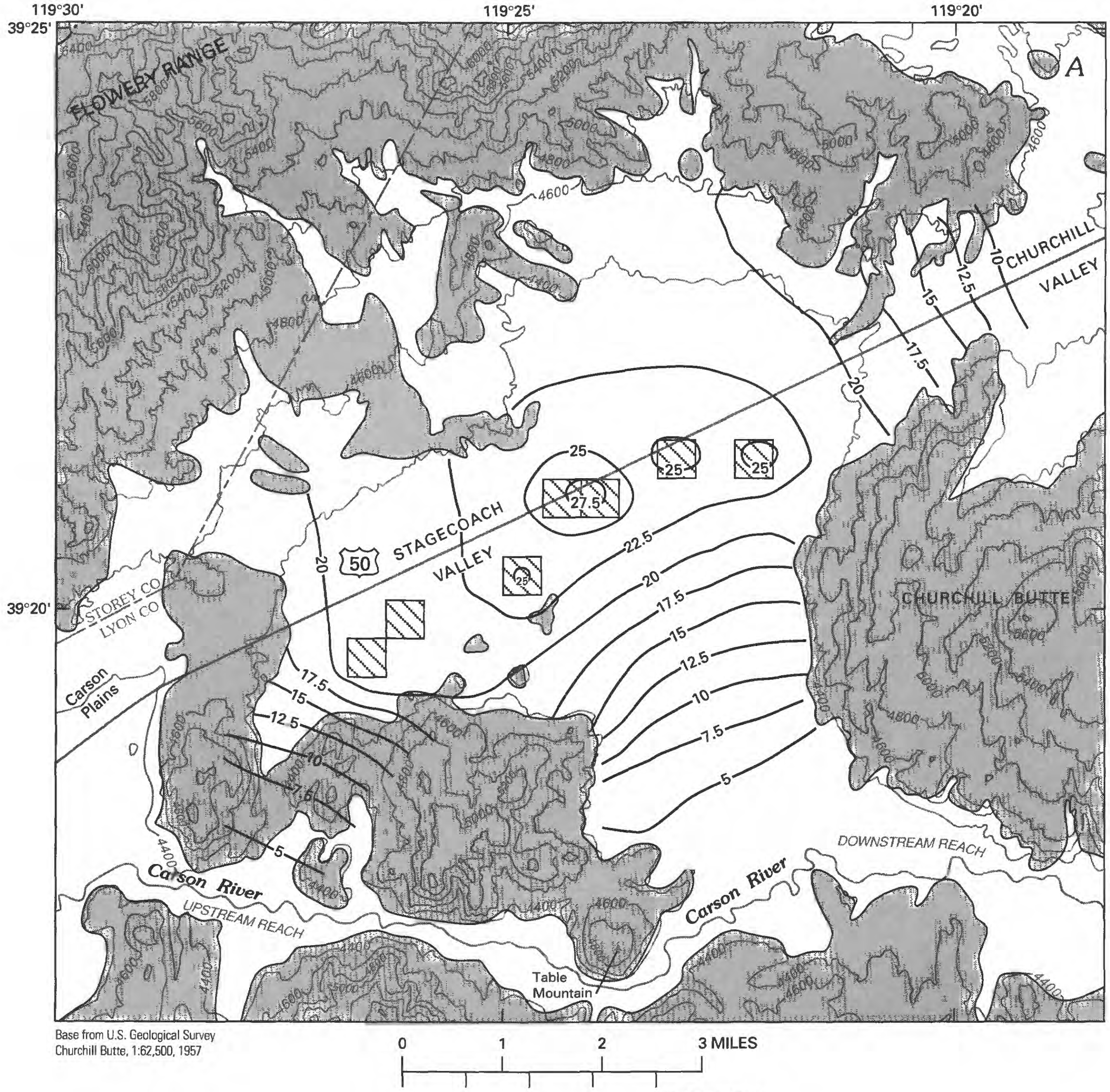
FIGURES 26–34 AND TABLES 10–18



EXPLANATION

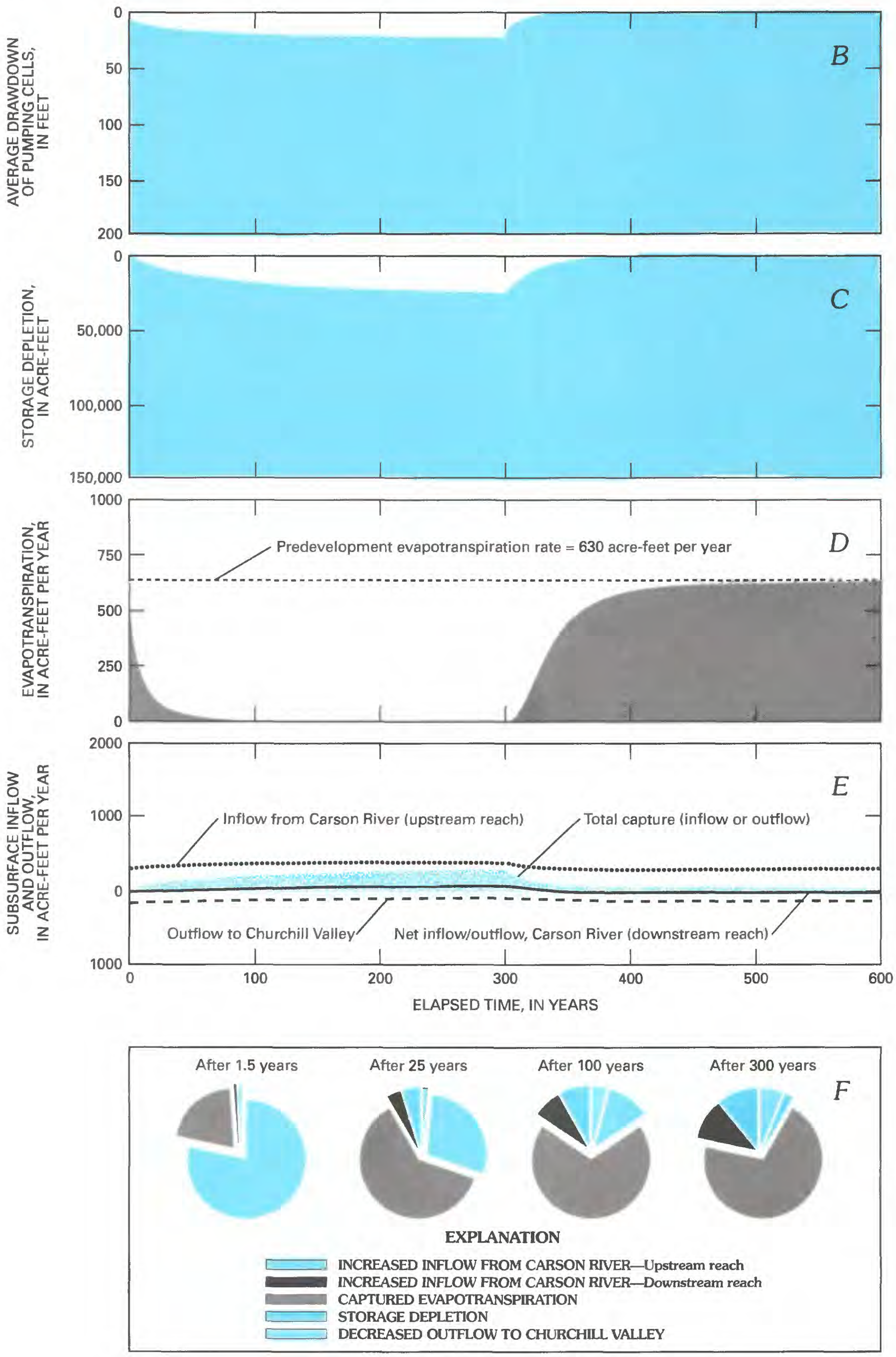
BASIN FILL

CONSOLIDATED ROCK

PUMPING CELL

—20— LINE OF EQUAL SIMULATED DRAWDOWN IN LAYER 2 AFTER 300 YEARS OF PUMPING—Interval, 2.5 feet

FIGURE 26.—Scenario A: Simulated response with pumping cells strategically located to capture evapotranspiration. Pumping rate is equal to estimated average predevelopment inflow and outflow (steady-state pumping rate of 920 acre-feet per year). *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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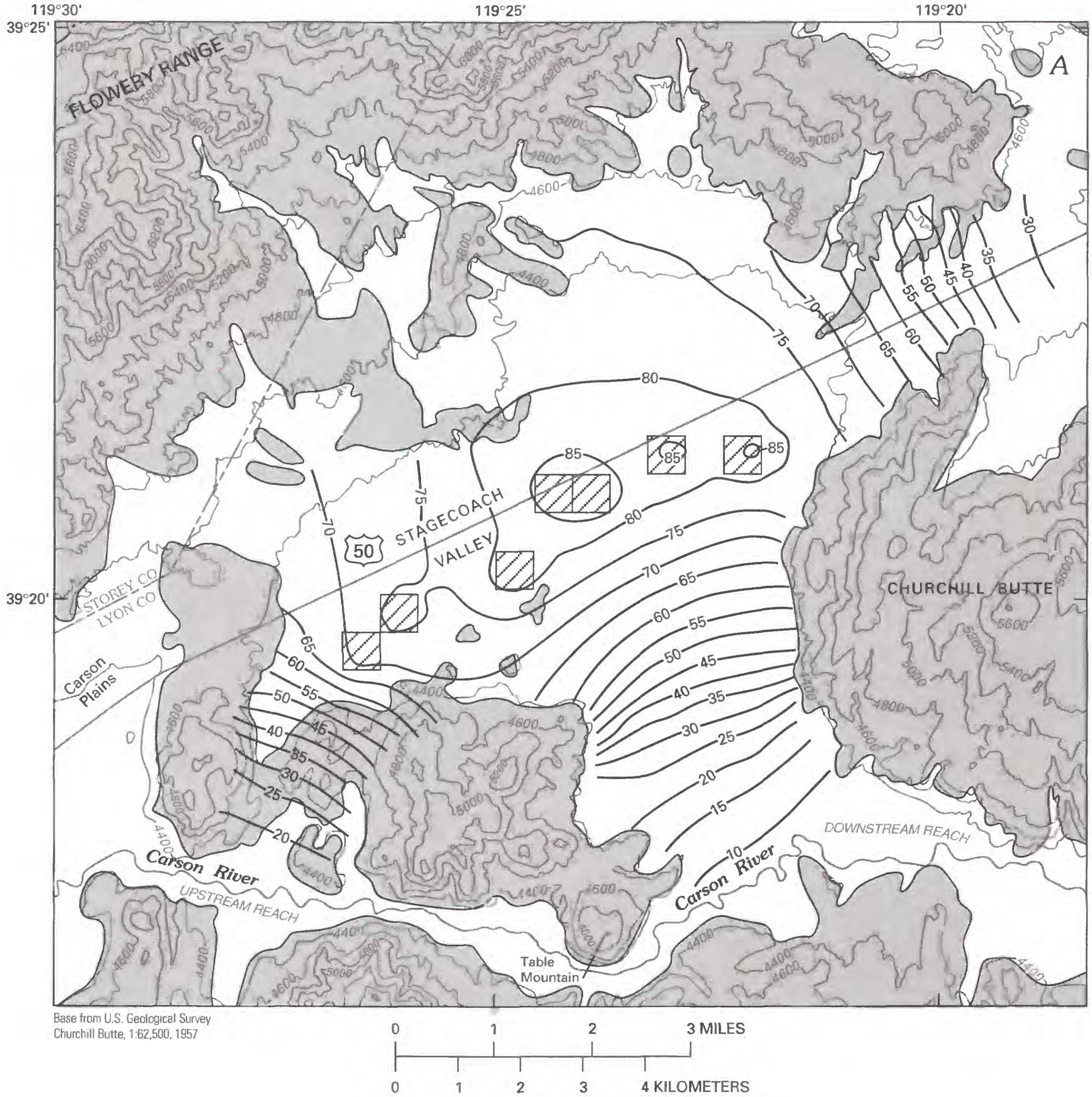


EXPLANATION

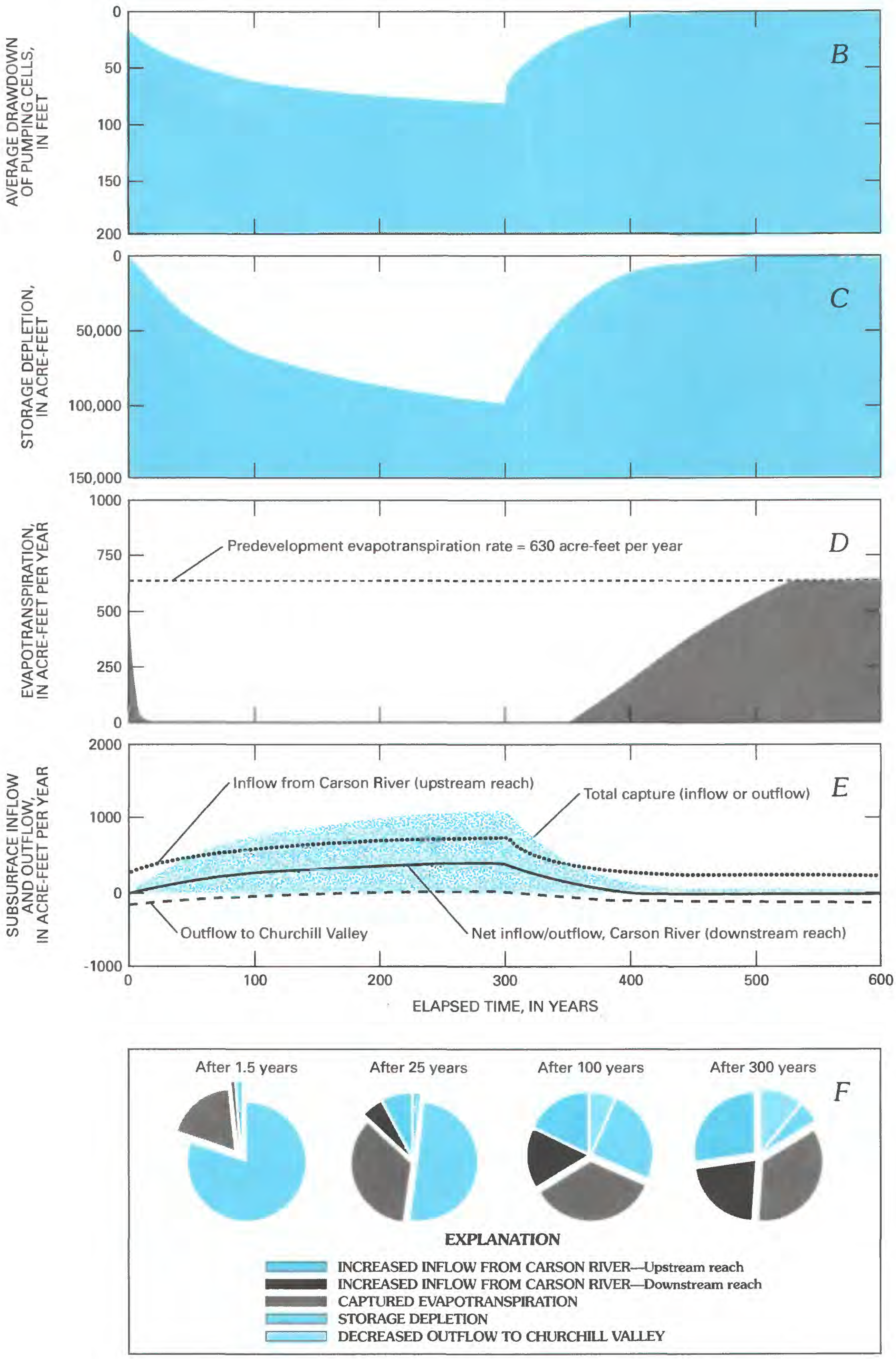
BASIN FILL

CONSOLIDATED ROCK

PUMPING CELL

—20— LINE OF EQUAL SIMULATED DRAWDOWN IN LAYER 2 AFTER 300 YEARS OF PUMPING—Interval, 5 feet

FIGURE 27.—Scenario B: Simulated response with pumping cells strategically located to capture evapotranspiration. Pumping rate is 1,840 acre-feet per year (equal to twice estimated average predevelopment inflow or outflow). *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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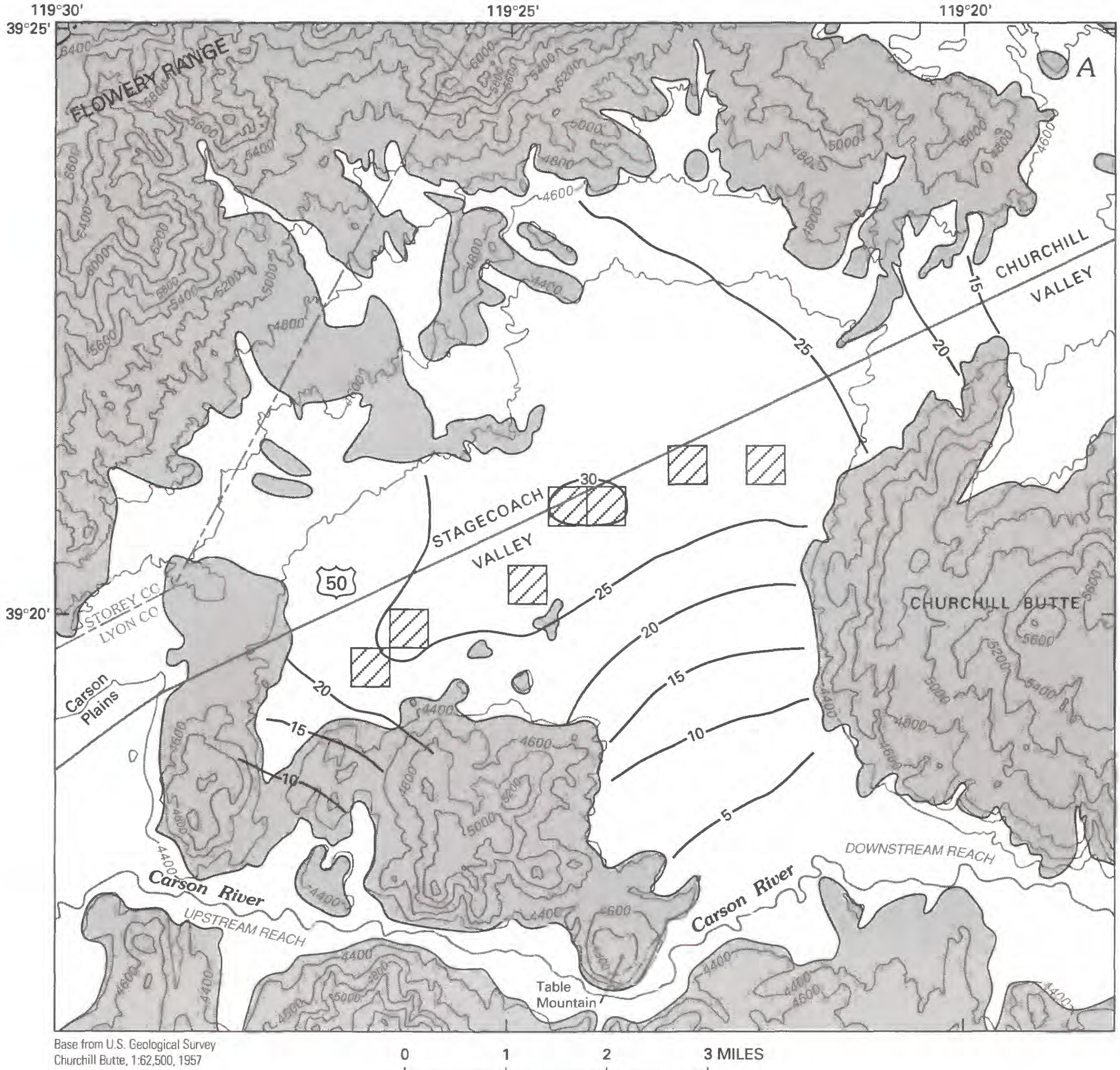
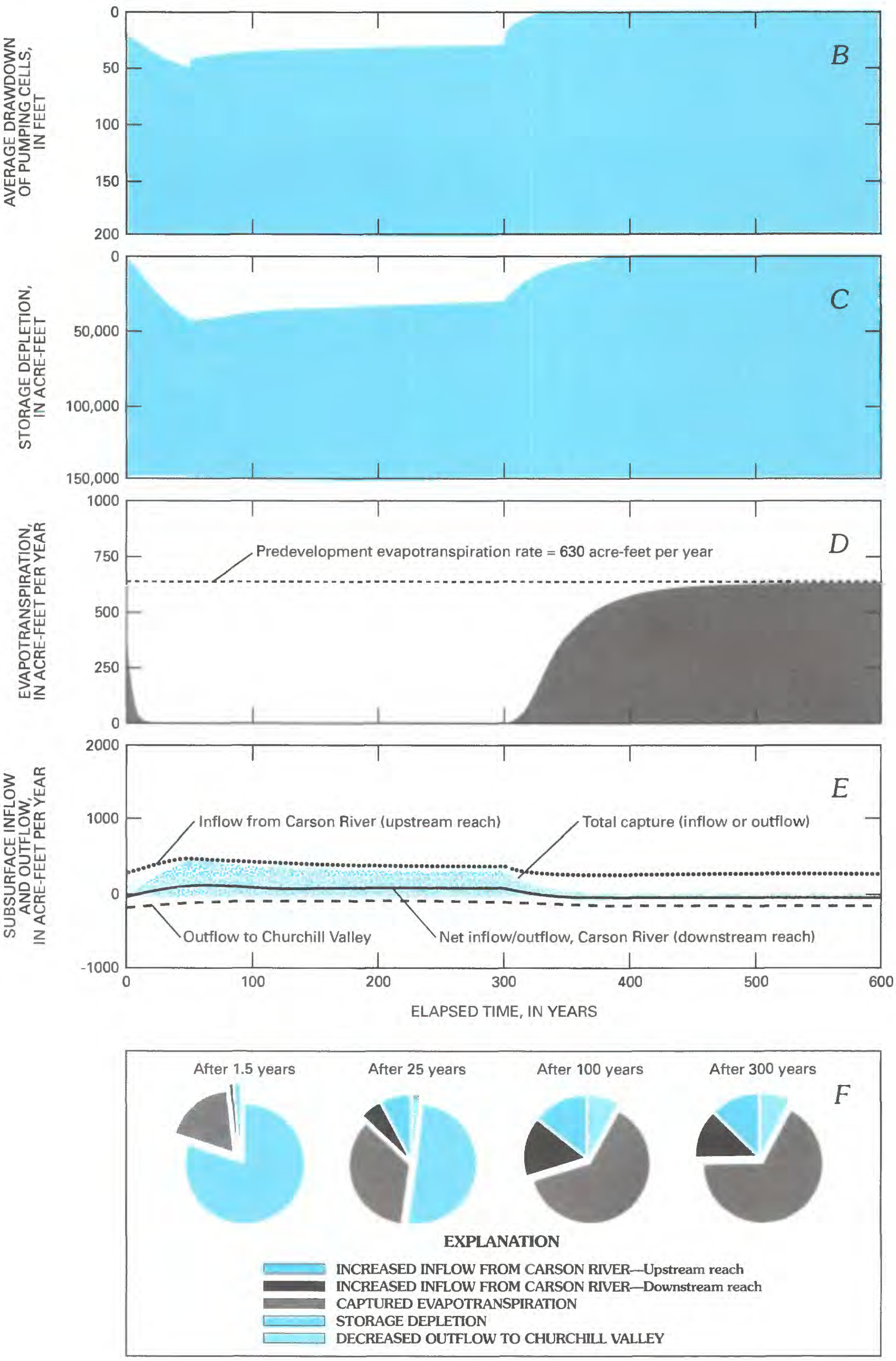


EXPLANATION

- BASIN FILL
- CONSOLIDATED ROCK
- PUMPING CELL
- —20— LINE OF EQUAL SIMULATED DRAWDOWN IN LAYER 2 AFTER 300 YEARS OF PUMPING—Interval, 5 feet

FIGURE 28.—Scenario C: Simulated response with pumping cells strategically located to capture evapotranspiration and pumping rate varied over time (50 years at 1,840 acre-feet per year, twice estimated average predevelopment inflow or outflow, then 250 years at 920 acre-feet per year, equal to estimated average annual predevelopment inflow or outflow). *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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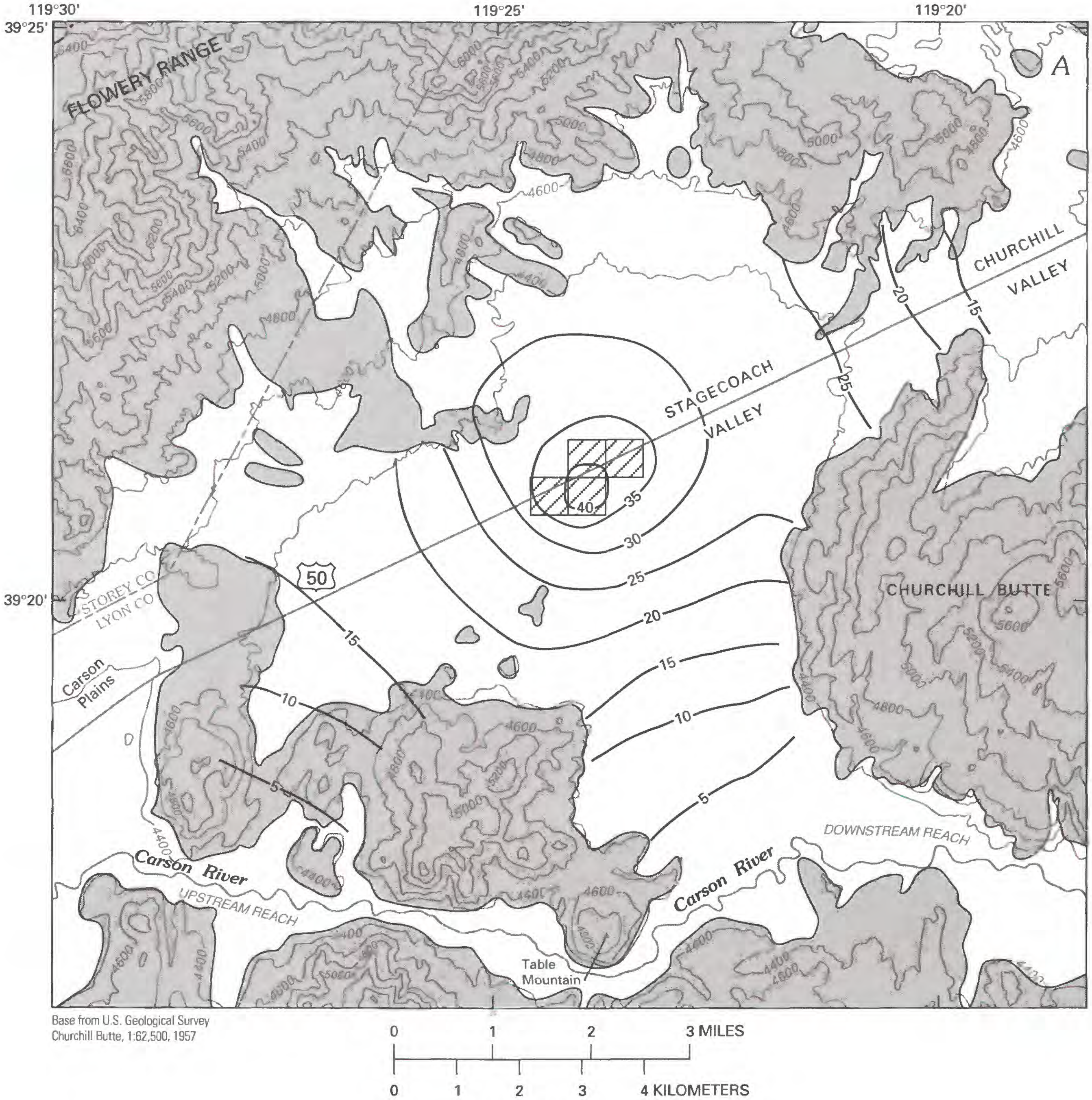


EXPLANATION

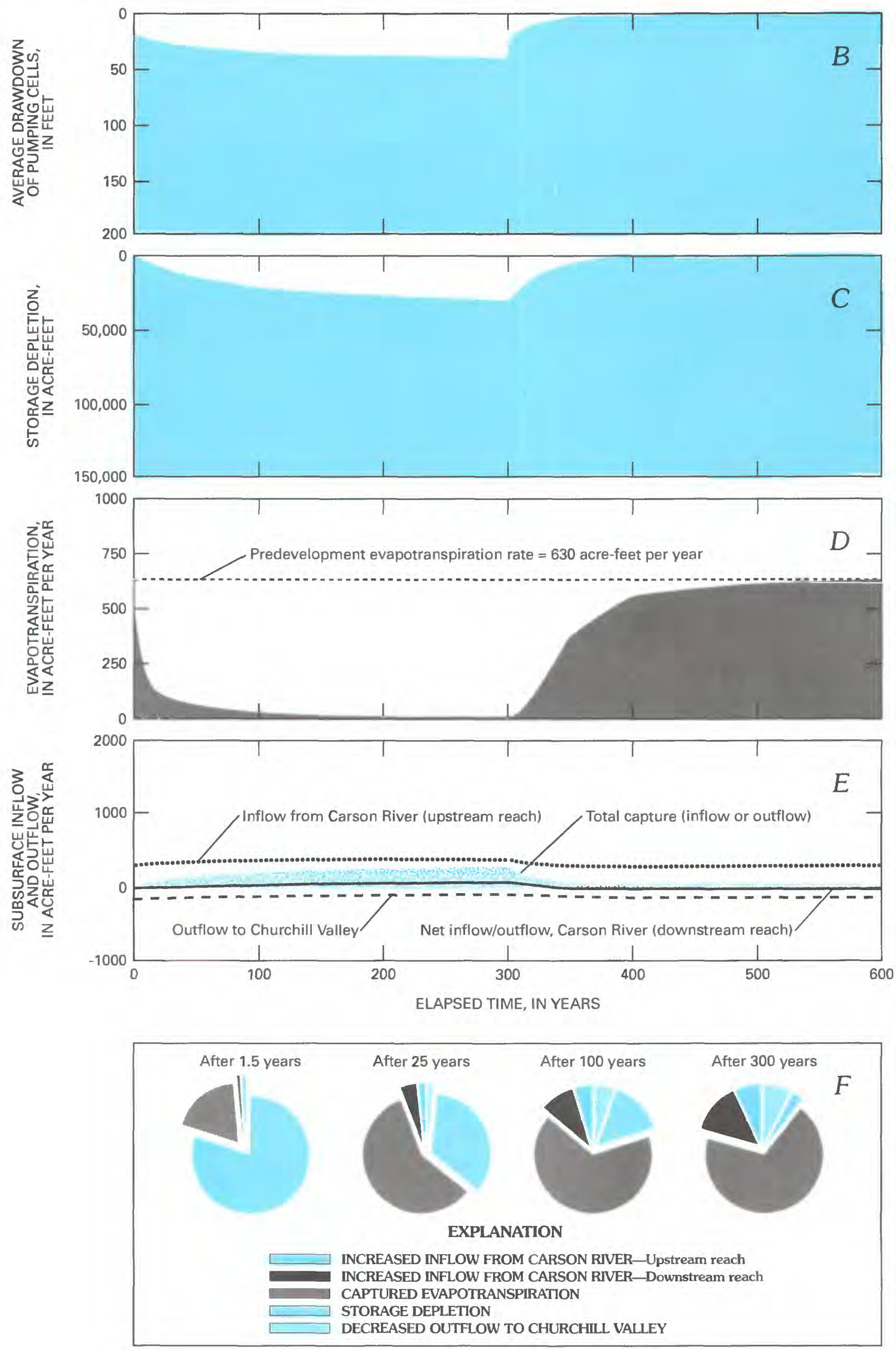
BASIN FILL

CONSOLIDATED ROCK

PUMPING CELL

—20— LINE OF EQUAL SIMULATED DRAWDOWN IN LAYER 2 AFTER 300 YEARS OF PUMPING—Interval, 5 feet

FIGURE 29.—Scenario D: Simulated response with pumping cells concentrated in area of evapotranspiration. Pumping rate is 920 acre-feet per year, equal to estimated average predevelopment inflow or outflow. *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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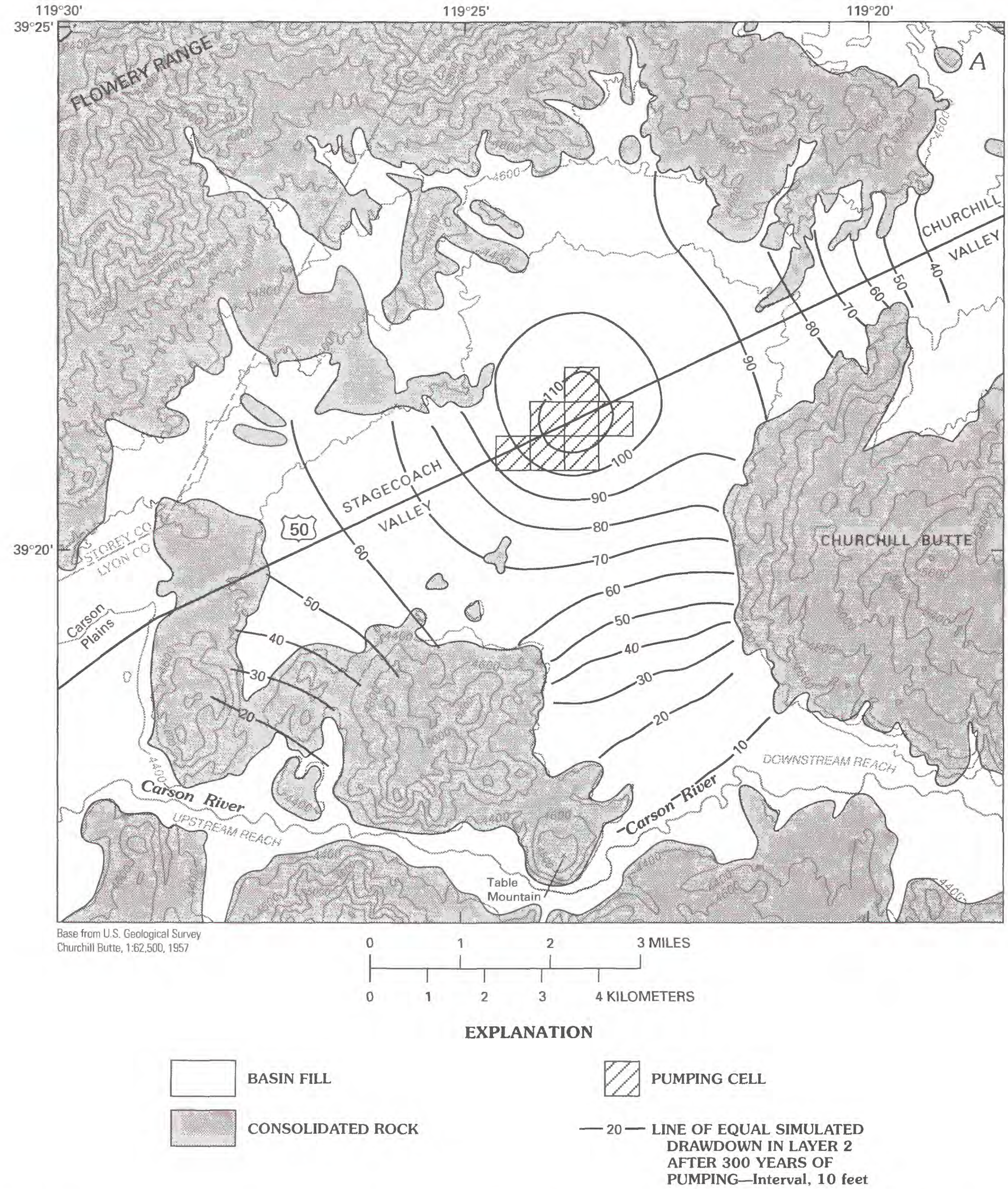
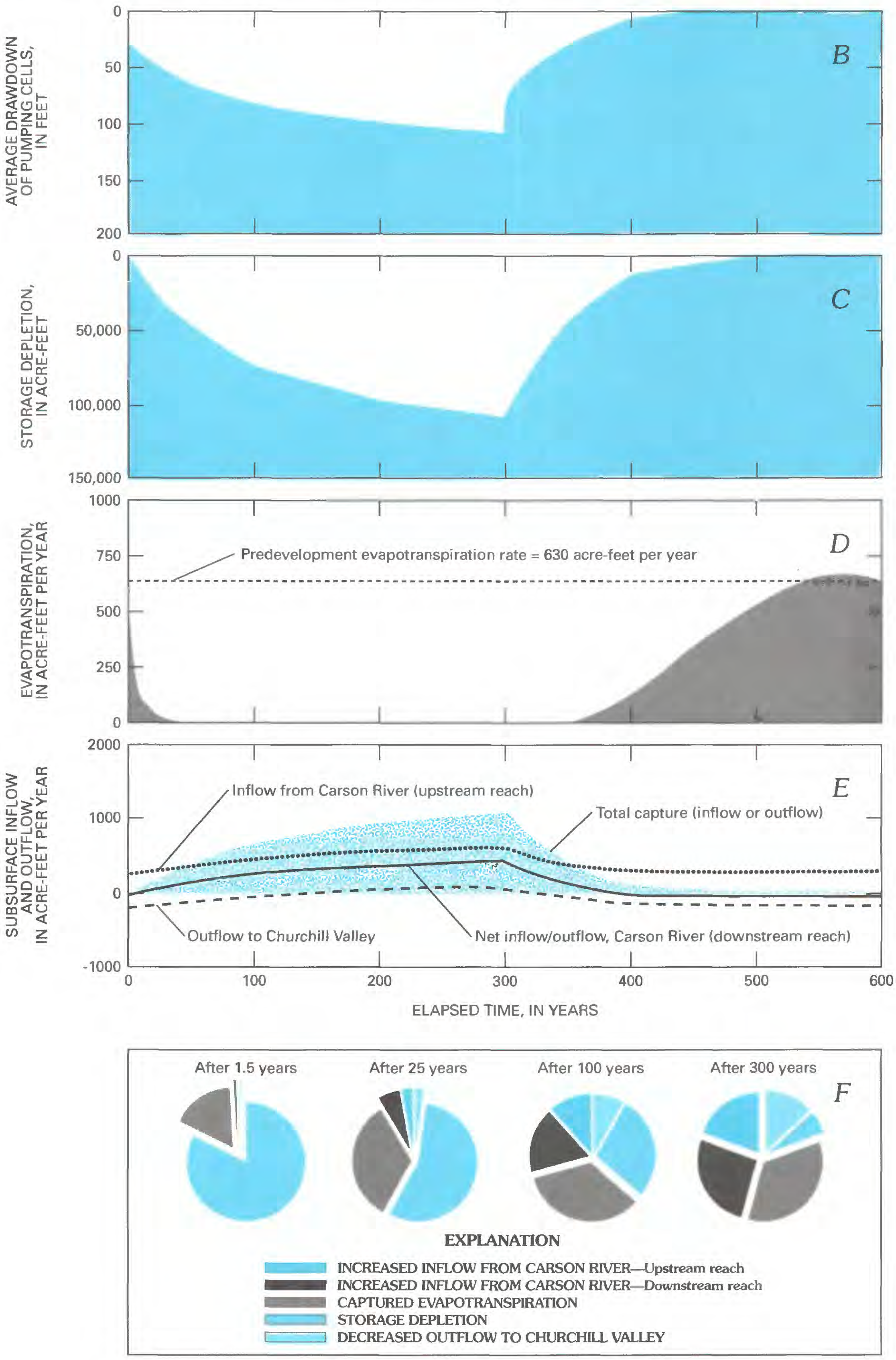


FIGURE 30.—Scenario E: Simulated response with pumping cells concentrated in area of evapotranspiration. Pumping rate is 1,840 acre-feet per year, equal to twice estimated average predevelopment inflow or outflow. *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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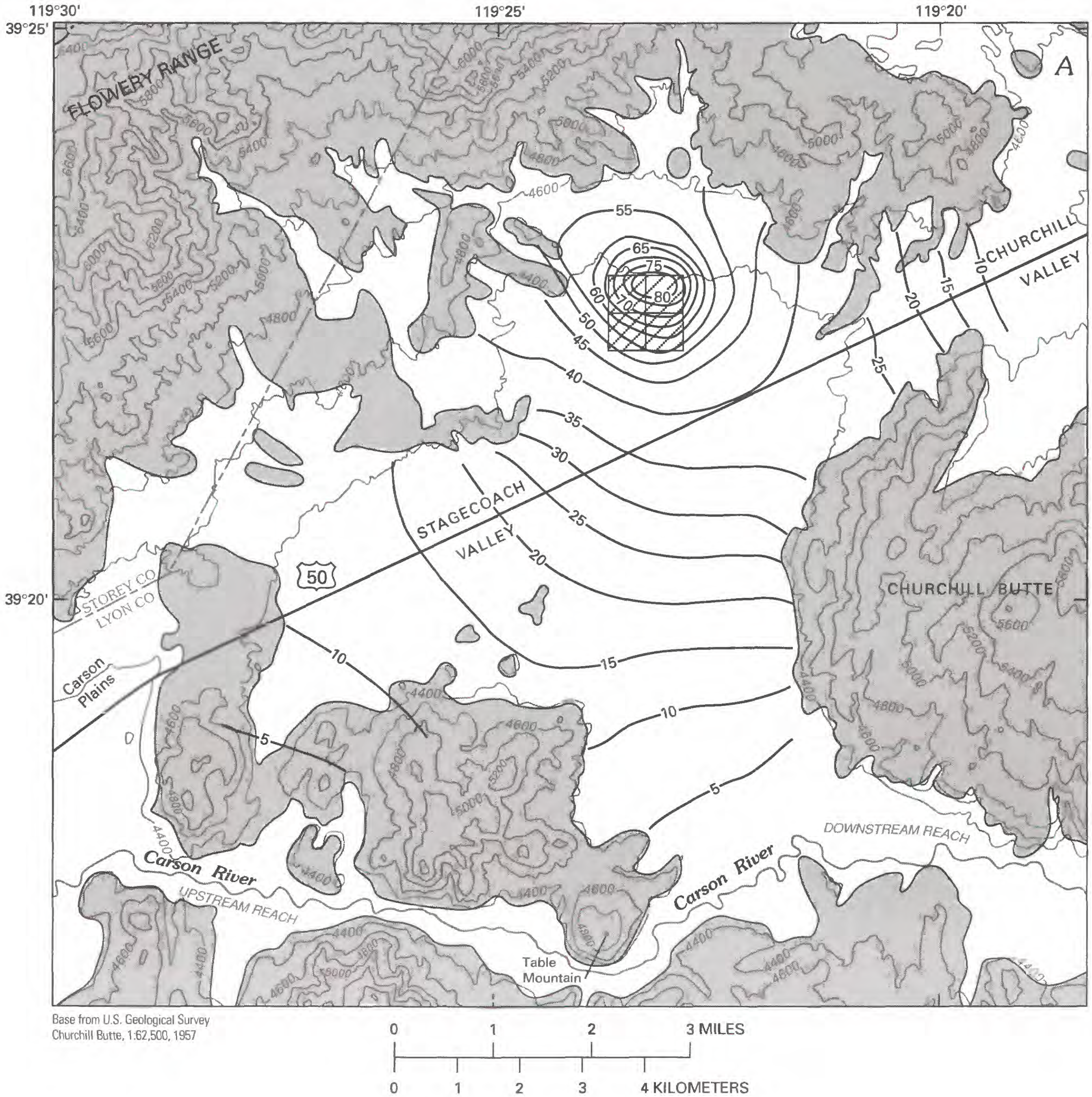


EXPLANATION

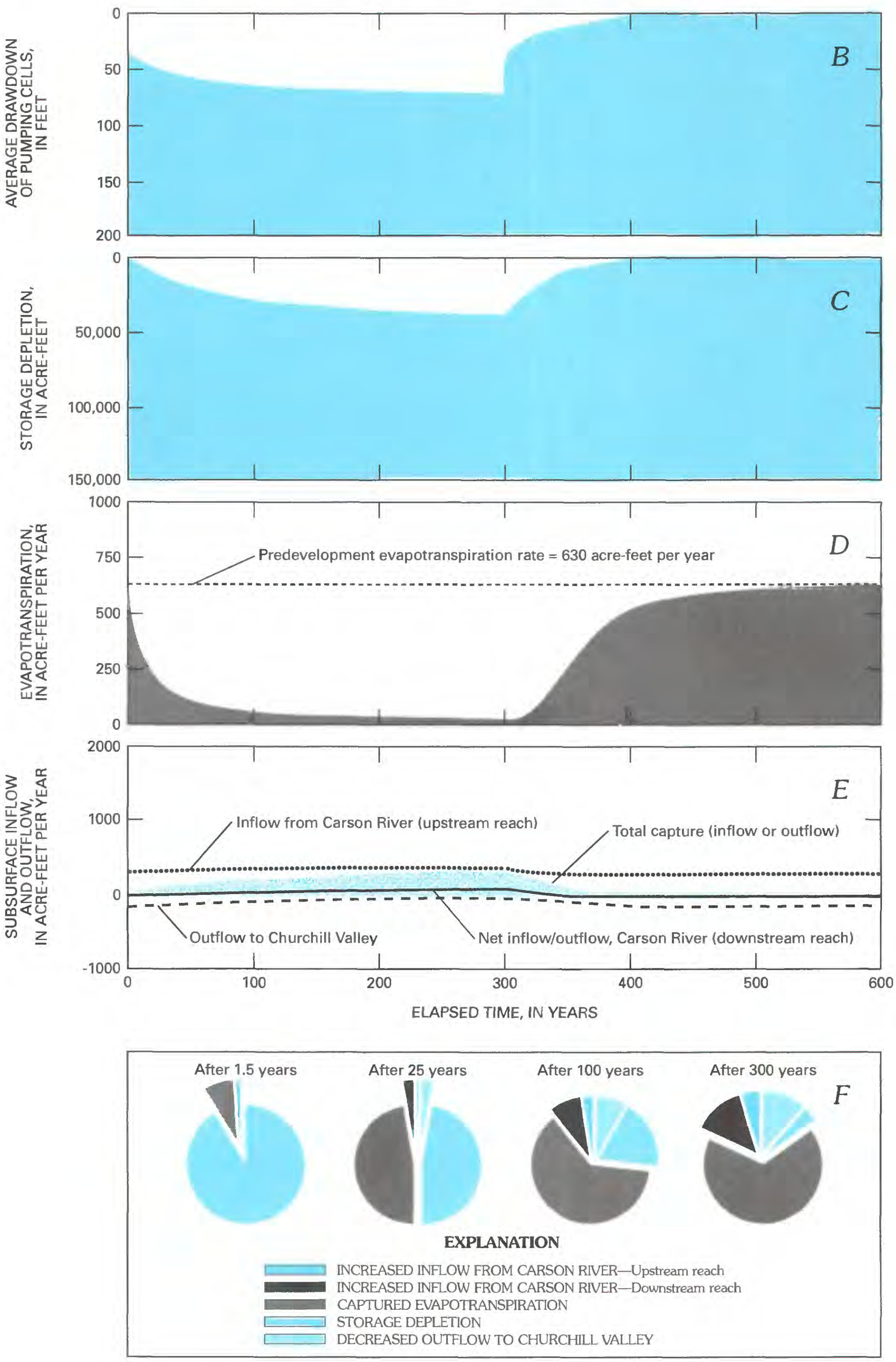
BASIN FILL

CONSOLIDATED ROCK

PUMPING CELL

—20— LINE OF EQUAL SIMULATED DRAWDOWN IN LAYER 2 AFTER 300 YEARS OF PUMPING—Interval, 5 feet

FIGURE 31.—Scenario F: Simulated response with pumping cells concentrated at north end of valley. Pumping rate is 920 acre-feet per year, equal to estimated average predevelopment inflow or outflow. *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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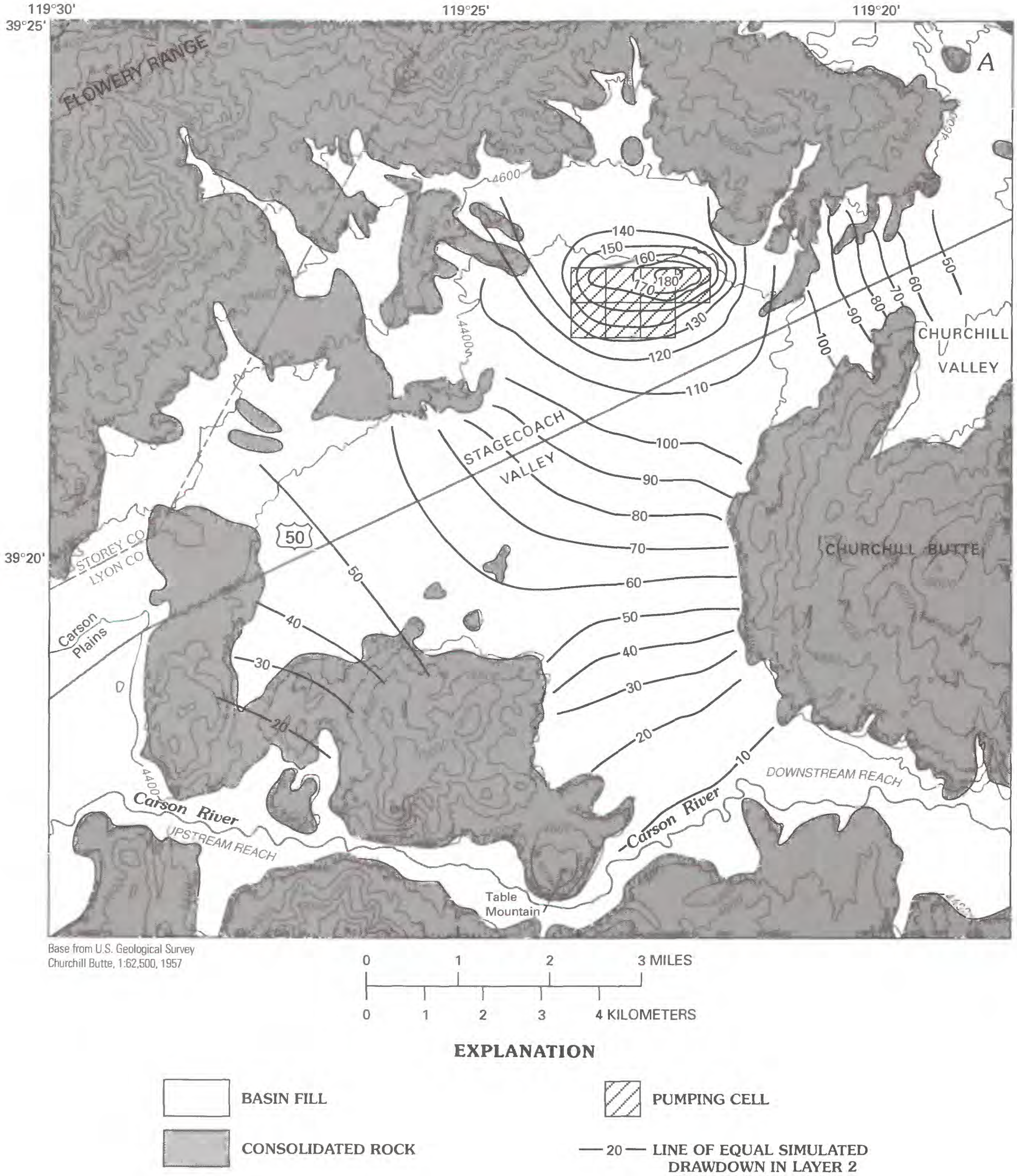
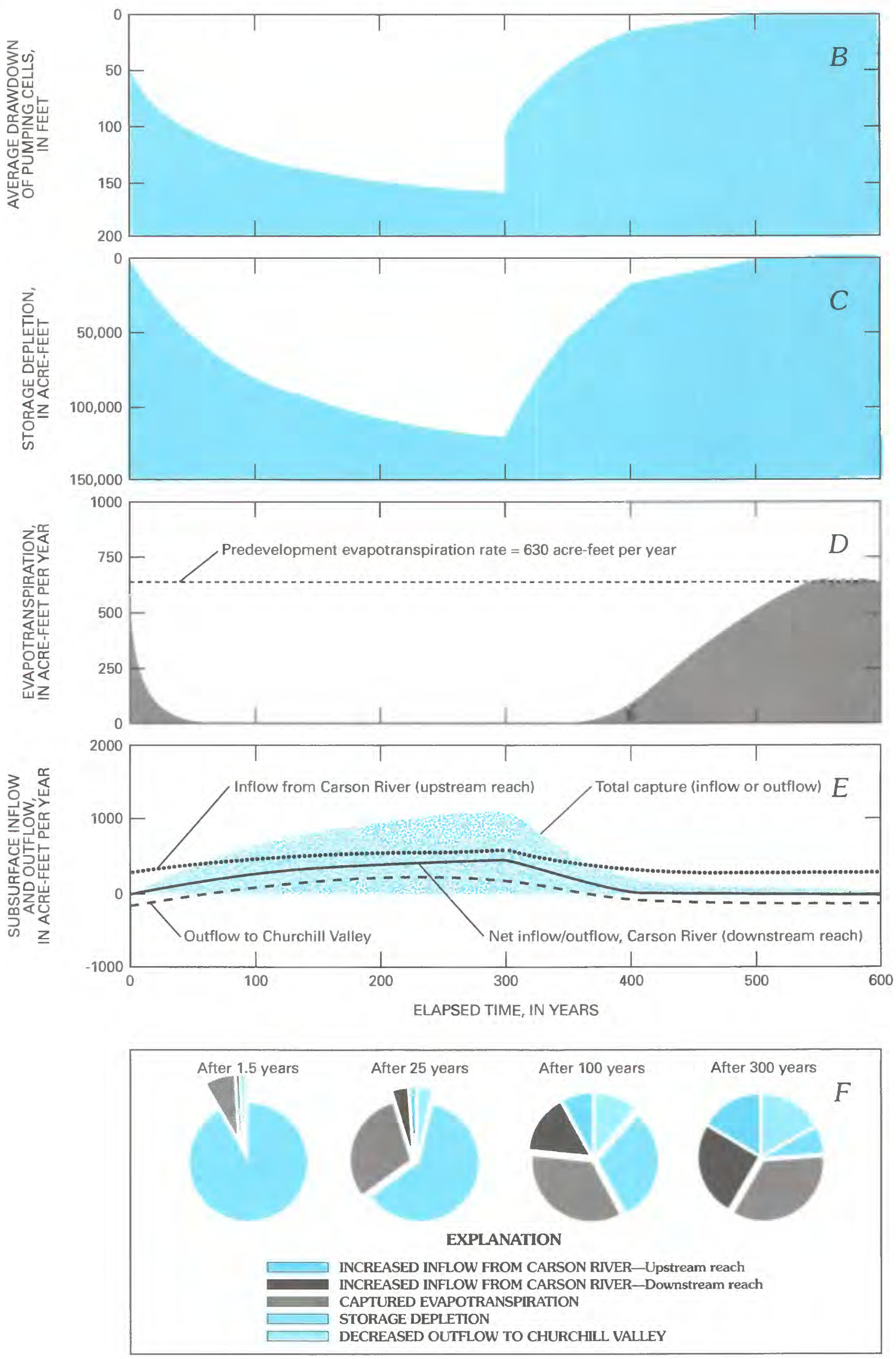


FIGURE 32.—Scenario G: Simulated response with pumping cells concentrated at north end of valley. Pumping rate is 1,840 acre-feet per year, equal to twice estimated average predevelopment inflow or outflow. *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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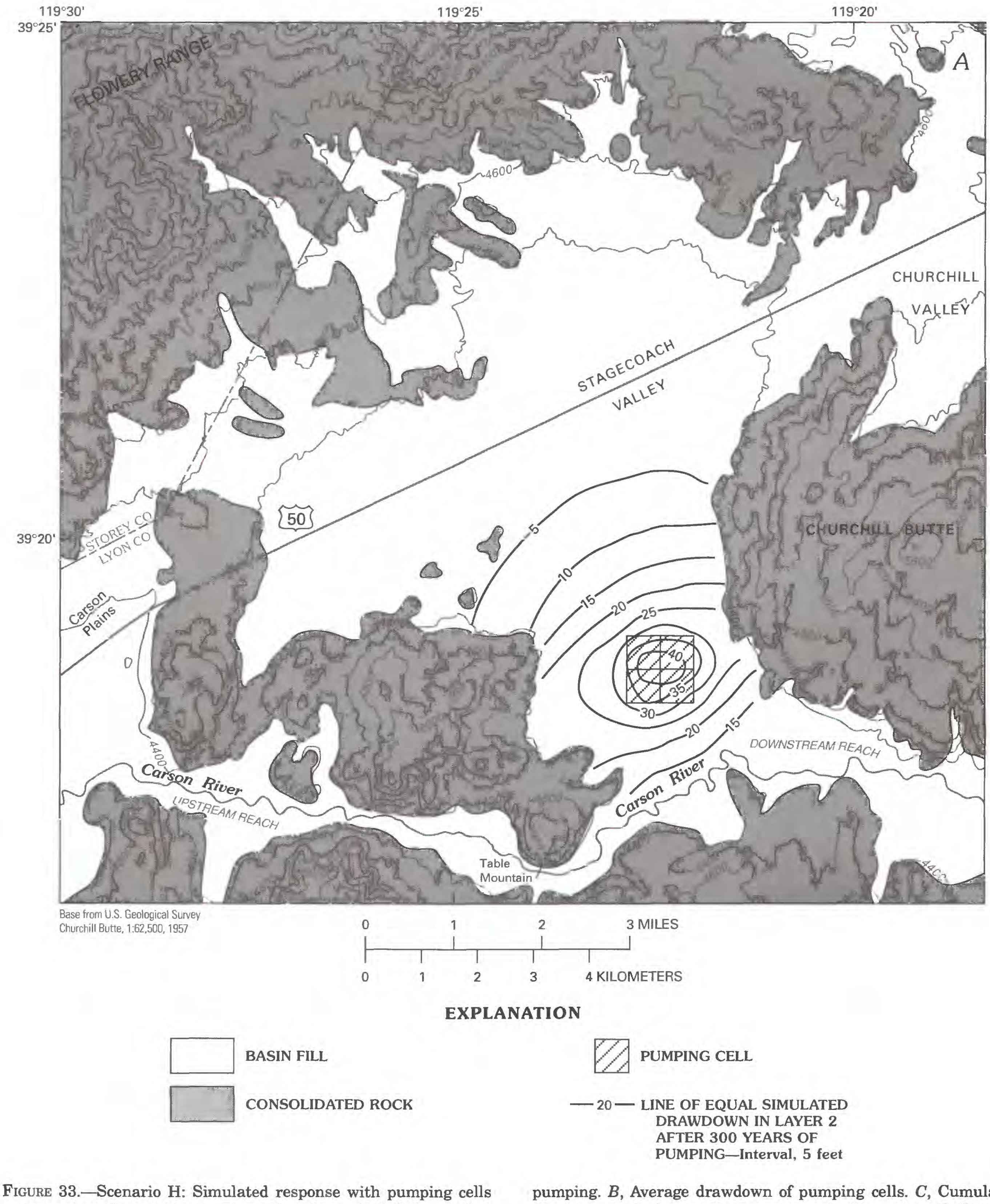
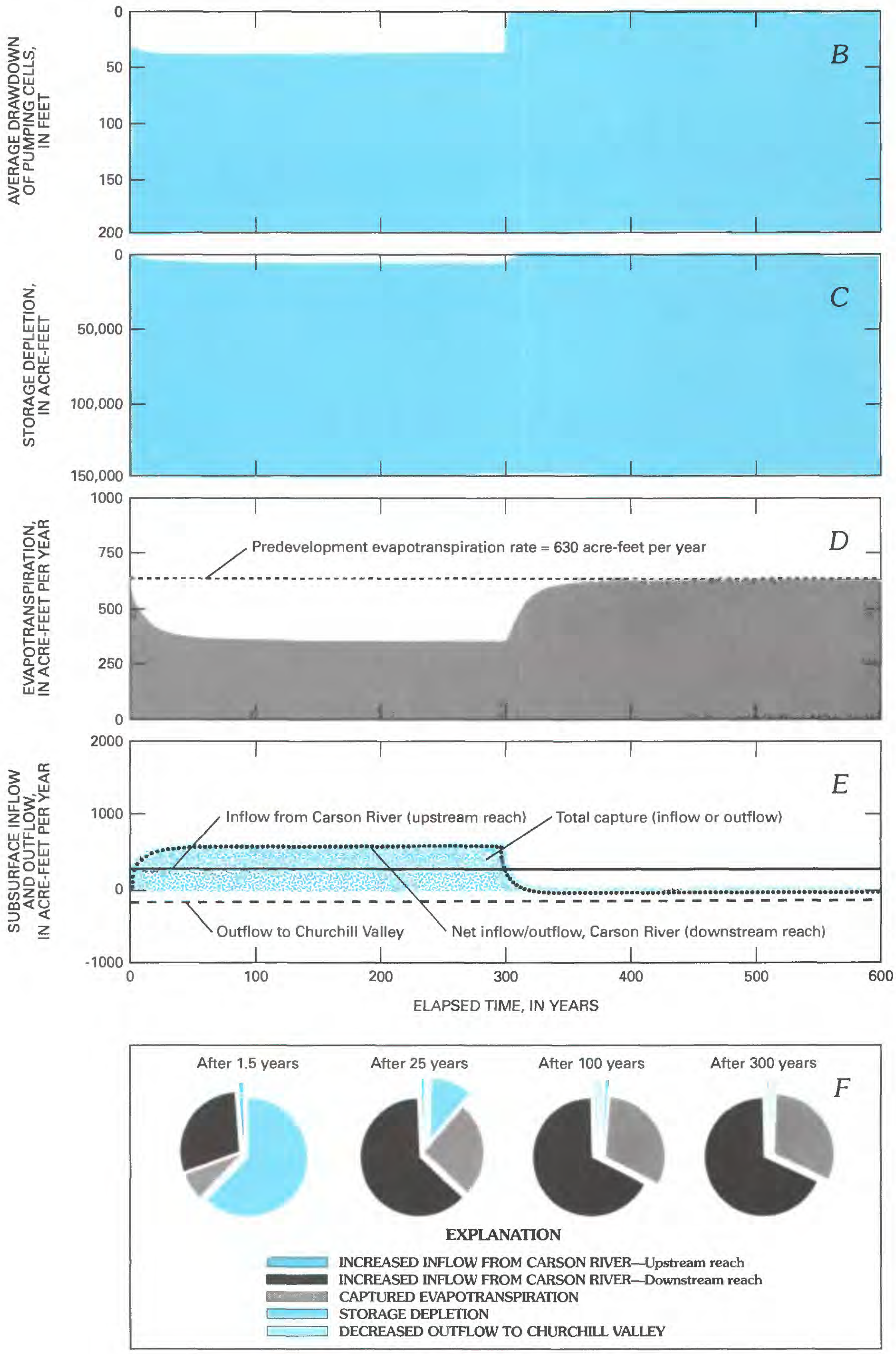


FIGURE 33.—Scenario H: Simulated response with pumping cells concentrated near Carson River. Pumping rate is 920 acre-feet per year, equal to estimated average predevelopment inflow or outflow. *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.



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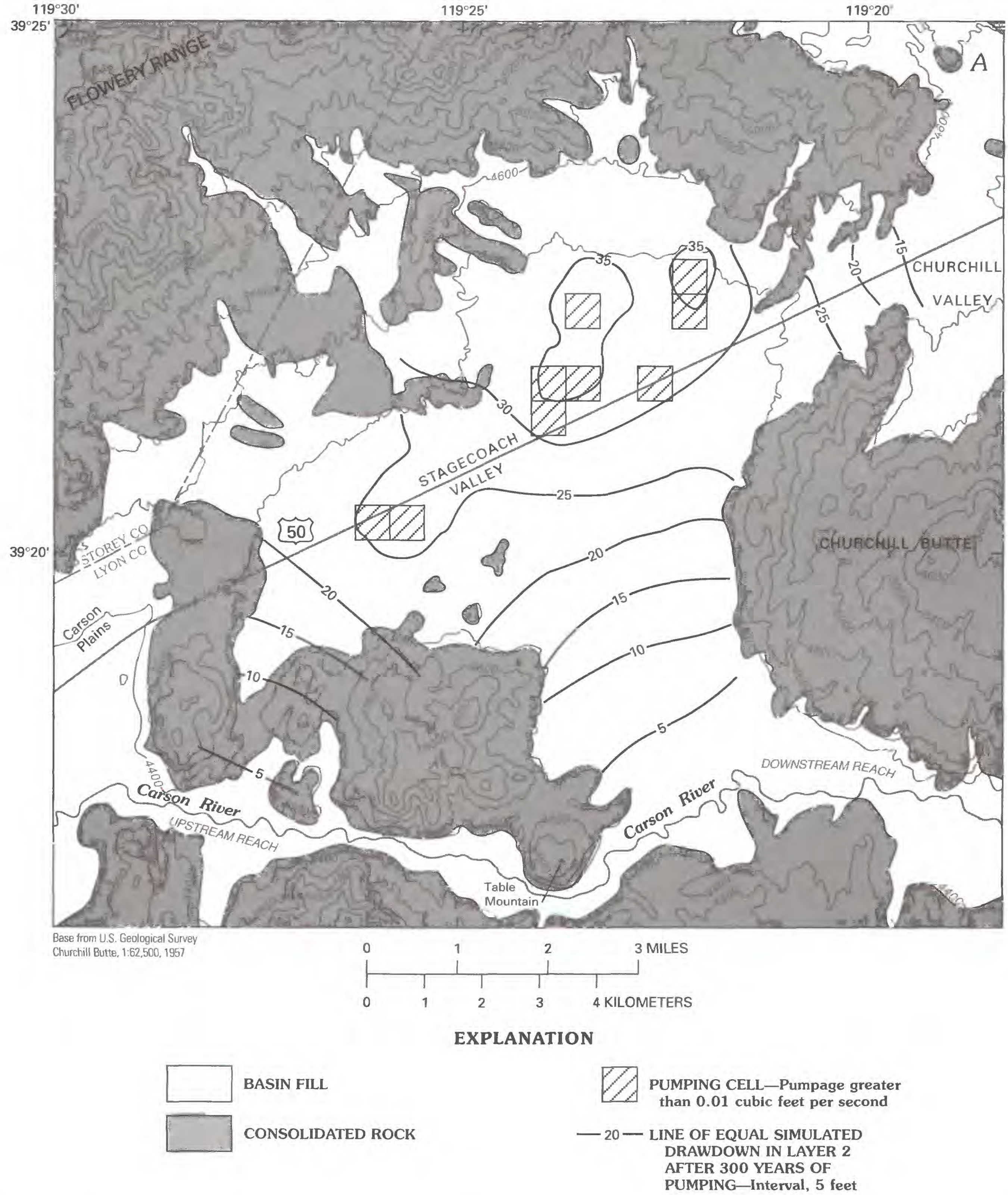

FIGURE 34.—Scenario I: Simulated response using 1981 well locations. Net pumping rate is equal to 1981 rate (about 966 acre-feet per year). *A*, Predicted drawdown in layer 2 after 300 years of pumping. *B*, Average drawdown of pumping cells. *C*, Cumulative storage depletion. *D*, Evapotranspiration rates. *E*, Rates of subsurface inflow and outflow. *F*, Sources of pumped water.

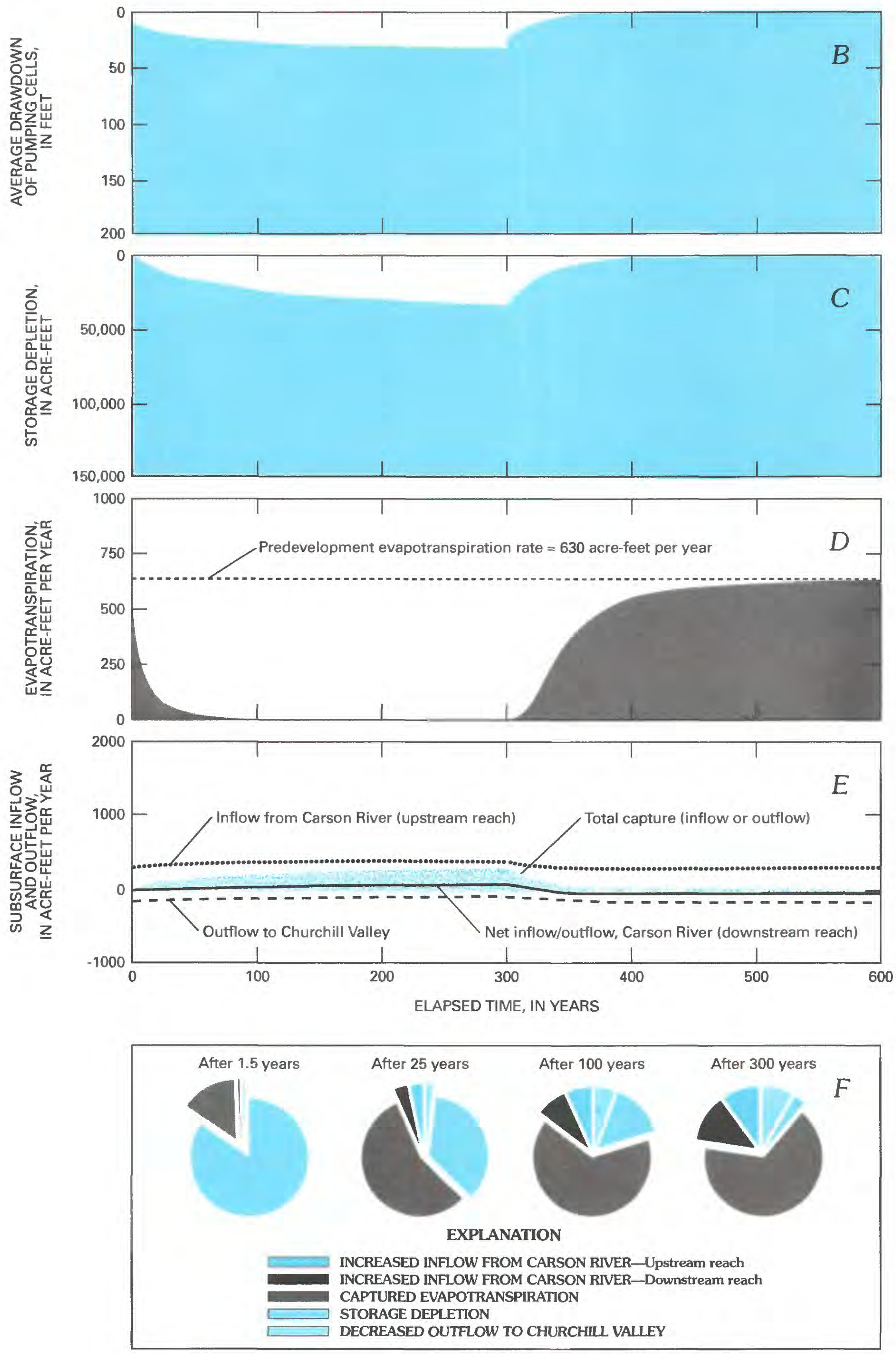

FIGURE 34.—Continued.

TABLE 10.—*Ground-water budgets for scenario A*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	287	324	338	356	379
Inflow from Carson River (downstream reach)	86	88	110	122	139	164
Subsurface inflow from Churchill Valley	—	—	—	—	—	—
Total	916	922	981	1,007	1,042	1,090
Outflow						
Subsurface outflow to Churchill Valley	166	165	153	141	127	109
Net pumpage	—	920	920	920	920	920
Evapotranspiration	635	447	71	23	4	—
Outflow to Carson River (downstream reach)	116	115	107	104	101	95
Total	917	1,647	1,251	1,188	1,152	1,124
Net results						
Net outflow-inflow[2]	1	725	270	181	110	34
Simulated storage depletion	—	717	265	176	106	20

[1]Predevelopment conditions, based on steady-state simulation.
[2]Ideally, the difference between simulated estimates of inflow and outflow is equal to simulated storage depletion, assuming outflow exceeds inflow. Discrepancies represent errors due primarily to truncation and rounding during iterative calculations performed by computer. These errors are independent of errors associated with estimation of recharge, discharge, or aquifer properties.

TABLE 11.—*Ground-water budgets for scenario B*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	302	424	510	610	738
Inflow from Carson River (downstream reach)	86	91	162	238	340	478
Subsurface inflow from Churchill Valley	—	—	—	—	—	40
Total	916	940	1,133	1,295	1,497	1,803
Outflow						
Subsurface outflow to Churchill Valley	166	164	132	94	40	—
Net pumpage	—	1,840	1,840	1,840	1,840	1,840
Evapotranspiration	635	299	—	—	—	—
Outflow to Carson River (downstream reach)	116	113	95	88	81	72
Total	917	2,416	2,067	2,022	1,961	1,912
Net results						
Net outflow-inflow[2]	1	1,476	934	727	464	109
Simulated storage depletion	—	1,471	927	720	457	99

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 12.—*Ground-water budgets for scenario C*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures; negative values indicate that water is being added to storage. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	302	424	510	435	403
Inflow from Carson River (downstream reach)	86	91	162	238	218	186
Subsurface inflow from Churchill Valley	—	—	—	—	—	—
Total	916	940	1,133	1,295	1,200	1,136
Outflow						
Subsurface outflow to Churchill Valley	166	164	132	94	82	91
Net pumpage	—	1,840	1,840	1,840	920	920
Evapotranspiration	635	299	—	—	—	—
Outflow to Carson River (downstream reach)	116	113	95	88	89	94
Total	917	2,416	2,067	2,022	1,091	1,105
Net results						
Net outflow-inflow[2]	1	1,476	934	727	–109	–31
Simulated storage depletion	—	1,471	927	720	–104	–12

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 13.—*Ground-water budgets for scenario D*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	277	298	310	324	346
Inflow from Carson River (downstream reach)	86	89	112	129	152	184
Subsurface inflow from Churchill Valley	—	—	—	—	—	—
Total	916	913	957	986	1,023	1,077
Outflow						
Subsurface outflow to Churchill Valley	166	165	151	136	117	94
Net pumpage	—	920	920	920	920	920
Evapotranspiration	635	460	101	62	30	8
Outflow to Carson River (downstream reach)	116	115	106	103	98	91
Total	917	1,660	1,278	1,221	1,165	1,113
Net results						
Net outflow-inflow[2]	1	747	321	235	142	36
Simulated storage depletion	—	745	315	229	134	25

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 14.—*Ground-water budgets for scenario E*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	278	331	395	495	637
Inflow from Carson River (downstream reach)	86	92	171	255	368	523
Subsurface inflow from Churchill Valley	—	—	—	—	—	84
Total	916	917	1,049	1,197	1,410	1,791
Outflow						
Subsurface outflow to Churchill Valley	166	164	123	76	11	—
Net pumpage	—	1,840	1,840	1,840	1,840	1,840
Evapotranspiration	635	321	23	—	—	—
Outflow to Carson River (downstream reach)	116	113	93	86	79	69
Total	917	2,438	2,079	2,002	1,930	1,909
Net results						
Net outflow-inflow[2]	1	1,521	1,030	805	520	118
Simulated storage depletion	—	1,516	1,024	798	513	113

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 15.—*Ground-water budgets for scenario F*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (downstream reach)	283	274	284	292	306	326
Inflow from Carson River (downstream reach)	86	86	102	119	144	183
Subsurface inflow from Churchill Valley	—	—	—	—	—	—
Total	916	907	933	958	997	1,056
Outflow						
Subsurface outflow to Churchill Valley	166	164	140	116	88	57
Net pumpage	—	920	920	920	920	920
Evapotranspiration	635	563	204	117	66	26
Outflow to Carson River (downstream reach)	116	116	109	105	99	91
Total	917	1,763	1,373	1,258	1,173	1,094
Net results						
Net outflow-inflow[2]	1	856	440	300	176	38
Simulated storage depletion	—	849	434	294	170	31

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 16.—*Ground-water budgets for scenario G*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	274	300	338	429	582
Inflow from Carson River (downstream reach)	86	87	138	211	328	499
Subsurface inflow from Churchill Valley	—	—	—	—	43	148
Total	916	908	985	1,026	1,347	1,776
Outflow						
Subsurface outflow to Churchill Valley	166	163	99	36	—	—
Net pumpage	—	1,840	1,840	1,840	1,840	1,840
Evapotranspiration	635	497	79	14	14	—
Outflow to Carson River (downstream reach)	116	115	100	89	81	70
Total	917	2,615	2,118	1,979	1,935	1,910
Net results						
Net outflow-inflow[2]	1	1,707	1,133	883	588	134
Simulated storage depletion	—	1,700	1,127	878	566	125

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 17.—*Ground-water budgets for scenario H*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	274	280	281	281	282
Inflow from Carson River (downstream reach)	86	323	607	635	643	645
Subsurface inflow from Churchill Valley	—	—	—	—	—	—
Total	916	1,144	1,434	1,463	1,471	1,474
Outflow						
Subsurface outflow to Churchill Valley	166	165	162	160	159	159
Net pumpage	—	920	920	920	920	920
Evapotranspiration	635	567	400	367	352	350
Outflow to Carson River (downstream reach)	116	77	60	58	58	58
Total	917	1,729	1,542	1,505	1,489	1,487
Net results						
Net outflow-inflow[2]	1	585	108	42	18	13
Simulated storage depletion	—	582	103	38	9	1

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

TABLE 18.—*Ground-water budgets for scenario I*

[All data values are in acre-feet per year and are considered reasonable only to two significant figures. —, none or negligible]

	Pre-1971 conditions[1]	Simulated elapsed pumping time				
		1.5 years	25 years	50 years	100 years	300 years
Inflow						
Recharge from precipitation	547	547	547	547	547	547
Inflow from Carson River (upstream reach)	283	282	314	327	346	377
Inflow from Carson River (downstream reach)	86	88	108	124	145	180
Subsurface inflow from Churchill Valley	—	—	—	—	—	—
Total	916	917	969	998	1,038	1,104
Outflow						
Subsurface outflow to Churchill Valley	166	164	146	130	110	85
Net pumpage	—	966	966	966	966	966
Evapotranspiration	635	492	93	34	9	—
Outflow to Carson River (downstream reach)	116	115	107	104	99	92
Total	917	1,737	1,312	1,234	1,184	1,143
Net results						
Net outflow-inflow[2]	1	820	343	236	146	39
Simulated storage depletion	—	818	339	232	139	29

[1]Predevelopment conditions, based on steady-state simulation.
[2]See footnote 2 of table 10.

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