GEOHYDROLOGY OF THE REGIONAL AQUIFER SYSTEM, WESTERN SNAKE RIVER PLAIN, SOUTHWESTERN IDAHO

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1408-G

Geohydrology of the Regional Aquifer System, Western Snake River Plain, Southwestern Idaho

By GARTH D. NEWTON

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

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

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

Dallas L. Peck
Director

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METRIC CONVERSION FACTORS

For readers who wish to convert measurements from the inch-pound system of units to the metric system of units, the conversion factors are listed below:

Multiply inch-pound unit	By	To obtain metric unit
acre	4,047	square meter (m^2)
acre-foot	1,233	cubic meter (m^3)
cubic foot per second (ft^3/s)	.02832	cubic meter per second (m^3/s)
foot (ft)	.3048	meter (m)
foot squared per day (ft^2/d)	.0929	meter squared per day (m^2/d)
gallon per minute (gal/min)	.06309	liter per second (L/s)
inch (in.)	25.4	millimeter (mm)
mile (mi)	1.609	kilometer (km)
square foot (ft^2)	.0929	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)

The conversion of degrees Celsius (°C) to degrees Fahrenheit (°F) is based on the equation

$$°F = (1.8)(°C) + 32.$$

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

GEOHYDROLOGY OF THE REGIONAL AQUIFER SYSTEM, WESTERN SNAKE RIVER PLAIN, SOUTHWESTERN IDAHO

By Garth D. Newton

ABSTRACT

The regional aquifer system in the western Snake River Plain is composed of three major rock units: (1) an upper unit of sedimentary and volcanic rocks about 500 feet thick, (2) a middle unit of fine-grained sedimentary and volcanic rocks about 4,000 feet thick, and (3) a lower unit of volcanic rocks about 7,000 feet thick. Ground-water recharge in the western plain is from infiltration of surface water used for irrigation, ground-water underflow across boundaries of the plain, and precipitation on the plain. Most ground water discharges to rivers and surface drains.

In 1980, recharge from irrigation was estimated to be about 1,400,000 acre-feet; from ground-water underflow, 310,000 acre-feet; and from precipitation on the plain, 40,000 acre-feet. Ground-water discharge to rivers and drains was estimated to be 1,450,000 acre-feet. Ground-water pumpage accounted for about 300,000 acre-feet. Almost no water leaves the western plain as underflow.

Ground-water levels in wells were measured during the spring of 1980. Because many wells are not cased, some measured water levels are a composite of heads in different aquifers. Only a few wells are completed in the middle and lower rock units.

A three-dimensional model employing a finite-difference technique was used to simulate ground-water flow in the regional aquifer system in the western Snake River Plain. Transmissivity, storage coefficient or specific yield, and vertical hydraulic conductivity for each of the three rock units were estimated using steady-state and transient model analyses. Limits were placed on estimated values of these hydraulic properties on the basis of published data for similar rock types and measured values where available. Specific yield of the upper unit (unconfined sand-and-gravel aquifer) was estimated to be about 0.10. Model-simulated transmissivity values for the upper rock unit ranged from 1,500 to 21,500 feet squared per day and, for the middle unit (sedimentary rock aquifer), from 900 to 12,100 feet squared per day. The lower unit (volcanic rock aquifer) was assigned a uniform transmissivity value of 8,600 feet squared per day. Storage coefficients for the middle and upper units were 4×10^{-3} and 7×10^{-3}, respectively.

Simulation showed that water from the volcanic rock unit leaks into overlying sedimentary rock units and that vertical hydraulic conductivity ranges from about 9 to 900 feet per day.

Data on transmissivity, storage capacity, vertical hydraulic conductivity, and ground-water underflow were not available for much of the modeled area. This general lack of data, combined with poor estimates of ground-water discharge to rivers and drains, hampered model calibration. Therefore, model results need to be evaluated with caution, except for the shallow flow system (upper sand-and-gravel unit) in the Boise River valley, where more data were available.

INTRODUCTION

This report is one in a series resulting from the U.S. Geological Survey's Snake River Plain RASA (Regional Aquifer-System Analysis) study that began in October 1979.

Lindholm (1981) presented a plan of study for the Snake River Plain RASA. Preliminary interpretive reports generated by the RASA study to date (1988) include (1) a regional water-table map and description of the ground-water-flow system (Lindholm and others, 1983; 1988); (2) a description of the geohydrologic framework (Whitehead, 1986); (3) an examination of water budgets and flow in the Snake River (Kjelstrom, 1986); (4) a description of water withdrawals for irrigation (Bigelow and others, 1986); (5) a map of land use showing irrigated acreage (Lindholm and Goodell, 1986); (6) a description of surface- and ground-water quality (Low, 1987); and (7) a ground-water-flow model of the eastern Snake River Plain (Garabedian, 1986).

Final interpretive results of the RASA study are presented in U.S. Geological Survey Professional Paper 1408, which consists of seven chapters as follows:

Chapter A is a summary of the aquifer system.

Chapter B describes the geohydrologic framework, hydraulic properties of rocks composing the framework, and geologic controls on ground-water movement.

Chapter C describes ground-water/surface-water relations and ground-water budgets.

Chapter D describes solute geochemistry of the cold-water and geothermal-water systems.

Chapter E describes water use.

Chapter F describes geohydrology and results of ground-water-flow modeling of the eastern Snake River Plain.

Chapter G (this report) describes geohydrology and results of ground-water-flow modeling of the western Snake River Plain.

PURPOSE AND SCOPE

The purposes of the study described in this report (chapter G) are to (1) define the geohydrology of the western Snake River Plain, incorporating new data and results of previous geologic, geophysical, and hydrologic investigations; and (2) develop a computer model of the regional ground-water system to simulate steady-state and transient hydrologic conditions.

The scope of the work included water-level measurements to determine the potentiometric surface and direction of ground-water flow, computation of recharge and discharge rates, and estimation of horizontal and vertical hydraulic conductivity, transmissivity, and storage capacity of regional aquifers.

DESCRIPTION OF THE STUDY AREA

The Snake River Plain is an arcuate topographic and structural depression that extends across southern Idaho. Geology and hydrology of eastern and western parts of the Snake River Plain are distinctly different; the west is predominantly sedimentary rocks, and the east is predominantly volcanic rocks. The line of separation between the eastern and western parts follows Salmon Falls Creek from the southern boundary of the plain to the Snake River, and along the Snake River to King Hill near the northern boundary. The study area described in this report includes the western part of the Snake River Plain and surrounding mountains (fig. 1). The boundary of the ground-water-flow model approximates the boundary of the western Snake River Plain from the junction of the Payette River with the Snake River southeast to the junction of Salmon Falls Creek with the Snake River.

The 144-mi-long, 50-mi-wide western Snake River Plain is flat relative to the Boise and Owyhee Mountains that border it on the northeast and southwest, respectively. Land-surface altitudes range from more than 5,000 ft above sea level south of Glenns Ferry to less than 2,000 ft near Weiser in the extreme northwestern part of the plain. A small part of the western Snake River Plain extends into Oregon.

The western Snake River Plain is drained by three major rivers. The largest is the Snake River, which crosses the entire study area from the mouth of Salmon Falls Creek to Weiser. The Boise and Payette Rivers are major tributaries of the Snake River. The central part of the plain in the vicinity of Mountain Home is a high plateau. No perennial rivers cross the plateau, and ephemeral drainage is to the Snake River.

Along the south side of the Mountain Home plateau to about Swan Falls Dam, the Snake River flows northwestward in a canyon as deep as 700 ft. Altitude of the river at Swan Falls Dam is about 2,300 ft; adjacent lands north of the canyon rise to about 3,000 ft. Downstream from Swan Falls Dam, the Snake River is less entrenched and flows across a broad flood plain about 100 ft below the general level of adjacent lands. Northwest of Swan Falls Dam, the plain is mainly a series of broad alluvial terraces with low, gentle slopes toward major rivers.

PREVIOUS INVESTIGATIONS

Many previous geologic and hydrologic studies described all or part of the western Snake River Plain, but none included computer ground-water-flow modeling. Russell (1902) described the geology and water resources of the Snake River Plain in an early regional study. The report is general and does not include quantitative data on hydraulic properties of the ground-water system. A more detailed summary of the development of ground water in the Snake River Plain was given by Mundorff and others (1964).

Lindgren (1898a, b) and Lindgren and Drake (1904) made early geologic studies of the Boise River valley; Piper (1924) studied the Bruneau plateau. Kirkham (1931) proposed that the Snake River Plain is a great structural depression, which he called the Snake River Plain downwarp. More recently, Malde (1965, p. 255) described the western Snake River Plain as a northwest-trending

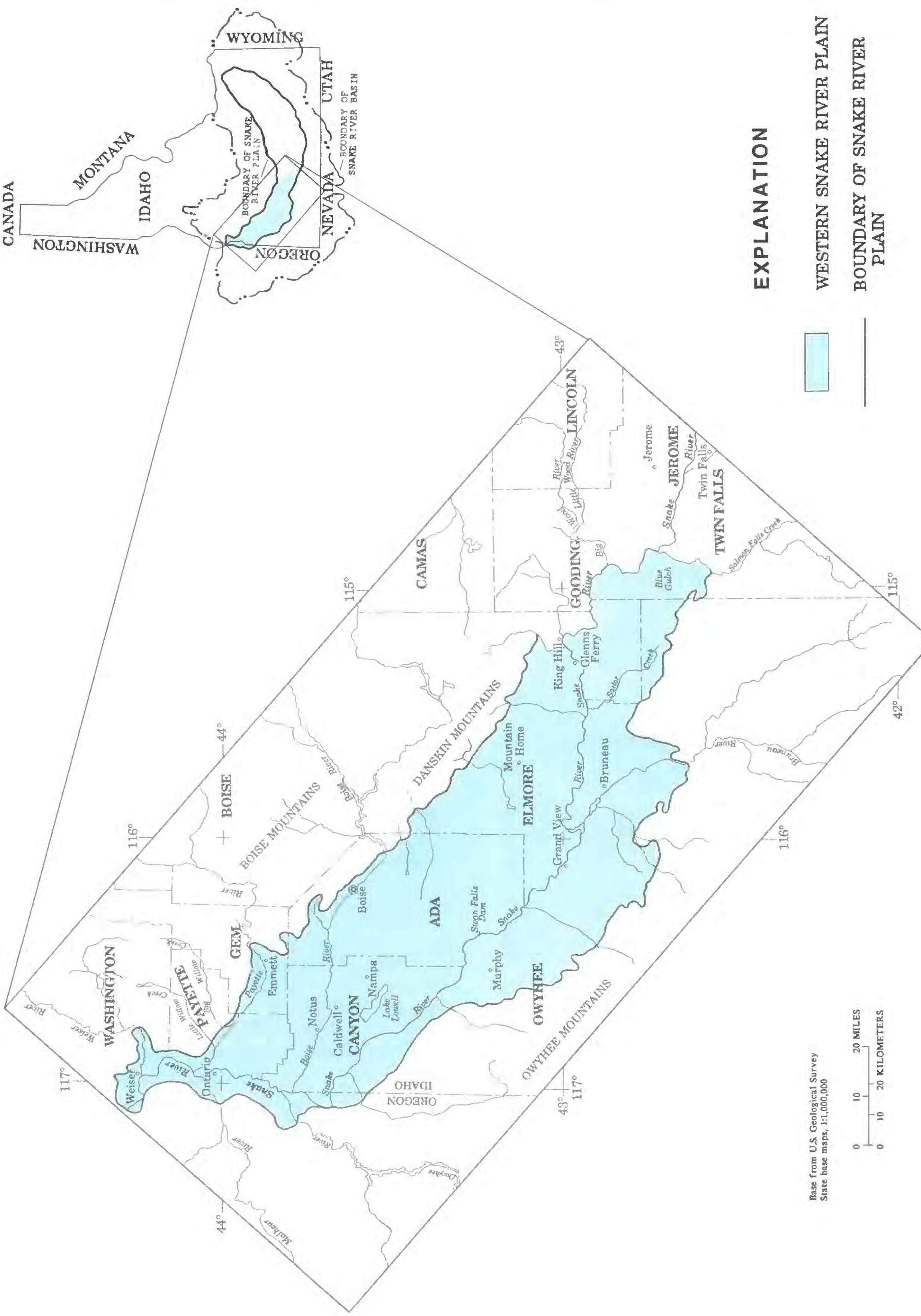


FIGURE 1.—Study area and vicinity.

graben. Wood and Anderson (1981, p. 9) supported that interpretation on the basis of lithologic logs from deep drilling, seismic-reflection profiles, and geologic mapping in the Nampa–Caldwell area.

Malde and others (1963) mapped geology of the west-central Snake River Plain. Mabey and others (1974) compiled a preliminary gravity map of southern Idaho, and Mabey (1976) published an interpretation of the original map.

Nace and others (1957) studied ground-water hydrology and geology of the Nampa–Caldwell area and described a plan to irrigate land near Mountain Home with water from the Boise River in exchange for ground water pumped from the Boise River valley. Littleton and Crosthwaite (1957) studied the geology and hydrology of the Bruneau–Grand View area, and Crosthwaite (1963) examined the Sailor Creek area. Ralston and Chapman (1969) studied northern Owyhee County and Ada and Elmore Counties, and Chapman and Ralston (1970) studied ground-water resources of the Blue Gulch area. Dion (1972) reported on the shallow ground-water system in the Boise–Nampa area.

As early as 1890, natural hot-water springs near Boise and the Snake River prompted development of geothermal water for heating and bathing. Geothermal water in the Bruneau–Grand View area is the primary source for irrigation. Young (1977) described the geothermal system near Mountain Home; Young and others (1979) studied ground-water discharge and heat flow in the Bruneau–Grand View area, and Young and Lewis (1980) described the chemistry of geothermal water in southwestern Idaho. Studies by Burnham (1979) and Wood and Anderson (1981) focused on more recent geothermal developments near Boise; their reports include detailed geologic logs of wells drilled to heat commercial buildings in downtown Boise.

Few data on hydraulic properties are available for aquifers in the western Snake River Plain. Nace and others (1957, p. 55) listed results of a few aquifer tests in the Boise River valley. Results of several tests on geothermal wells in the Boise River valley are included in reports by Burnham (1979), Wood and Anderson (1981, p. 34), and Anderson (1981, p. 6).

WELL-NUMBERING SYSTEM

The well-numbering system used by the U.S. Geological Survey in Idaho indicates the location of wells within the official rectangular subdivision of the public lands, with reference to the Boise base line and Meridian. The first two segments of the number designate the township and range. The third segment gives the section number, followed by three letters and a number, which indicate the ¼ section (160-acre tract), the ¼-¼ section (40-acre tract), the ¼-¼-¼ section (10-acre tract), and the serial number of the well within the tract. Quarter sections are lettered A, B, C, and D in counterclockwise order from the northeast quarter of each section (fig. 2). Within the quarter sections, 40-acre and 10-acre tracts are lettered in the same manner. For example, well 6S-5E-24BCA1 is in the NE¼SW¼NW¼ sec. 24, T. 6 S., R. 5 E., and is the first well inventoried in that tract.

GEOLOGIC SETTING

The western Snake River Plain is a deep structural depression bounded by major northwest-trending faults (pl. 1). High mountains that surround the plain on the northeast and southwest are composed largely of Tertiary rhyolitic and basaltic rocks and Cretaceous granitic rocks of the Idaho batholith. Tertiary and Quaternary sedimentary and volcanic rocks fill the depression. A geologic map and section and generalized description of geologic units in the western Snake River Plain and surrounding area are shown on plate 1.

Tertiary volcanic rocks crop out along the margins of the plain and also are present beneath the plain. The most hydrologically significant are rhyolitic rocks that crop out south of the Bruneau–Grand View area. Tertiary volcanic rocks are as thick as 11,000 ft.

Tertiary and Quaternary sedimentary rocks (mainly the Idaho Group) are as thick as 5,000 ft near the Idaho–Oregon State line and overlie Tertiary volcanic rocks. The sediments are unconsolidated to poorly consolidated clay, silt, sand, volcanic ash, diatomite, freshwater limestone, and conglomerate of the Idaho Group with interflows of basalt. The sediments are mainly of fluviatile origin; finer grained materials accumulated in localized lake basins. Facies changes within the Idaho Group and other parts of the stratigraphic section complicate the subsurface geology and make correlation of individual stratigraphic units difficult. Unconsolidated Quaternary deposits of silt, sand, and gravel unconformably overlie sediments of the Idaho Group.

In the central part of the western plain near Mountain Home, Quaternary basalt overlies and is interfingered with Quaternary sediments. The basalt consists of fine- to coarse-textured fissure flows and pyroclastics.

Clay, silt, sand, and gravel associated with late Quaternary (Pleistocene and Holocene) floods are cemented with varying amounts of calcium carbonate. Coarse fragments consist of crystalline rocks including rhyolite, quartz monzonite, diorite, and arkose, and some scoria and basalt pebbles.

The youngest sediments in the area are clay, silt, sand, and gravel along present flood plains. The oldest of these sediments are in low terraces above the present flood plains.

GEOHYDROLOGY

The regional aquifer system in the western Snake River Plain is composed of three major rock units—upper and middle units of Tertiary and Quaternary sedimentary and volcanic rocks, and a lower unit of Tertiary volcanic rocks (pl. 1).

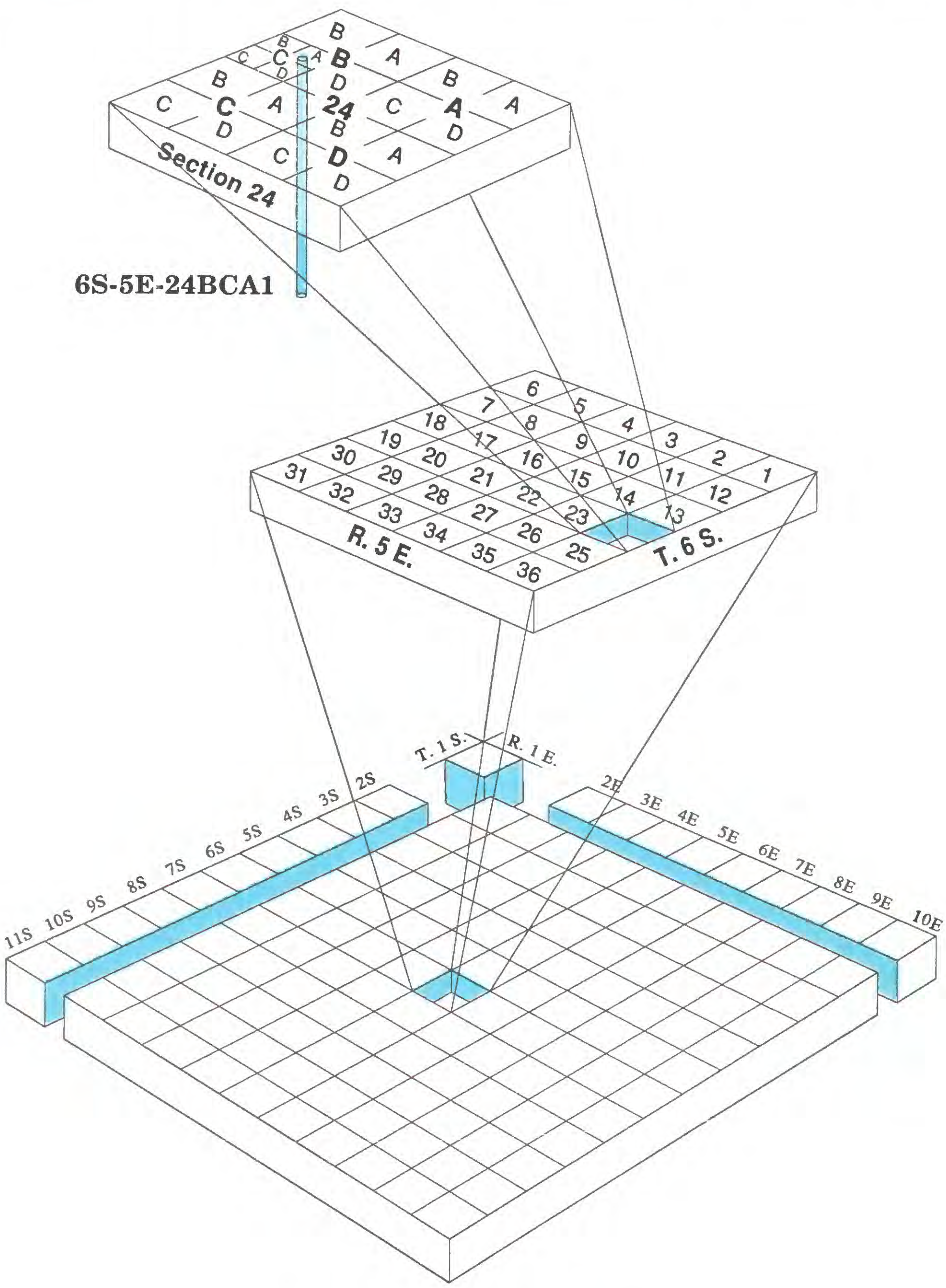


FIGURE 2.—Well-numbering system.

Most wells in the upper unit, which is about 500 ft thick, are completed in sand and gravel. In the Boise River valley, the upper unit also contains many discontinuous layers of clay that are locally confining. A layer of clay composes the base of the upper unit in the Boise River valley but is not present under the Mountain Home plateau. However, near Mountain Home, fine-grained sediments above layers of Quaternary basalt impede downward water movement and form perched aquifers.

The middle unit, consisting of fine-grained sedimentary and volcanic rocks, is about 4,000 ft thick. Although some layers are highly permeable, thick sequences of clay greatly reduce horizontal and vertical hydraulic conductivity. The sedimentary rocks of low hydraulic conductivity confine water in the underlying volcanic rocks (lower unit) and separate hot water in the lower unit from cold water in the upper unit. Water-temperature gradients in some wells increase sharply where wells penetrate layers of clay that confine the geothermal system.

In the lower unit, hydraulic conductivity is due largely to faults and fractures. The volcanic rocks supply geothermal water to numerous springs and wells, particularly south of the Snake River and in the Boise area. The lower unit is estimated to be about 7,000 ft thick.

WATER-TABLE CONFIGURATION AND GROUND-WATER MOVEMENT

The configuration of the water table shown in figure 3 is based on water levels measured in 1980 in aquifers that include discontinuous rock layers of low hydraulic conductivity. The direction of ground-water movement is controlled largely by topography and surface drainage. Movement (shown by arrows) is from topographic highs to rivers and other surface drains. Water-table contours parallel to the boundary of the plain indicate underflow from areas adjacent to the plain or recharge along the foothills. Regional water movement is to the Snake River; local movement is to the Boise River and other tributaries. The northeast-southwest section across the western plain (fig. 4) shows upward water movement in the Boise River valley. Measured water levels were assumed to represent the hydraulic head of the aquifer at the first opening in the well casing. These values of hydraulic head then were contoured to show head changes with depth, directions of ground-water movement, and areas of recharge and discharge.

Near Mountain Home, water levels in perched aquifers are as much as 200 ft higher than those in the regional aquifer system (Young, 1977). Water eventually moves downward from these perched aquifers into the regional system.

The middle and lower rock units in the western plain consist of a series of confined aquifers in Tertiary and Quaternary sedimentary and Tertiary volcanic rocks. Confined aquifers are recharged largely where these rocks crop out along the boundaries of the plain (pl. 1). Water discharges as leakage to upper layers and as spring flow along fault zones. Geothermal springs are common along the margins of the plain. Water is heated at depth to temperatures as high as 77 °C and moves upward along faults.

DEPTH TO WATER AND WATER-LEVEL CHANGES

In the spring of 1980, the U.S. Geological Survey measured water levels in several hundred wells in the western Snake River Plain. Some of these wells are completed in confined aquifers, and water in them rises above the base of the confining bed. In the Boise River valley, depth to water is several feet to several tens of feet below land surface. Most of these wells are less than 200 ft deep and are completed in unconfined sand-and-gravel aquifers. Near Mountain Home, many wells are completed in perched aquifers. In the Bruneau–Grand View area, some wells are as deep as 4,000 ft. Many wells are uncased, and water levels in them represent a composite head from more than one aquifer. The depth to water below land surface in the spring of 1980 is shown in figure 5. This figure provides general information on pumping lift.

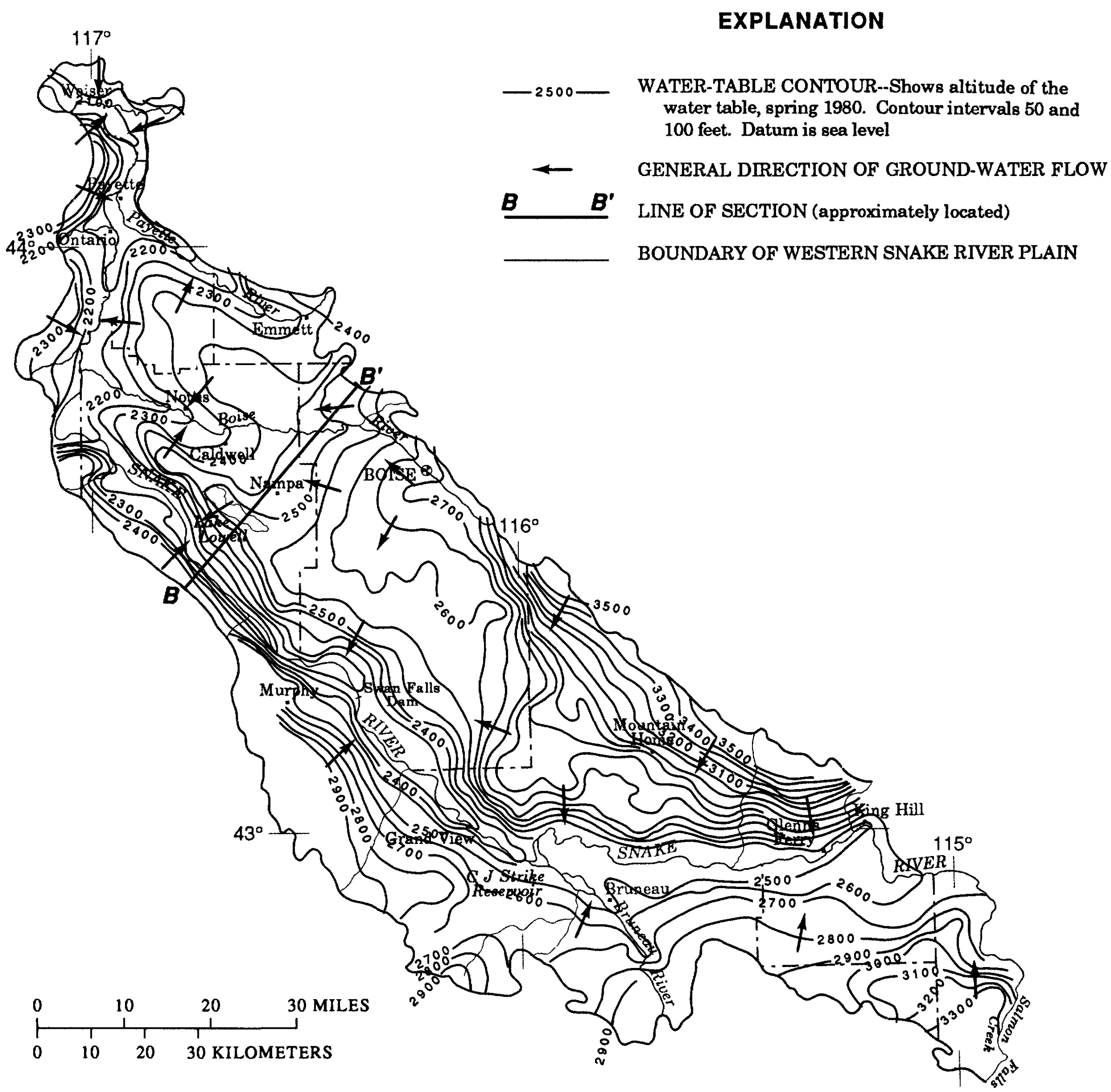
Since the early 1900's, water levels have changed considerably in much of the western plain (fig. 6). For example, the hydrograph for well 2N-1W-4DDA1 (fig. 6, well 3) in Ada County shows a rise in water level of about 80 ft since 1914.

Water levels have declined in some areas as a result of ground-water withdrawals. In Elmore County, ground-water levels have declined as much as 35 ft since 1966. For example, the water level in well 4S-10E-30BBA1 (fig. 6, well 8) has declined about 15 ft since 1967, and the water level in well 5S-4E-5CAA1 (fig. 6, well 9) has declined about 35 ft since 1966. Declines are evident in both shallow and deep aquifers. Near Mountain Home, water-level declines were severe enough that the Idaho Department of Water Resources closed the area to further ground-water development (Norton and others, 1982).

Blue Gulch near Salmon Falls Creek in Twin Falls County was closed to further ground-water development in 1970. Water levels in that area, as shown by the hydrograph for well 10S-12E-11DBD1 (fig. 6, well 12), have declined about 40 ft since 1968, and seasonal ground-water-level fluctuations have doubled in magnitude since 1968. The water level in well 7S-5E-19CCC1 (fig. 6, well 11), Owyhee County, has declined about 30 ft since 1953, and seasonal water-level fluctuations have increased from a few feet to about 10 ft. The water level in well 6S-3E-14BCB1 (fig. 6, well 10) has declined more than 40 ft since 1957, and seasonal fluctuations have increased to about 30 ft. Since 1980, water levels in parts of Owyhee County have risen as a result of increased precipitation.

HYDRAULIC PROPERTIES

Nace and others (1957, p. 54, 55) estimated from aquifer tests that the transmissivity of sand-and-gravel aquifers in the upper rock unit in the Boise River valley ranged from about 5,000 to about



Modified from Lindholm and others (1988)

FIGURE 3.—Configuration of the water table, spring 1980. (Ground-water-flow section shown on fig. 4.)

230,000 ft²/d (table 1). The storage coefficient of shallow, confined aquifers in the upper rock unit ranged from about 0.00007 to 0.006, and the specific yield of unconfined aquifers in this unit ranged from 0.025 to 0.43. Ferris and others (1962, p. 76, 78) stated that the storage coefficient of confined aquifers ranges from about 0.00001 to 0.001, and that the specific yield of unconfined aquifers ranges from about 0.05 to 0.30. The locations of wells where aquifer tests were conducted in the Boise River valley are shown in figure 5.

Chapman and Ralston (1970, p. 13) estimated that the transmissivity of sedimentary rocks composing the middle unit in the Blue Gulch area (fig. 1) west of Salmon Falls Creek ranged from about 500 to 2,280 ft²/d. They also estimated that the transmissivity of volcanic rocks composing the lower unit in the Blue Gulch area ranged from 8,000 to 112,000 ft²/d. Young (1977, p. 19) estimated that the average transmissivity of basalt in the upper unit near Mountain Home was about 50,000 ft²/d. Anderson (1981, p. 6) estimated that the transmissivity of welded tuff in the middle unit in the Boise River valley was about 670 ft²/d, and that the storage coefficient was 0.00005.

In general, transmissivity is highest in coarse-grained sedimentary and volcanic rocks in the upper unit. Transmissivity is lowest in fine-grained sedimentary rocks in the middle unit. The transmissivity of individual layers of sand and gravel or basalt may be higher than that of the rock units. The transmissivity of volcanic rocks in the lower unit probably is between the middle-unit and upper-unit values, although it may be greatly increased if the rocks are highly fractured or faulted. Much of the water movement in the lower unit is through faults and fractures parallel to the Snake River. Transmissivity estimated from aquifer tests along fault zones is typically higher than the regional transmissivity of volcanic rocks. Transmissivity is highest in the direction of faults.

Many factors affect specific capacity, such as well diameter, depth of well penetration into the aquifer, type and condition of well openings, meth-

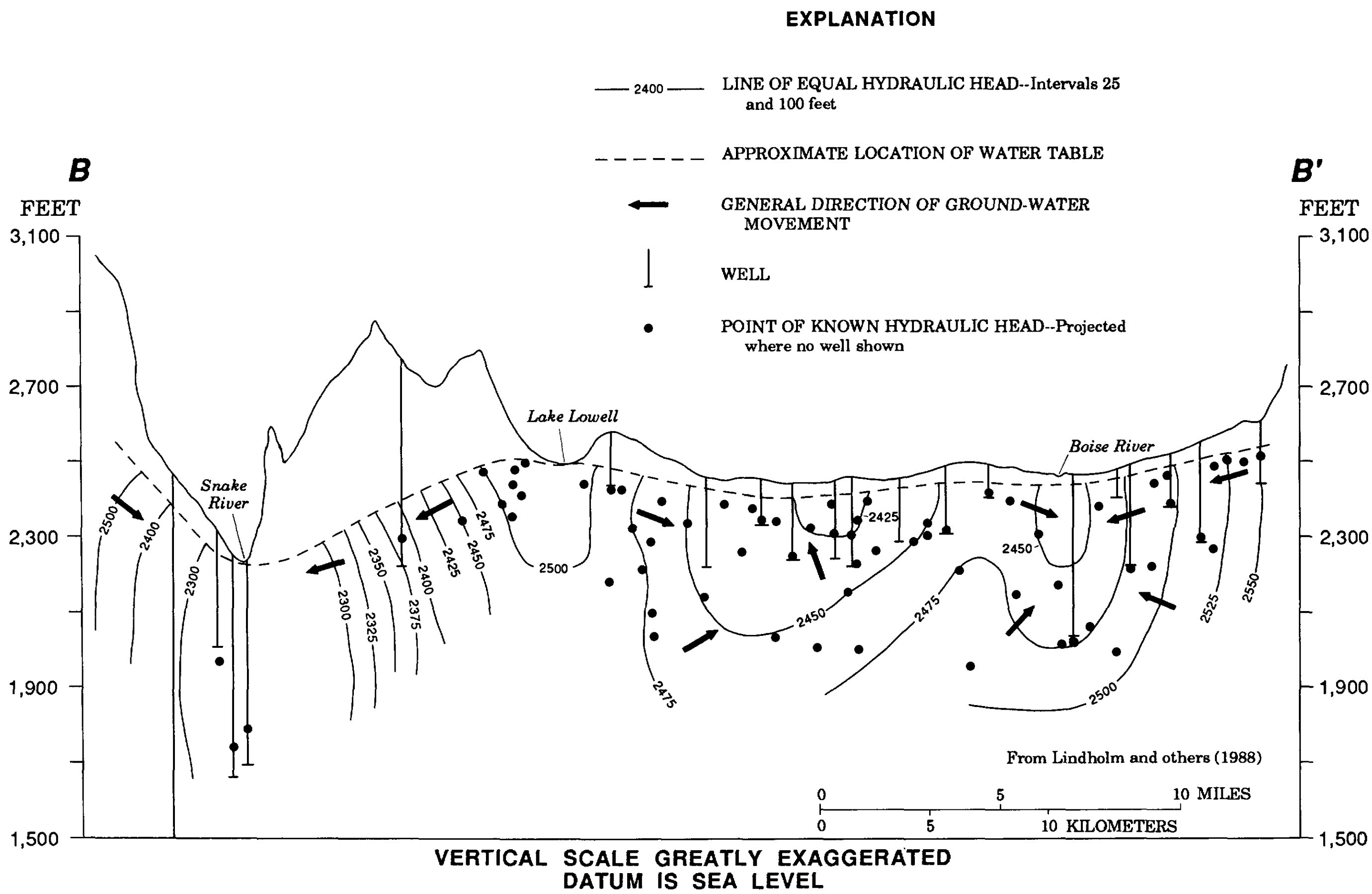


FIGURE 4.—Generalized section showing ground-water flow across the western Snake River Plain, 1980. (Line of section shown on fig. 3.)

od of development, length of pumping time, and hydraulic properties of the aquifer. Specific-capacity values used in this report were estimated from drillers' logs. Although estimated specific capacities varied greatly, average values for different rock units were consistent with transmissivity values obtained from aquifer tests. Locations of wells for which specific-capacity data are available are shown in figure 7.

Specific capacities of wells completed in sand and gravel in the upper rock unit near Mountain Home are 40 to 100 (gal/min)/ft and, in shallow alluvium in the Boise River valley, about 40 (gal/min)/ft. Specific capacities of wells completed in sand and gravel in the middle unit in the Boise River valley and along the Payette River are 10 to 20 (gal/min)/ft. Near the Snake River, specific capacities of wells in the middle unit generally are less than 10 (gal/min)/ft; these low values reflect the predominance of fine-grained sediments. Highest specific capacities average about 80 (gal/min)/ft and are for wells completed in Tertiary volcanic rocks in the lower unit in the Bruneau–Grand View area.

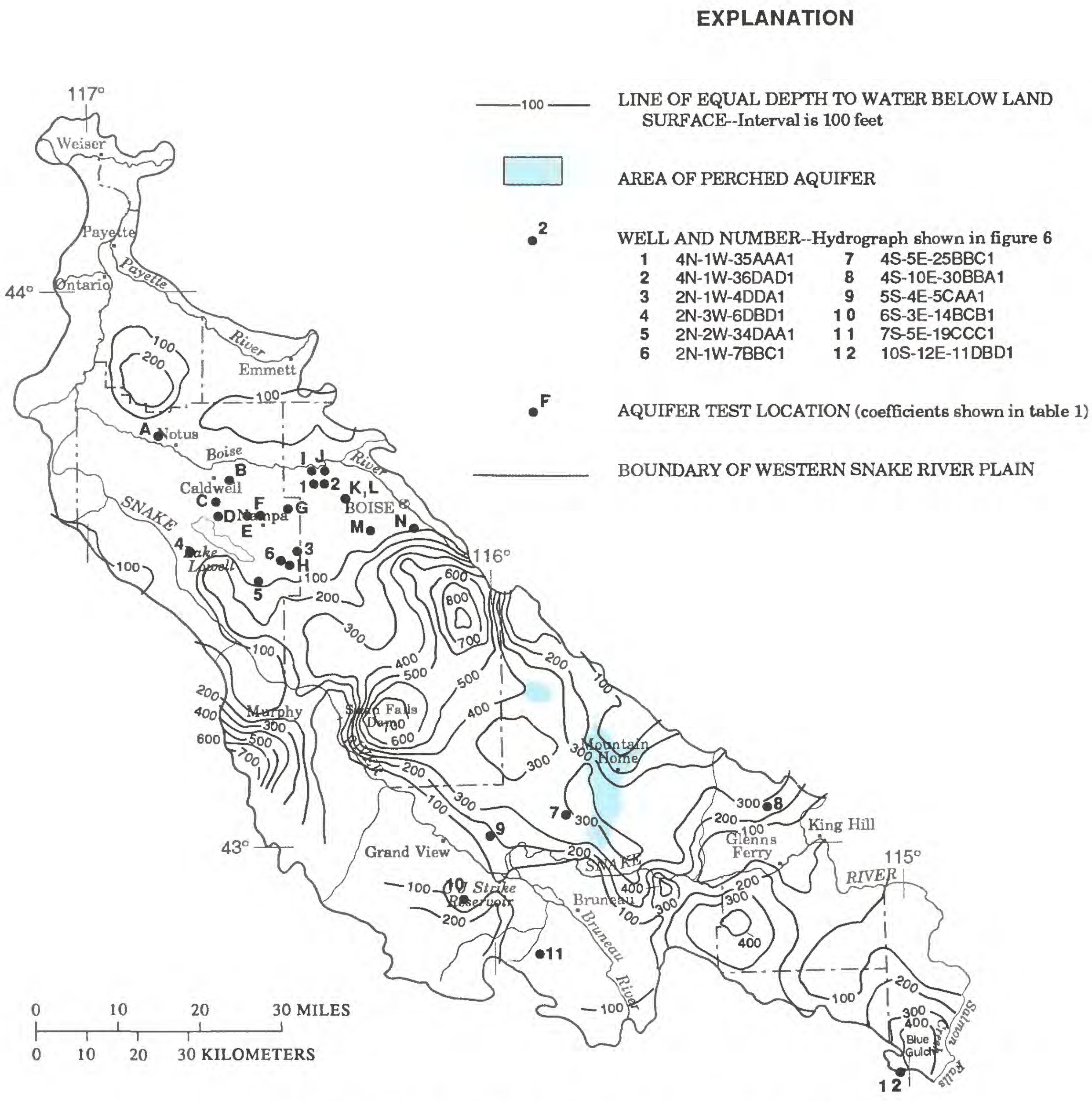


FIGURE 5.—Depth to water, spring 1980, and locations of wells.

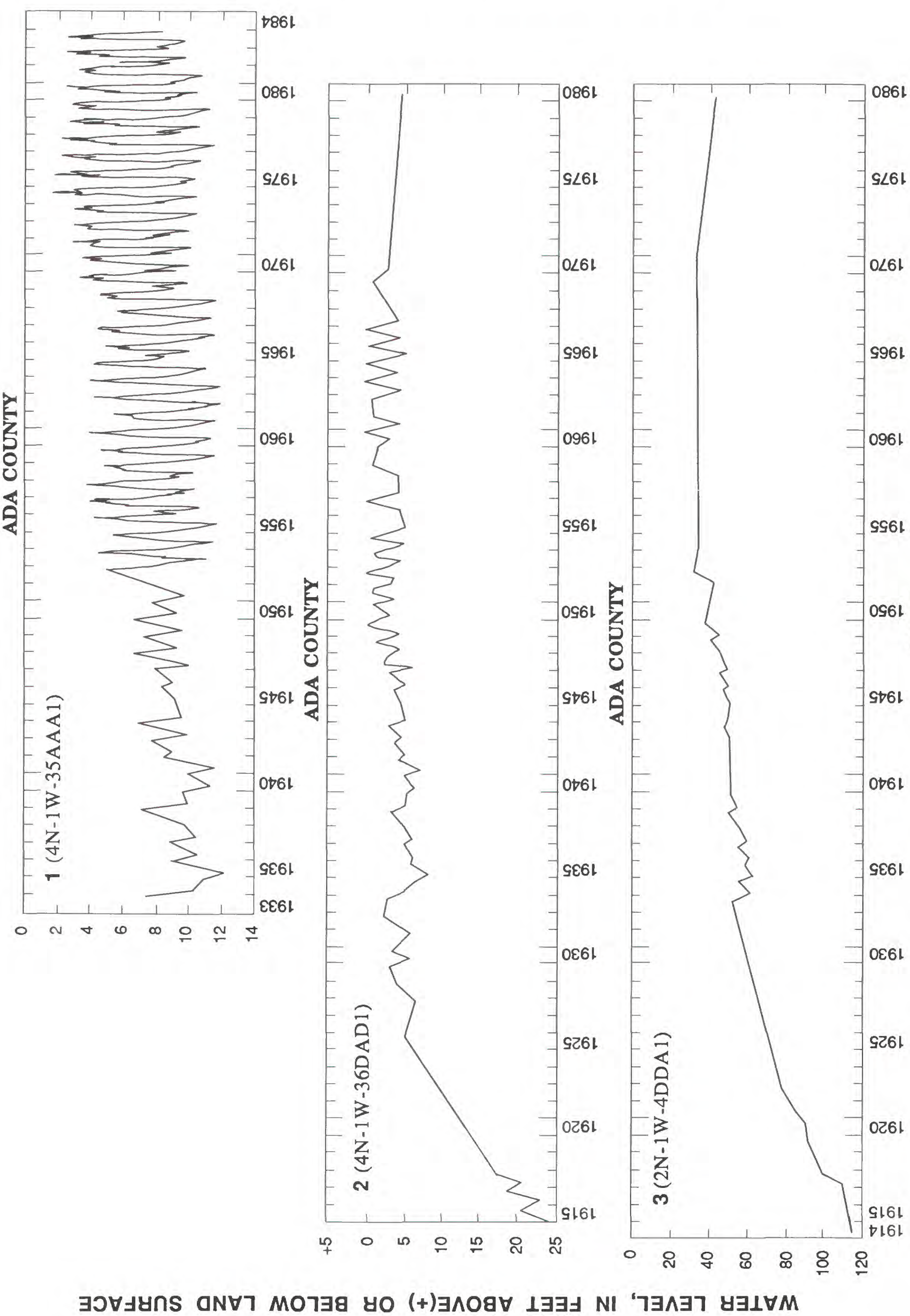


FIGURE 6.—Hydrographs for selected wells. (Locations of wells shown in fig. 5.)

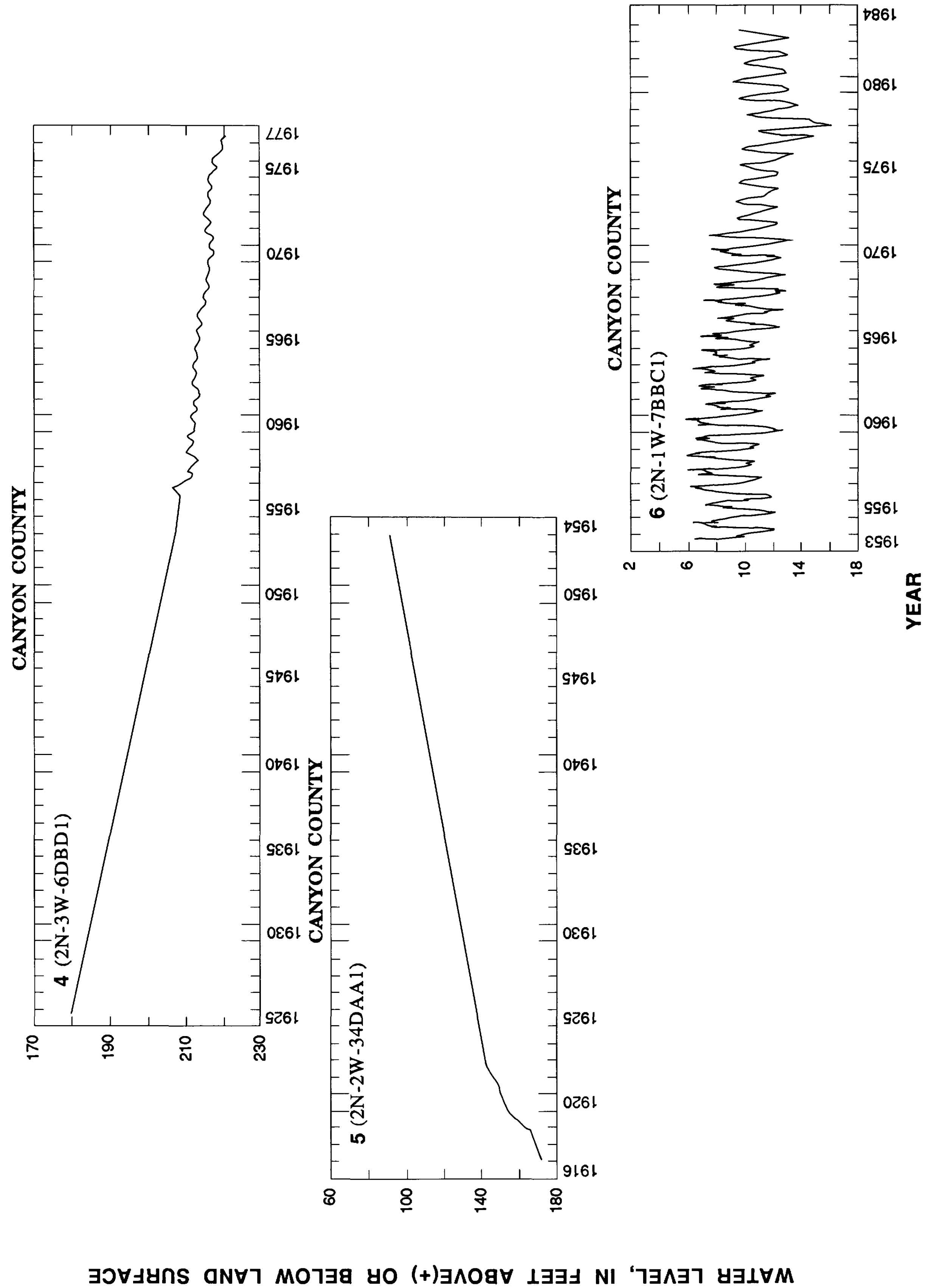


FIGURE 6.—*Continued.*

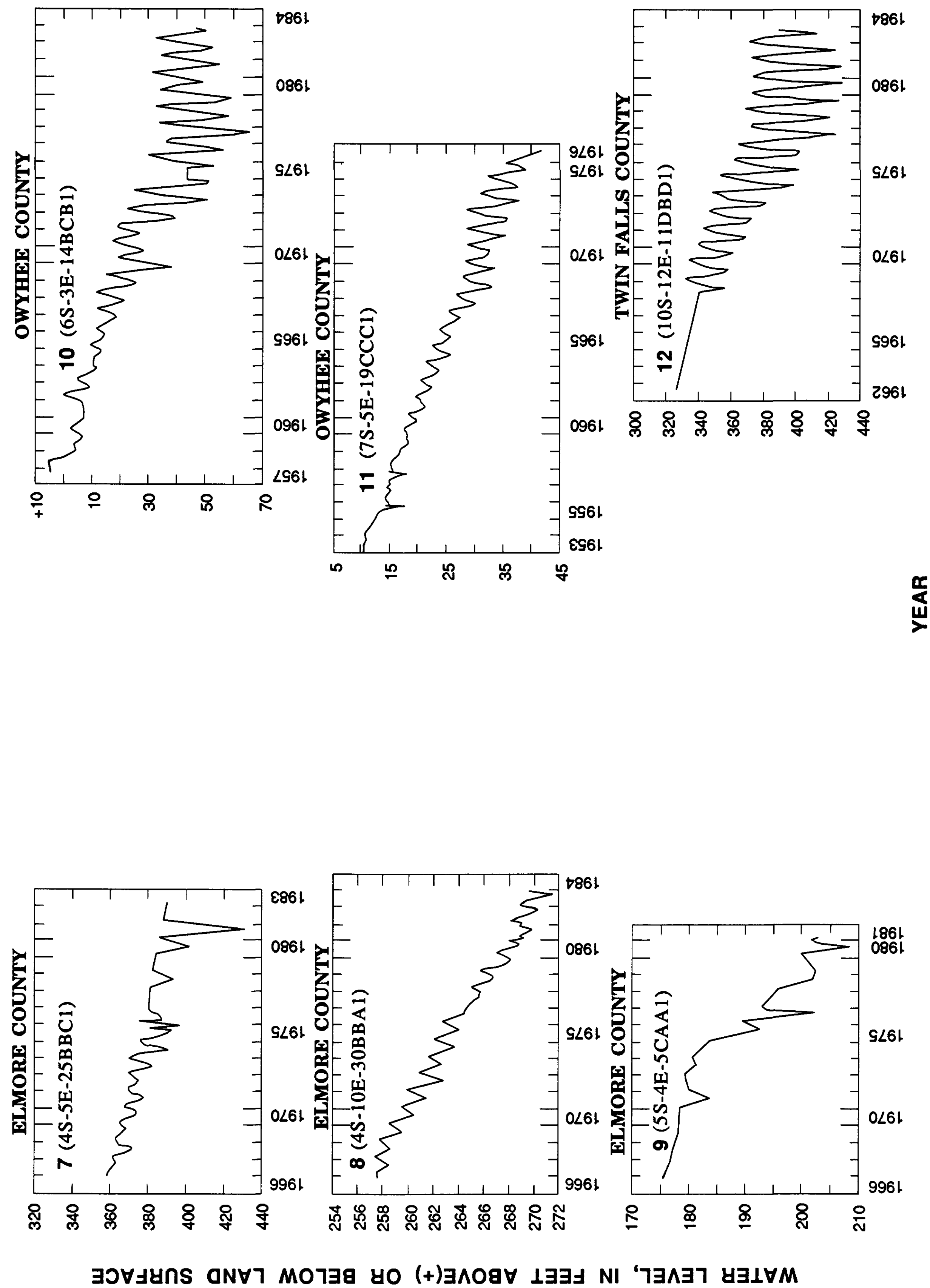


FIGURE 6.—*Continued.*

TABLE 1.—*Selected aquifer coefficients for the upper rock unit in the Boise River valley*

[Modified from Nace and others, 1957; locations shown in figure 5; —, data not available]

Well	Location	Transmissivity (feet squared per day)	Specific yield or storage coefficient
Canyon County			
A	5N-4W-28CC1	43,300	0.025
B	4N-3W-25DA1	27,800	.004
C	3N-3W-3BB1	128,400	.230
D	3N-3W-11DA1	160,400	.006
E	3N-2W-8CC1	18,200	.0006
F	3N-2W-9DD4	36,900	.0001
G	3N-1W-7BB1	22,100	.003
H	2N-1W-7BC4	227,300	.004
Ada County			
I	4N-1W-13DC1	7,200	—
J	4N-1W-13DC2	16,000	.001
K	3N-1E-5AB1	4,900	.001
L	3N-1E-5AB1	25,100	.006
M	3N-1E-36AD2	36,100	.00007
N	3N-2E-25BB1	17,400	.43

Specific-capacity data were used to estimate aquifer transmissivity according to a method developed by Theis and others (1963, p. 331–341). Because specific capacities of wells pumped at low rates generally are lower than those of wells pumped at high rates, only wells discharging more than 200 gal/min were used to estimate transmissivity. High average specific capacity generally indicates high transmissivity, and low specific capacity indicates low transmissivity. A comparison of hydraulic conductivities estimated from specific-capacity data with model-calculated estimates is made later, in the section "Model Development and Calibration." The distribution of transmissivity values calculated from specific-capacity data is shown in figure 8. Most transmissivity values are for the upper unit, where most wells are completed.

No vertical hydraulic-conductivity data are available for the study area. Values used in this study were estimated from reported values for rock types similar to those in the study area.

GROUND-WATER BUDGET, WESTERN SNAKE RIVER PLAIN, 1980

A ground-water budget was prepared for 1980 to define components of recharge and discharge. The range of uncertainty associated with the estimated ground-water budget (table 2) is large because some of the values used in budget estimates are not well defined. For example, recharge from irrigation water generally is estimated as the difference between applied irrigation water and losses due to ET (evapotranspiration) and excess water returned to canals and drains. Rates of ET in irrigated areas probably can be estimated reasonably well; however, rates of ET in nonirrigated areas cannot be estimated accurately owing to lack of data. Surface-water diversions and ground-water pumpage for irrigation generally are measured or estimated; however, the amount of irrigation water that returns to canals and drains is largely unknown and not measured. Even if flow in the canals and drains were known, the amount of irrigation water returned to them still would be difficult to estimate because that flow also contains a ground-water component that discharges through the surficial aquifers. For these reasons, an alternative approach toward estimating recharge from irrigation water was taken. This approach was to use estimates made by the U.S. Bureau of Reclamation of crop water requirements, irrigation-water delivery requirements, and losses from delivery systems in the Boise River valley. Further discussion on recharge estimates is presented in the subsection "Recharge."

Estimates of ground-water discharge to the Boise River and drains are considered fair because they are based on measured stream discharge during low-flow conditions. However, estimates of ground-water discharge to the Snake River are poor because discharge accounts for only 4 to 5 percent of the total river discharge. A small measurement error in Snake River discharge may result in a large error in computed ground-water discharge. Ground-water pumpage estimates are considered good because they were determined from power-consumption records for individual wells. The accuracy of pumpage estimates depends a great deal on estimates of the total pumping head, which includes depth to water (fig. 5) and pressure requirements for distribution.

Underflow across plain boundaries from tributary basins was computed as the difference between total discharge and estimated recharge from surface-water-irrigation return flow and from precipitation on the plain. Estimates of annual precipitation are considered good because long-term records from weather stations are available; however, estimates of recharge from precipitation generally are poor because many factors that affect infiltration of precipitation are not well determined. The distribution of underflow is poorly known. Water-table contours (fig. 3) indicate some underflow along the entire boundary of the plain. However, because of the more permeable nature of volcanic rocks north and south of Mountain Home, more underflow was assumed in these areas.

RECHARGE

Sources of recharge to the regional aquifer in the western Snake River Plain are infiltration of surface water used for irrigation, underflow across plain boundaries, and precipitation on the plain.

The largest source of ground-water recharge is infiltration of surface water diverted for irrigation. Of 826,000 acres irrigated in 1980, surface water was used to irrigate about 696,500 acres; ground water was used to irrigate the remaining 129,500 acres (fig. 9).

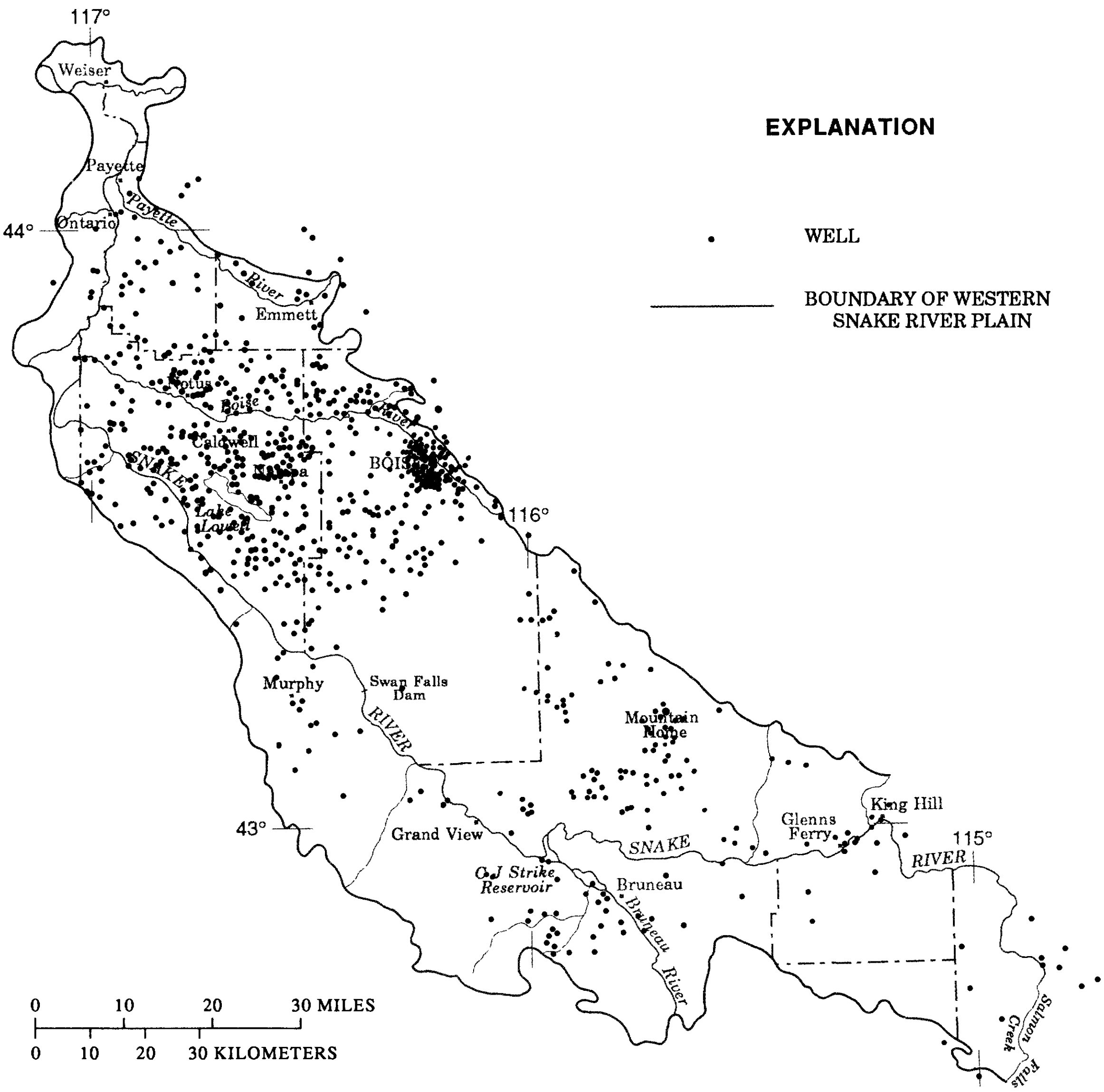


FIGURE 7.—Locations of wells for which specific-capacity data are available.

Surface-water diversions for irrigation from 1901 to 1980 are shown in figure 10. Total diversions in 1980 were 3,700,000 acre-ft. Most were gravity diversions, but about 669,000 acre-ft were pumped from rivers to irrigate adjacent lands (Bigelow and others, 1986). In some reaches, Snake River water is lifted as much as 500 ft.

Total ground-water recharge from surface-water irrigation in 1980 was estimated to be about 1,400,000 acre-ft. The budget equation in the section titled "Ground-Water Budget, Western Snake River Plain, 1980" was not used to compute recharge in the Boise River valley because surface-water return flow could not be differentiated from ground-water discharge. Instead, recharge rates determined by the U.S. Bureau of Reclamation (1964, p. 26) were used. The Bureau estimated that consumptive water use in the Boise River valley averages 2.4 ft/yr. After precipitation during the growing season is subtracted, the crop water requirement that must be supplied by irrigation averages 2.1 ft/yr. The Bureau also estimated that farm delivery requirements for the Boise River valley are about 3.8 ft/yr of water, which allows for a farm efficiency of about 55 percent. On the basis of

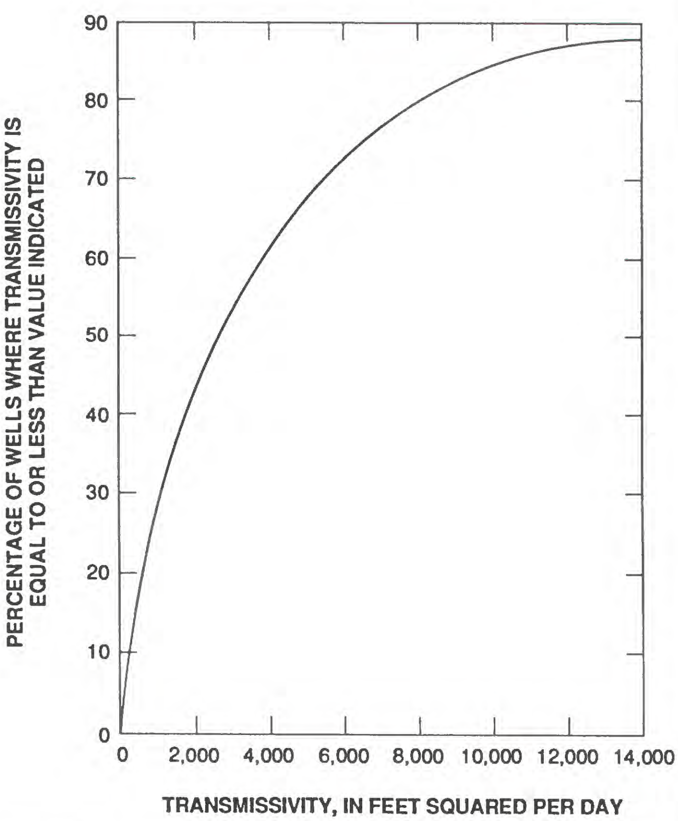


FIGURE 8.—Distribution of transmissivity values computed from specific-capacity data.

TABLE 2.—*Ground-water budget, 1980*

Source	Quantity (acre-feet)
Recharge	
Infiltration from surface-water irrigation	1,400,000
Underflow across boundaries	310,000
Direct precipitation	40,000
Total recharge	1,750,000
Discharge	
Ground-water discharge to rivers and drains	1,450,000
Ground-water pumping	300,000
Total discharge	1,750,000

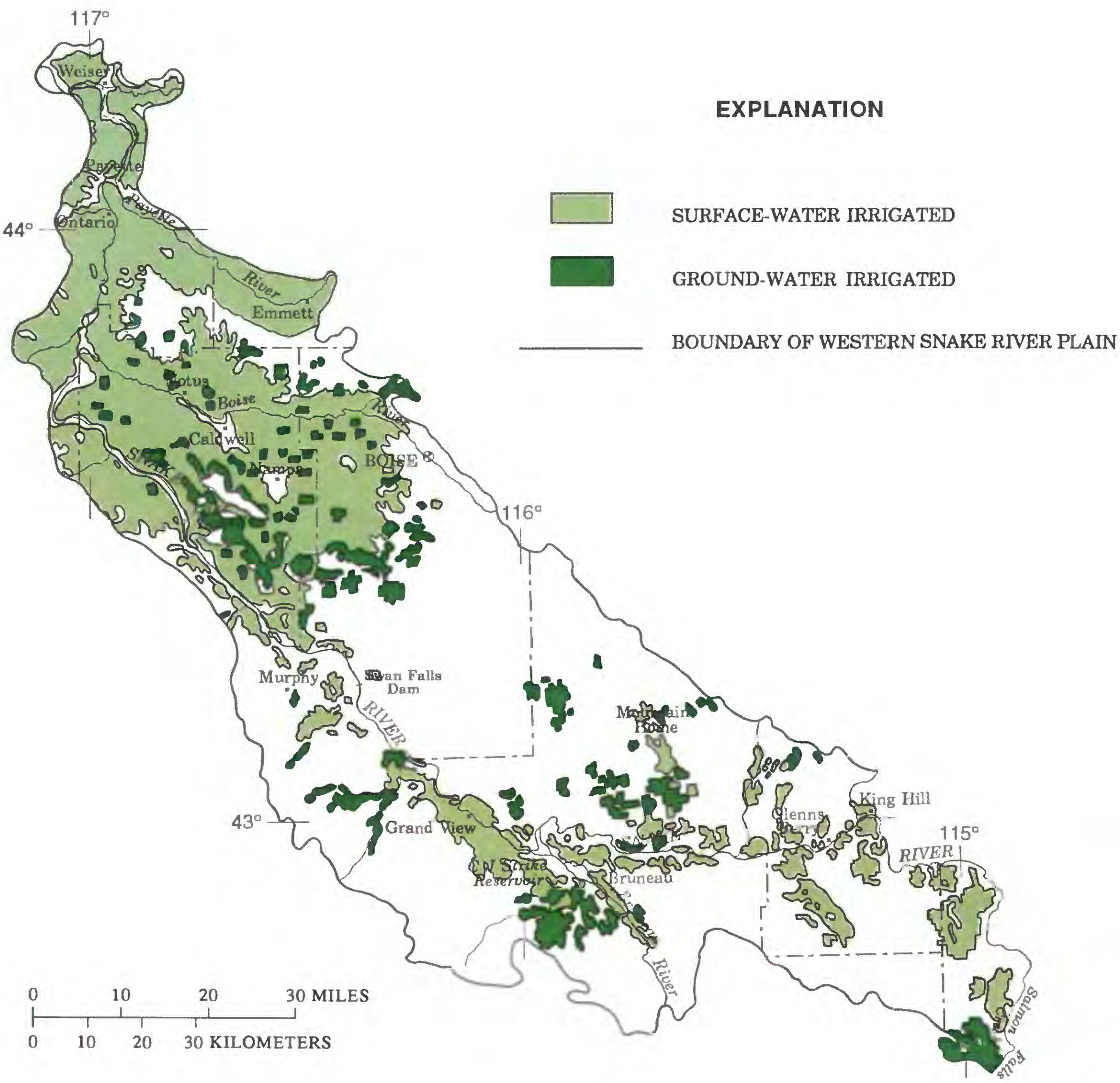
canal and lateral losses, the Bureau estimated that conveyance losses in the Boise River valley average 30 to 33 percent of total diversions. Allowing for this loss, the diversion requirement is about 5.4 to 5.7 ft/yr. Ground-water recharge from irrigation in the Boise River valley is therefore about 3.5 ft/yr.

Not all irrigated areas receive an adequate water supply. Near Mountain Home, farm deliveries are as low as 1.2 ft/yr (U.S. Bureau of Reclamation, 1964, p. 14) and recharge from surface-water irrigation is near zero. Norton and others (1982, p. 45) also showed that crop consumptive-use requirements near Mountain Home exceed available water supply in some years.

Water enters the plain as underflow through fractures and permeable rocks that extend beyond the study-area boundary. Underflow from the area of the Idaho batholith (pl. 1) is difficult to estimate but is assumed to be small. Granitic rocks composing the batholith are nearly impermeable except where highly fractured. Volcanic rocks that extend outside the boundary of the plain are more permeable than the granitic and sedimentary rocks. Almost no data are available to estimate underflow; therefore, it was estimated from the ground-water budget. In the water-budget equation, recharge must equal discharge plus change in storage. Because change in storage was assumed to be small or zero in 1980, underflow was estimated to be about 310,000 acre-ft. If there were a net decrease in storage, underflow would be overestimated; if there were a net increase, underflow would be underestimated.

Precipitation on the western plain was estimated from data supplied by climatologists at the University of Idaho. Average annual precipitation from 1971 to 1980 (fig. 11) ranged from 7 in. near Grand View to 19 in. north of Boise; most precipitation falls during the winter. The average annual volume of precipitation on the western plain from 1971 to 1980 was estimated to be about 2,000,000 acre-ft.

Evapotranspiration from nonirrigated lands in 1980 was estimated to be nearly 1,500,000 acre-ft and, from irrigated lands, nearly 2,000,000 acre-ft. Monthly ET exceeds precipitation except during a few winter months. Where annual precipitation on nonirrigated lands is less than 9 in., recharge from precipitation was assumed to be negligible. In those areas, ET from native vegetation exceeds precipitation. In irrigated areas, soil-moisture conditions are satisfied throughout much of the growing season, and all precipitation is assumed to recharge the ground-water system. On the basis of these assumptions, average recharge from precipitation on the plain was estimated to be about 2 percent of the annual precipitation. The amount of recharge varies, depending on soil-moisture condi-



Modified from unpublished irrigated acreage maps by the U.S. Bureau of Reclamation for the Pacific Northwest River Basins Commission, 1979

FIGURE 9.—Irrigated area, 1980.

tions and the amount of precipitation. Recharge from the 2,000,000 acre-ft of precipitation on the western plain in 1980 was estimated to be about 40,000 acre-ft.

DISCHARGE

Ground-water discharge is largely seepage and spring flow to rivers and drains (canals) and ground-water pumpage. Although ET from shallow ground water may be significant in local areas, ET is negligible relative to the total ground-water budget. Measured surface-water inflows and outflows were used to estimate ground-water discharge to rivers and drains. Estimates were made for winter months, when surface-water diversions for irrigation are few and stream and canal flow is nearly all from ground water. An accurate estimate of Snake River gains is important because nearly half of the total ground-water discharge in the western Snake River Plain is to the Snake River between King Hill and Weiser (table 3). However, accurate estimates are difficult to make because ground-water discharge is small relative to flow in the Snake River. In 1980, ground-water discharge between King Hill and Weiser was only about 5 percent of the total Snake River discharge at Weiser. An estimated 1,450,000 acre-ft of ground water discharged to the Snake, Boise, and Payette Rivers and Salmon Falls Creek in 1980 (table 3). An unknown but probably small amount of ground water was discharged to a few other minor streams.

Ground-water discharge to the Snake River between King Hill and Murphy in 1980 was about 4 percent of the total discharge at Murphy. Only a small part of the area adjacent to that reach is irrigated with surface water (fig. 9) because supplies are generally inadequate. Consequently, irrigation

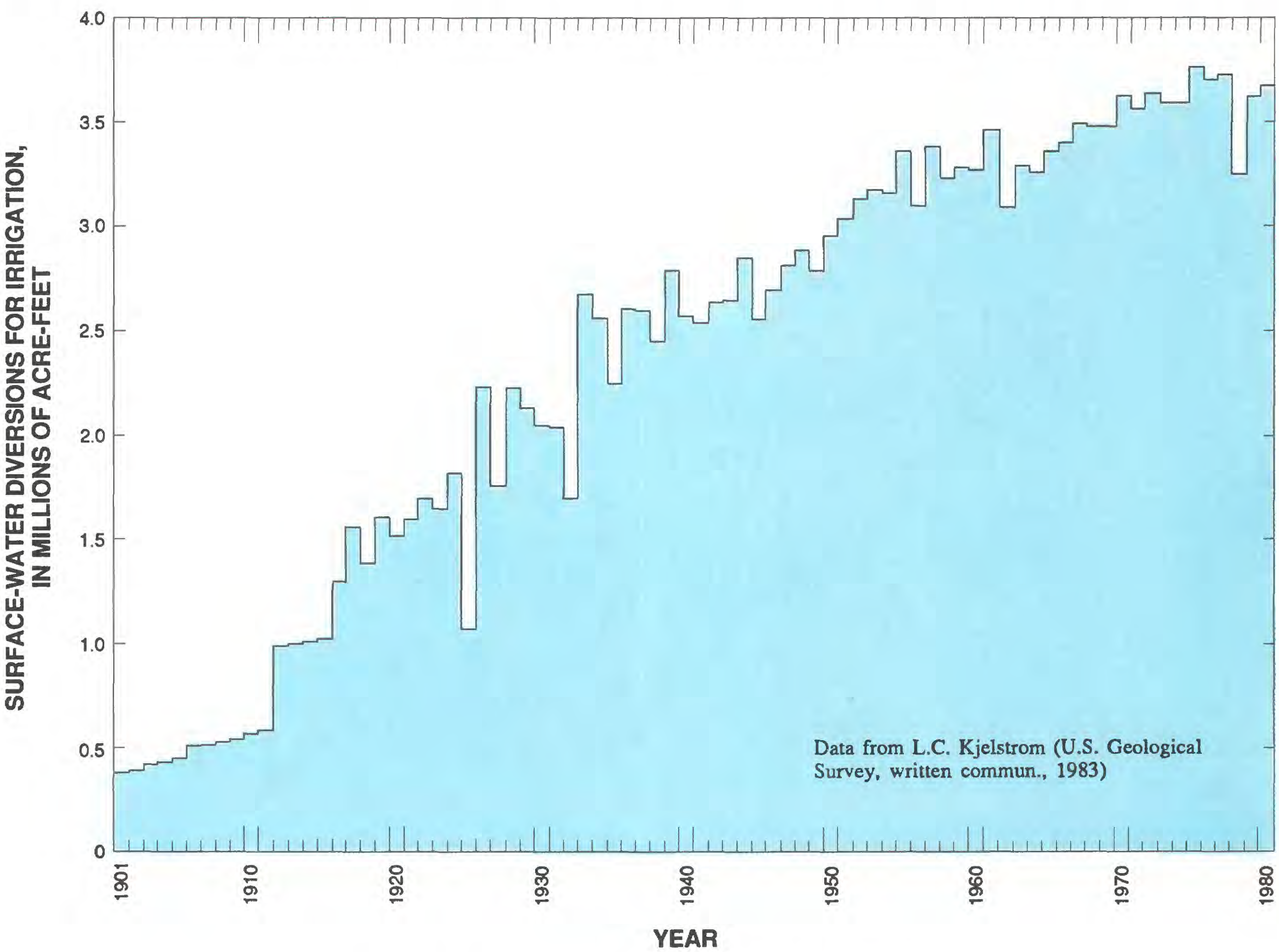


FIGURE 10.—Surface-water diversions for irrigation, 1901–80.

application rates are much lower than in the Boise River valley, and ground-water discharge is not significantly increased by local recharge. Precipitation on the plain between King Hill and Murphy is less than 9 in/yr, and most streams crossing the plain are dry most of the year.

In the Boise River valley, ground water discharges to the river and to an extensive network of

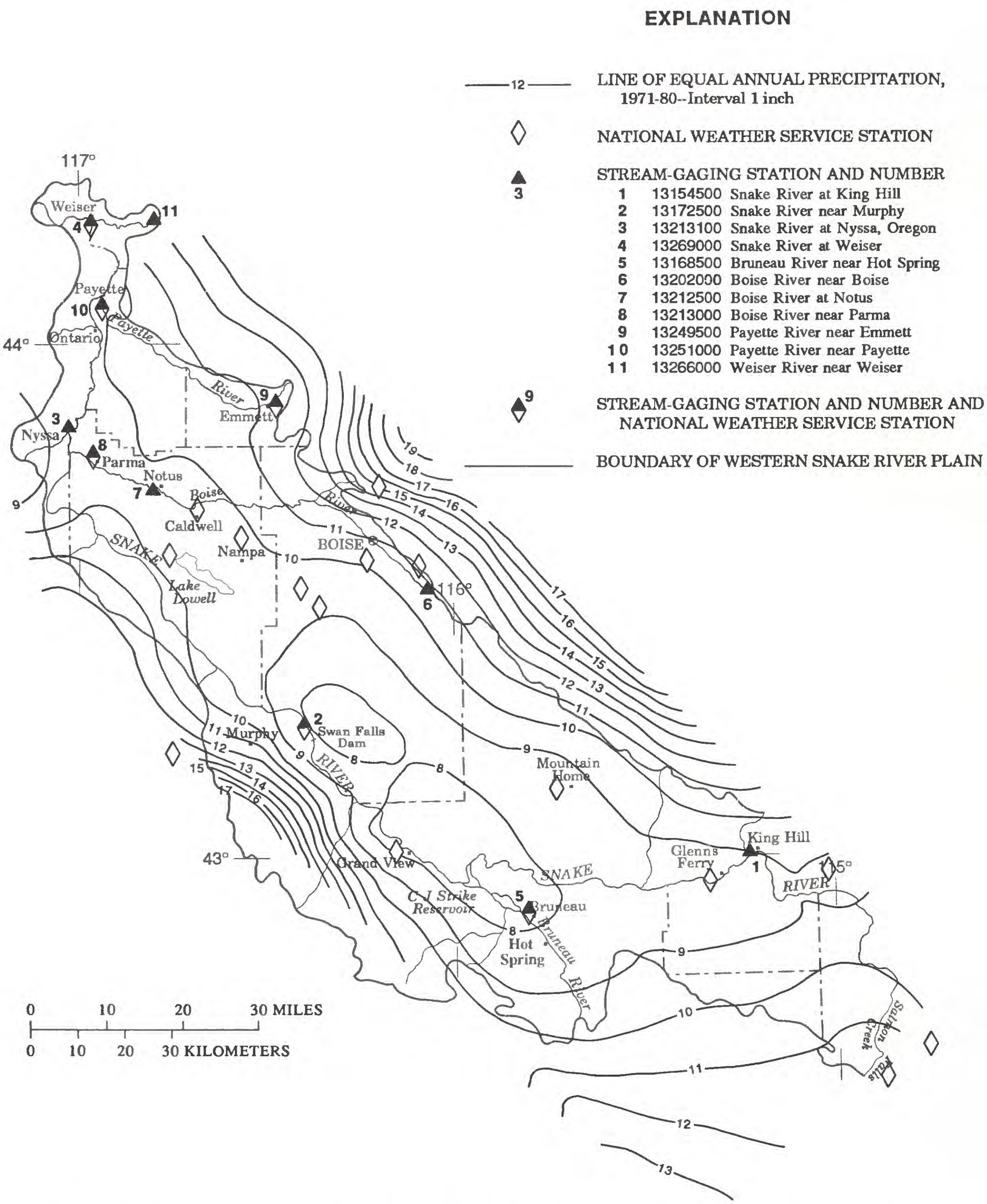


FIGURE 11.—Average annual precipitation, 1971–80, and locations of weather and stream-gaging stations.

TABLE 3.—*Ground-water discharge to major streams, 1980*

Stream	Quantity (acre-feet)
Snake River:	
above King Hill	120,000
King Hill to Murphy	280,000
Murphy to Weiser	350,000
Boise River	460,000
Payette River	220,000
Salmon Falls Creek	20,000
Total	1,450,000

canals and drains. Discharge to the river varies greatly from summer to winter but averages about 800 ft^3/s during winter (Thomas and Dion, 1974, p. 40); during the summer, discharge is increased by irrigation return flow.

Thomas and Dion (1974, p. 40) showed that the increase in Boise River flow between gaging stations near Boise and at Notus (fig. 11) results from ground-water discharge to drains and to the Boise River. Ground-water discharge to drains in the Boise River valley was estimated from Boise River discharge records. From October 1 of one year to March 31 of the next year, river gains from irrigation return flows average about 455,000 acre-ft. During winter months, when surface-water diversions are few, the increase in flow is nearly all from ground water. Ground-water discharge peaks at about 1,000 ft^3/s in September, when ground-water levels are highest, then decreases until about April, just before irrigation diversions begin. Ground-water discharge to canals and drains during the irrigation season is difficult to measure because irrigation return flows cannot be differentiated from ground-water discharge.

Ground-water discharge to the Payette River between Emmett and Payette (fig. 11) was estimated by subtracting flow of the Payette River near Emmett and inflow from Big and Little Willow Creeks from flow in the Payette River near Payette. Discharge from several other small tributaries was not measured. Estimated ground-water discharge to this reach from November 1, 1979, to January 31, 1980, was 217,000 acre-ft. Inflow from other small streams would reduce the computed gain.

Ground-water discharge to Salmon Falls Creek is mostly from outside the study area. No water is released into Salmon Falls Creek from instream reservoirs, except for possible leakage around dams. Therefore, any gains in the reach below the reservoir are largely from ground water and irrigation return flow. During winter months, when no water is diverted for irrigation, essentially all streamflow is from ground water. From November 1, 1979, to January 31, 1980, Salmon Falls Creek gained about 130,500 acre-ft from the dam to its mouth. Only a short part of the reach is within the plain; therefore, total ground-water discharge was assumed to be about 20,000 acre-ft.

Little ground water was pumped on the western Snake River Plain in the early 1900's. However, since the late 1940's, the number of irrigation pumps has increased steadily (fig. 12). Data shown in figure 12 are for the Idaho Power Company service area, which includes all of the western plain and about half of the eastern plain. The data

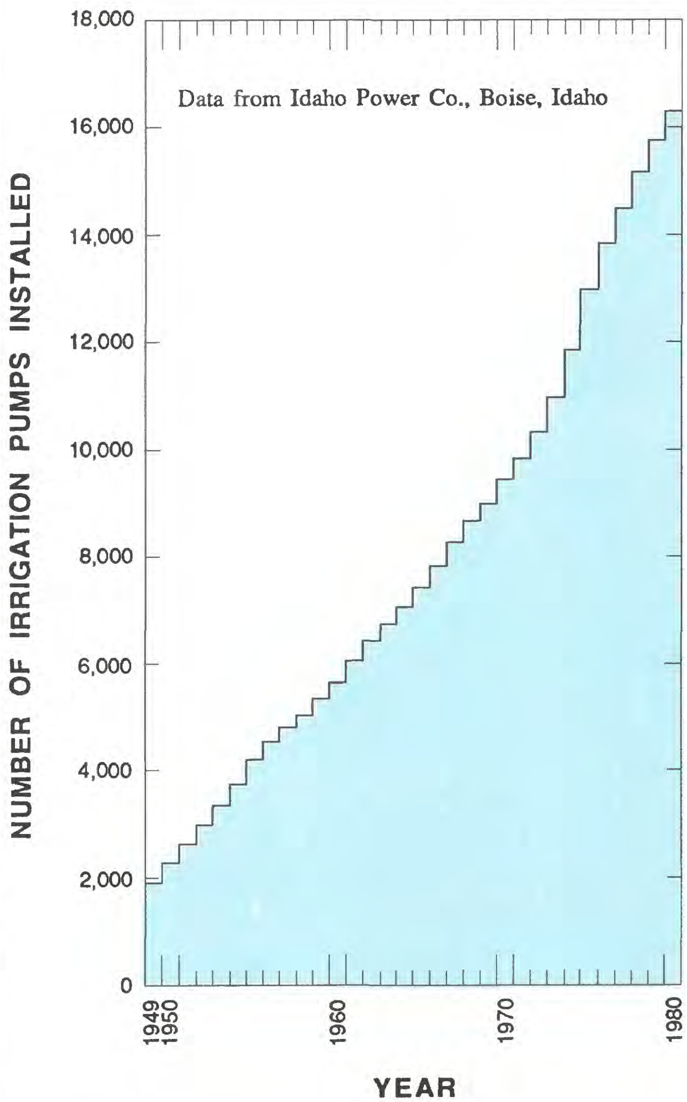


FIGURE 12.—Number of irrigation pumps installed in area served by Idaho Power Company, 1949–80.

include some relift pumps on canals and streams. In the Boise River valley, where a high water table causes waterlogging, some pumps included in figure 12 are actually used in dewatering wells. Much of this pumped water is used for irrigation downstream.

Piper (1924, p. 46) estimated that total discharge from flowing wells in the Bruneau–Grand View area was about 7,100 acre-ft in 1922. Littleton and Crosthwaite (1957, p. 51) estimated pumpage in the same area was 22,300 acre-ft in 1954, and Young and others (1979, p. 5) estimated pumpage was 50,000 acre-ft in 1978.

Power-consumption records were used to estimate that about 300,000 acre-ft of ground water were pumped to irrigate about 130,000 acres on the western Snake River Plain in 1980 (Bigelow and others, 1986). The mean application rate was about 2.2 acre-ft/acre. Irrigation well locations determined from power-company records are shown in figure 13.

In the Boise River valley, ground-water discharge as ET could be significant. Where the water table is at or near land surface, the potential exists for ET of ground water. In several areas, alkaline soils have developed as a result of deposition of salts left by ET of ground water.

In the Bruneau–Grand View area, ground water is discharged naturally by springs and ET. The amount of water discharged by ET is not known

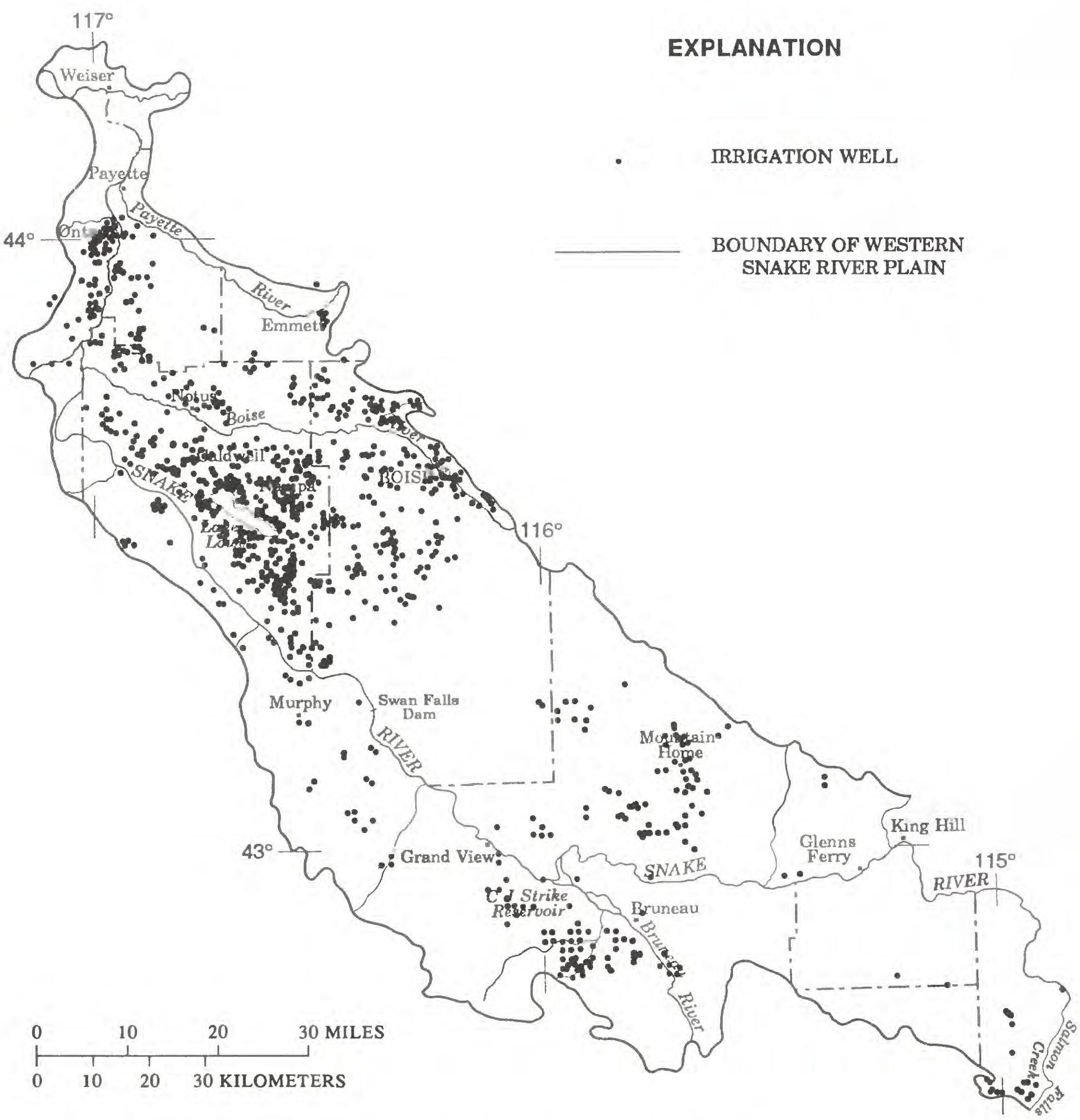


FIGURE 13.—Locations of pumped irrigation wells, 1980.

but probably is substantial because extensive nonirrigated lowlands include areas of alkaline and saline soils (Littleton and Crosthwaite, 1957, p. 173).

SIMULATION OF GROUND-WATER FLOW

A finite-difference computer program (McDonald and Harbaugh, 1984) was used to simulate the ground-water-flow system underlying the western Snake River Plain. The three-dimensional model enabled simulation of vertical flow between major rock units. Because the model grossly simplifies a complex hydrologic system, results of the model should be used with caution.

Response of the ground-water system in the western plain to hydraulic stress, such as pumping, depends on (1) proximity and nature of the geologic and hydrologic boundaries, (2) aquifer properties (transmissivity and storage capacity), and (3) distribution of recharge to and discharge from the system. Values of the following hydraulic properties were defined and incorporated into the simulation model: (1) boundary conditions, (2) transmissivity, (3) storage coefficient or specific yield, (4) recharge and discharge rates, (5) thickness and vertical hydraulic conductivity of confining layers, and (6) head distributions.

MODEL DESCRIPTION

The boundary of the modeled area approximates the boundary of the western Snake River Plain from the junction of the Payette River with the Snake River southeast to the junction of Salmon Falls Creek with the Snake River (fig. 14).

The ground-water system was divided into a network of cells arranged in 25 rows, 72 columns, and 3 layers. Each cell is 2 mi on a side and represents a land area of 2,560 acres. The grid origin (row 25, column 1) is at the intersection of 43°43'14" latitude and 117°23'44" longitude. The grid is oriented 45° clockwise from the north to minimize the number of inactive cells in the model and to align the grid with the principal direction of ground-water flow (shown by arrows in fig. 3).

The three model layers represent the upper, middle, and lower rock units, as previously defined (fig. 15). Model layer 1 represents about 500 ft of sedimentary and volcanic rocks in the upper unit. The top of layer 1 was defined on the basis of water levels measured in the spring of 1980. The bottom of layer 1 was assumed to be 500 ft below the depth-to-water contours (fig. 3). Model layer 2 represents about 4,000 ft of predominantly fine-grained sedimentary rocks and volcanic rocks in the middle unit. Water is confined in sand lenses by sediments of low vertical hydraulic conductivity. Model layer 3 represents about 7,000 ft of volcanic rocks in the lower unit. Water is confined under high pressure and, along the northeastern and southwestern margins of the plain, wells yield geothermal water.

The modeled area was divided into 11 geohydrologic subareas on the basis of geologic and hydrologic characteristics of the aquifer system (fig. 16). The same subareas were used for all layers. In this manner, a complex aquifer system was simplified for simulation without great deviation from reality.

BOUNDARY CONDITIONS

Model boundaries were simulated as constant flux, no flow, or head dependent (figs. 16, 17).

The rate of flow assigned to a constant-flux cell remains constant throughout steady-state simulations; however, the head may change. The rate of flow assigned to a constant-flux cell may be changed for each time period during a transient simulation. Head-dependent boundaries are used to simulate flow from or into external sources such as rivers and drains and internal boundaries between model layers. The flow across head-dependent boundaries is the product of the head difference across the boundary and the conductance of the boundary. The equation for determining hydraulic conductance is

$$C = K(A/L),$$

where

C = hydraulic conductance,
K = hydraulic conductivity,
A = cross-sectional area perpendicular to the ground-water flowpath, and
L = distance along the flowpath.

The ground-water flowpath, based on water-level data, indicates some underflow into the modeled area from the surrounding mountains and adjacent areas. Constant-flux boundaries were used on the north, south, and west sides to simulate such underflow from outside the modeled area. Underflow was simulated by recharge wells placed in appropriate boundary cells. Almost no data were available on the amount of underflow; therefore, the amount was estimated from the ground-water budget.

The eastern boundary, formed by the Snake River, was simulated as head dependent in layer 1 and no flow in layers 2 and 3 because scant hydrologic evidence indicates little or no underflow between the eastern and western plain in layers 2 and 3.

The Snake River, Payette River, Salmon Falls Creek, and Lake Lowell were simulated as head dependent. Almost all ground-water discharge is to rivers. Groups of river cells were combined to correspond with reaches where ground-water discharge was measured or estimated.

Rivers may contribute water to or receive water from the ground-water system, depending on whether ground-water levels are above or below river stage. A river loses water when aquifer head

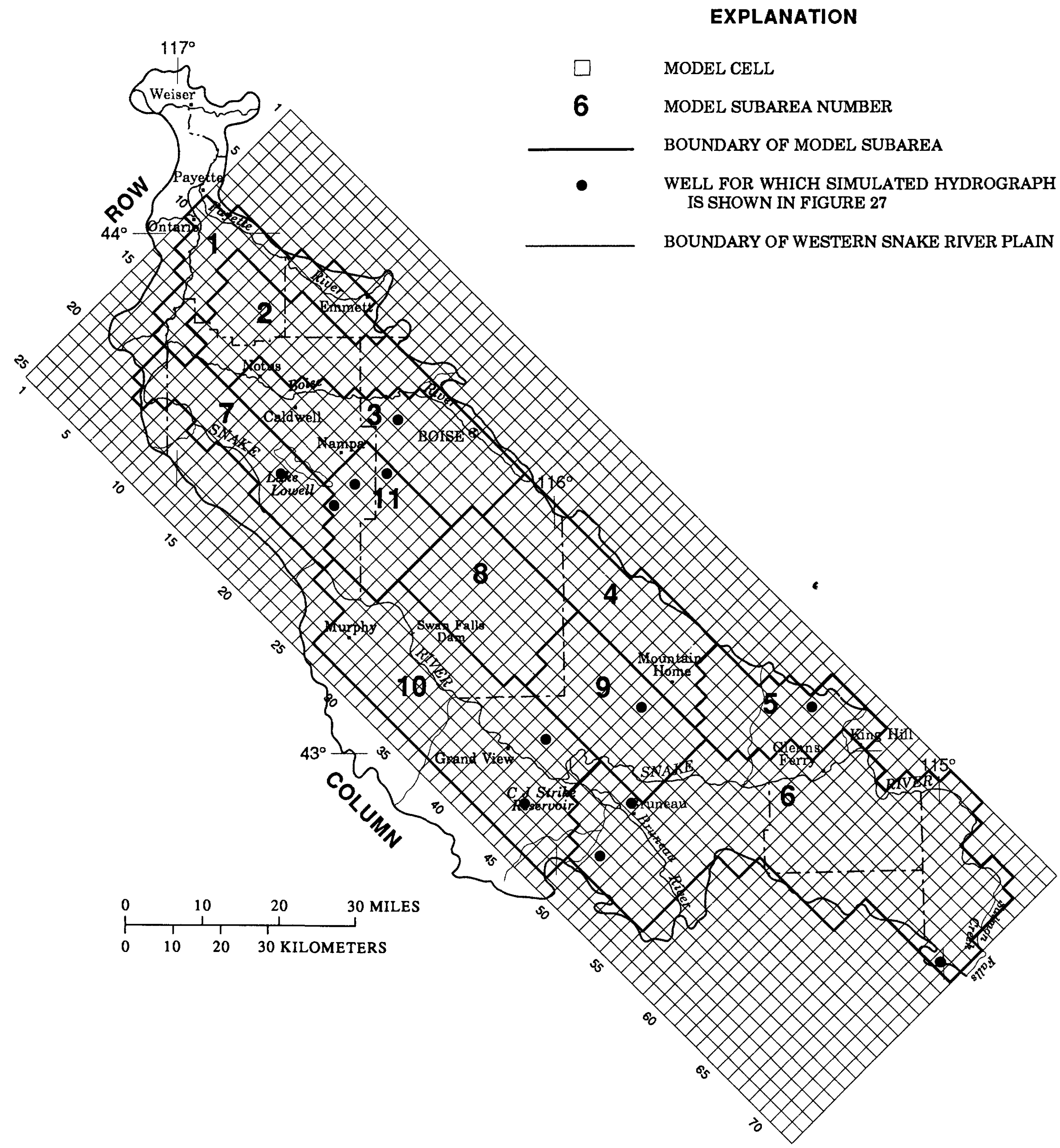


FIGURE 14.—Finite-difference grid and model subareas.

is below river stage and gains water when aquifer head is above river stage. To simulate the river-aquifer connection, a conductance value was assigned to each river cell. Conductance is calculated by dividing vertical hydraulic conductivity of the riverbed by riverbed thickness and multiplying by the area of the river within the cell. Field data on riverbed hydraulic conductivity and thickness gen-

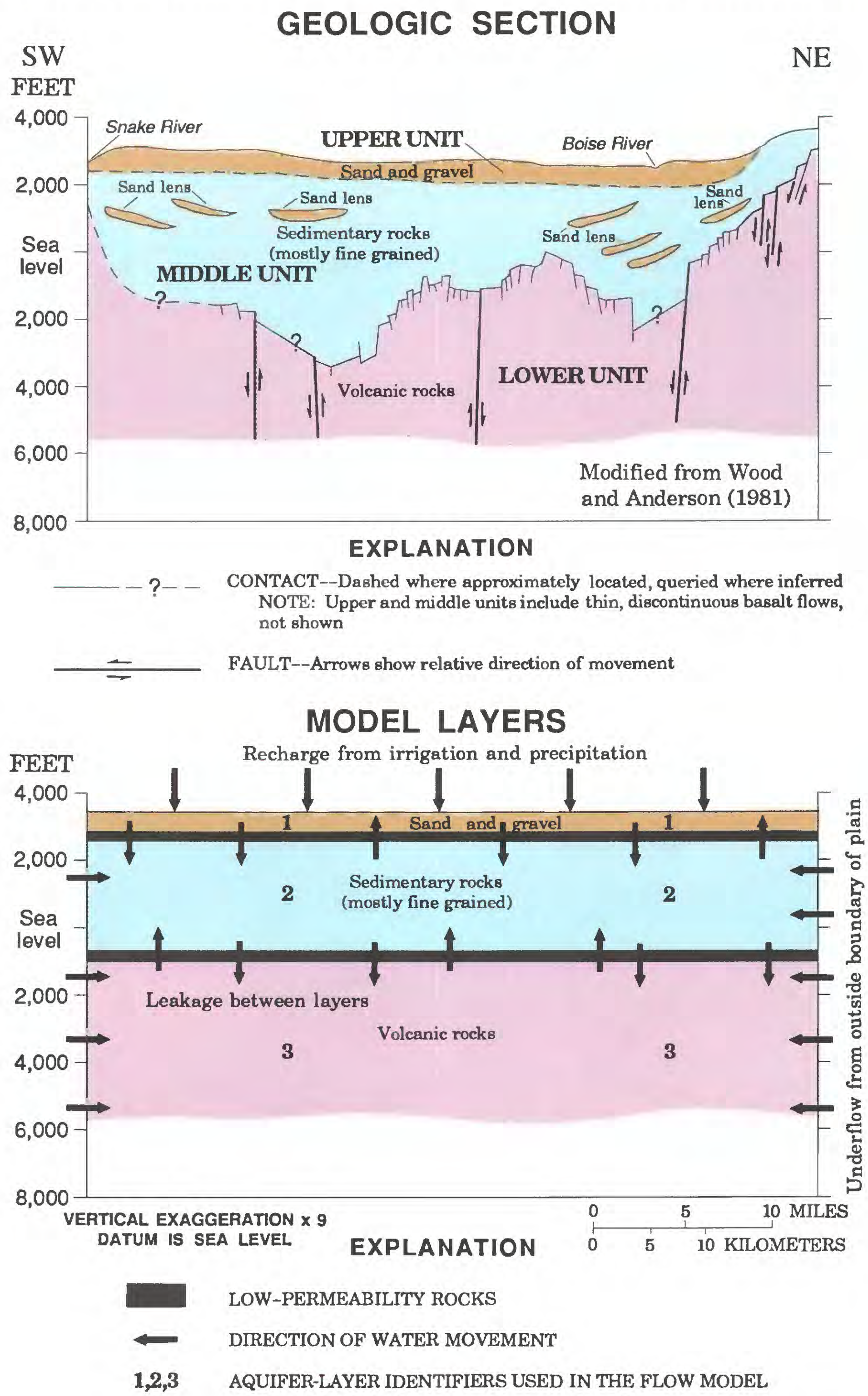


FIGURE 15.—Generalized geologic section and model layers.

erally were unavailable. Assuming an average river width of 1,300 ft, an average length of 2 mi per cell, an average riverbed thickness of 10 ft, and a hydraulic conductivity for a gravel riverbed of 1×10^{-5} ft/s, riverbed conductance would be about 14 ft^2/s.

Parts of the area between the Boise River and the Snake River and part of the Payette River valley were simulated with drains in layer 1 (fig. 16). The bottom of the drain for each cell was approximated by averaging land-surface altitude within a cell. If both drains and the Boise River were in the same cell, the cell was assigned as a drain. During the simulated time period, the Boise River was always a gaining stream.

Canals in the Boise River valley are related to the ground-water system in much the same way as rivers. Average length of major canals in each

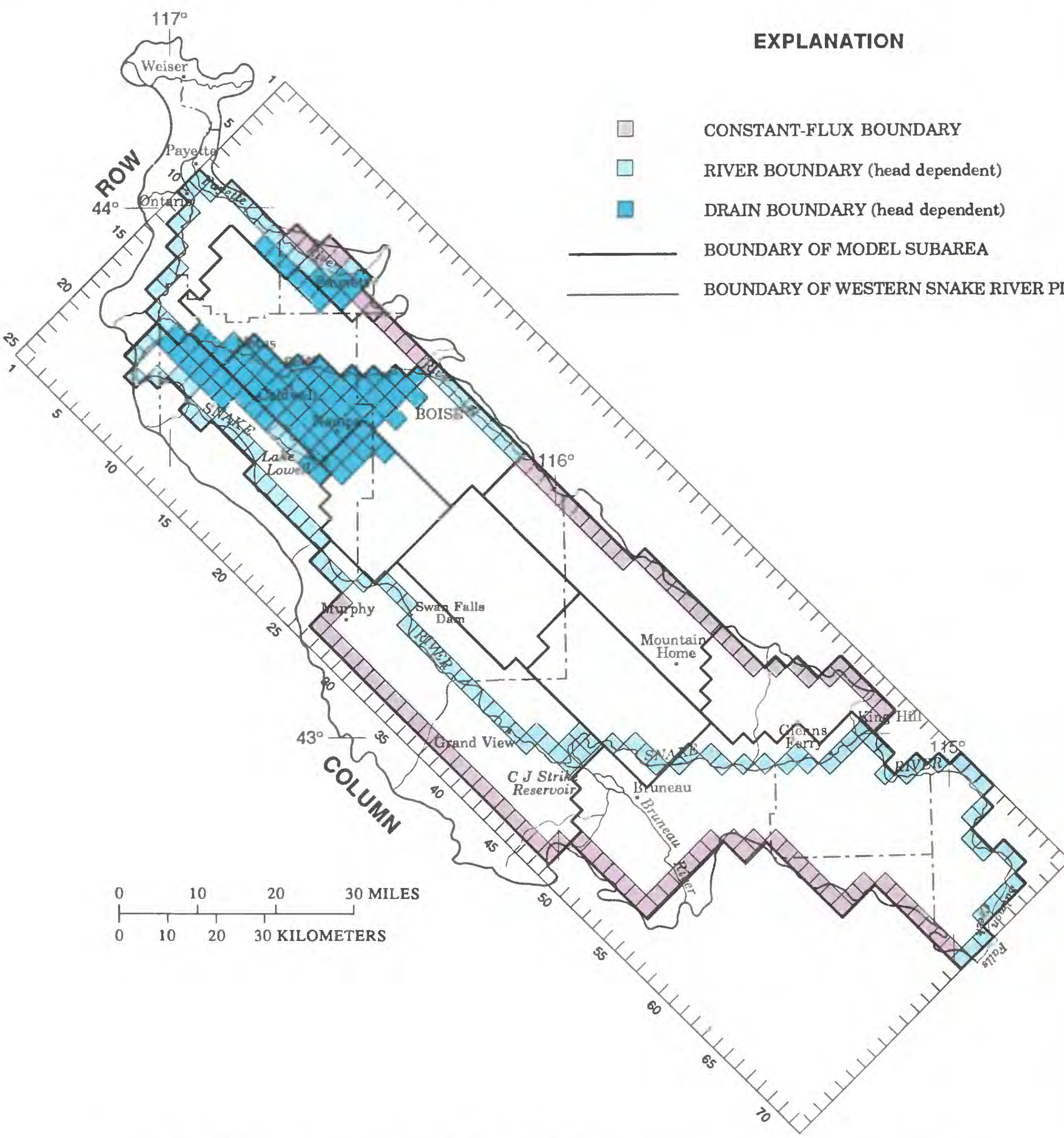


FIGURE 16.—Boundary conditions, rivers, and drains, model layer 1. (Model layers shown in fig. 15.)

model cell was estimated from topographic maps to be about 9 mi, and average canal width was assumed to be about 20 ft. Total canal area is about 1.3 percent of the irrigated area in a cell, or about 333 acres per cell in the Boise River valley. Assuming a canal-bed thickness of 1 ft and a hydraulic conductivity of 1×10^{-6} ft/s, canal-bottom conductance would be about 14 ft^2/s.

Boundaries between layers were simulated as head dependent. Individual confining layers were not simulated because they are poorly defined. Rather, effective vertical-conductance values between model layers were estimated. The conductance values represent the average conductance between adjacent layers.

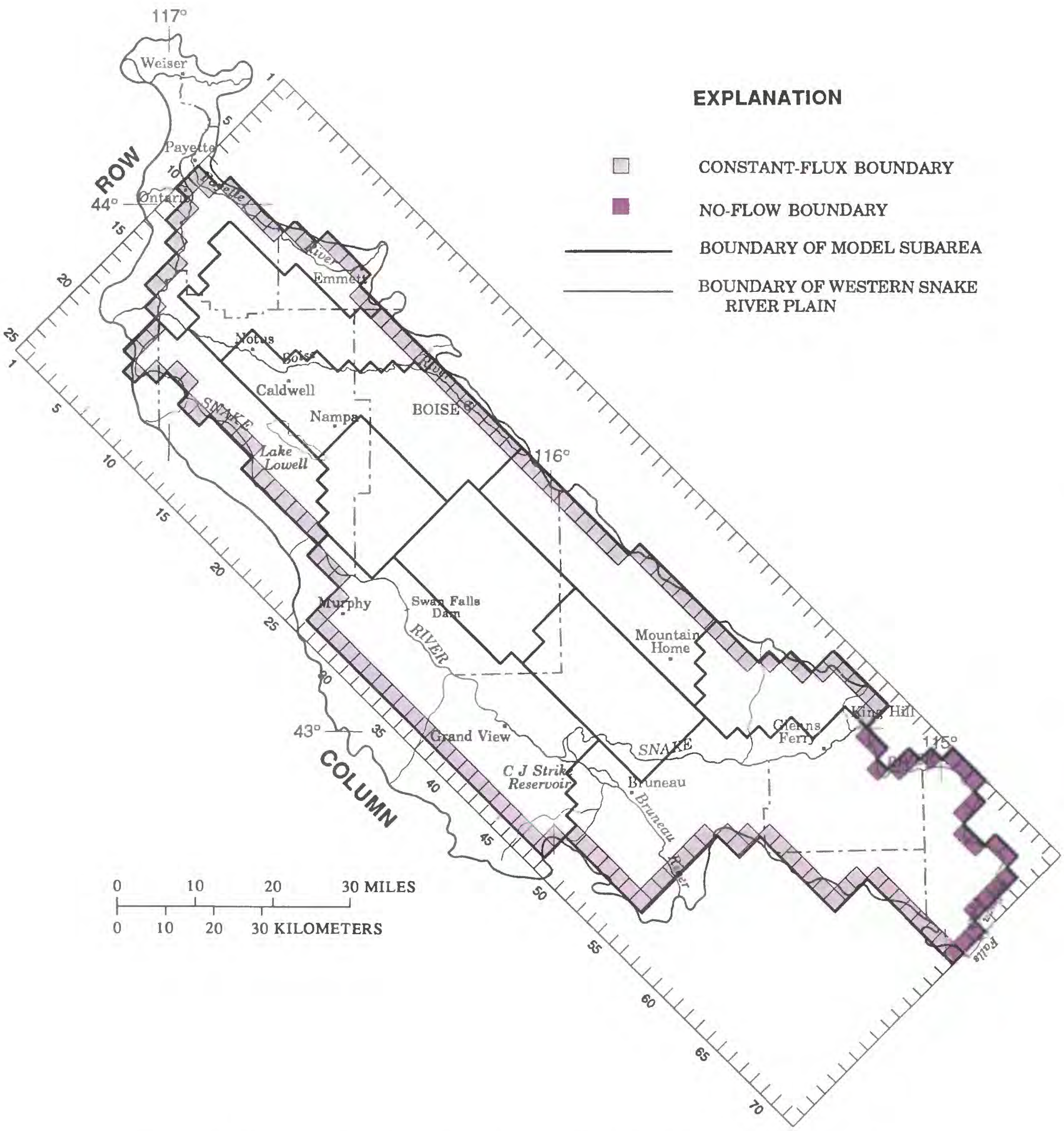


FIGURE 17.—Boundary conditions, model layers 2 and 3. (Model layers shown in fig. 15.)

MODEL DEVELOPMENT AND CALIBRATION

The model was developed and calibrated in two stages. Stage 1 was a simulation of 1980 steady-state conditions. Calibration consisted of adjusting transmissivity, river and drain conductance, and vertical hydraulic conductivity within reasonable ranges. Under steady-state conditions, ground-water storage does not change. Stage 2 was a transient analysis of simulated aquifer response to measured pumping and recharge stresses, with the objective of reproducing long-term water-level changes from 1880 (preirrigation) to 1980. Calibration in stage 2 consisted of adjusting storage coefficients until measured water-level changes were simulated reasonably.

Because few field data were available on hydraulic properties used in the model, a range of values was more appropriate than a single number. The model can be used to bracket assumed best values of transmissivity, storage coefficient, and other properties. A sensitivity analysis was performed to show the effects of using a range of values on simulation results.

STEADY-STATE ANALYSIS

Objectives of steady-state analysis were to (1) calibrate the model to a selected equilibrium condition in which the stress and response characteristics of the system are well documented, and (2) provide the initial conditions for subsequent transient simulations.

No information on preirrigation water levels was available for steady-state calibration; therefore, 1980 hydrologic conditions were chosen as an alternative because data on precipitation, pumpage, streamflow, estimates of recharge, and water levels were available. After years of agricultural development in the study area, use of water for irrigation reached a stable condition. The aquifer system also adjusted to agricultural development and reached an approximate steady-state condition in most of the study area, as indicated by small water-level fluctuations measured in the 1970's.

The calibrated steady-state model was used to simulate preirrigation water levels by assuming there was no ground-water pumpage or recharge from irrigation water, with all other input remaining the same. The simulated preirrigation water levels were used as initial conditions for transient simulations for the period 1881–1980.

During calibration, hydraulic-conductivity, transmissivity, and vertical-conductance values were adjusted to obtain the best possible match between model-simulated water levels and spring 1980 measured water levels. Model-simulated river gains from ground water were compared with estimated gains. Simulated values of riverbed and canal-bed conductance were adjusted to maintain estimated 1980 ground-water discharge to rivers and drains.

MODEL INPUT

Hydraulic-conductivity values estimated from specific-capacity data were assigned to each model subarea for layer 1. A uniform transmissivity of 8,600 ft^2/d (0.10 ft^2) was assigned to layers 2 and 3.

No data for vertical hydraulic conductivity were available; however, thick clay layers are common throughout layer 2 and greatly restrict vertical flow. Vertical flow rates through volcanic rocks in layer 3 are probably higher. The value of vertical hydraulic conductivity between layers 1 and 2 was assumed to be about 9 ft/d and, between layers 2 and 3, about 22 ft/d, on the basis of published values for similar rock types (Walton, 1984, p. 23).

The model requires vertical-leakance values between layers as input. Vertical leakance was calculated by dividing the assumed vertical hydraulic conductivity by the average assumed thickness between layers. The leakance between layers 1 and 2 and between layers 2 and 3 was computed to be about 4×10^{-3} 1/d.

Model cells that include surface-water-irrigated areas in 1980 are shown in figure 18; cells that include ground-water-irrigated areas are shown in figure 19. If both ground water and surface water are used for irrigation in the same area, ground water is usually a supplemental supply. Irrigated acreage in each model subarea is given in table 4.

Estimated ground-water recharge from surface-water irrigation in each model subarea in 1980 is given in table 5. Estimated recharge from precipitation in each model subarea in 1980 is given in table 6. Ground-water pumping was simulated using discharge wells. Pumpage from all wells within each cell was simulated by a single well in the center of each cell. Ground-water pumpage in excess of consumptive use was assumed to be returned directly to the aquifer. The estimated pumpage used in model simulation is net pumpage, which is approximately equal to estimated consumptive use.

Modeled ground-water pumpage and application rates in each model subarea in 1980 are given in table 7. Underflow was simulated using wells, as described in the section "Boundary Conditions."

RESULTS

The model was calibrated so that simulated head values and ground-water discharge to rivers and drains agreed reasonably well with measured and estimated values.

Water levels measured in 305 wells in the spring of 1980 were compared with simulated head values for layer 1 (fig. 20). Simulated head values were interpolated at the well locations for comparison purposes. The simulated 1980 steady-state potentiometric surface in model layer 1 is shown in figure 21. Comparison of figure 21 with figure 3 (1980 water-table map based on field measurements) indicates that, although the general configuration of the contours is similar, the simulated surface has smoother, less distorted contours. The smoother contours reflect the inability of the model to simulate complexities of local geology and changes in hydraulic properties of the rocks. Differences between measured and simulated heads were least in the Boise River valley and greatest along the boundaries of the plain, where geology is more complex.

Values of hydraulic conductivity for model layer 1, used in the model to generate figure 21, were compared with hydraulic conductivities estimated from specific-capacity data (table 8). Overall, simulated hydraulic-conductivity values are reasonably consistent with hydraulic conductivities computed from specific-capacity data. The comparison is poorest for subarea 4, where hydraulic conductivity estimated from specific-capacity data exceeds the simulated estimate by a factor of 10. The comparison is best for subarea 2 (Boise River valley), where both estimates are nearly equal. The apparent error in subarea 4 may be due to inadequate simulation of this poorly defined area and to errors caused by using specific-capacity data to compute hydraulic conductivity.

Simulated 1980 steady-state potentiometric surfaces for model layers 2 and 3 are shown in figures 22 and 23. The validity of the simulated heads in layers 2 and 3 cannot be judged because almost no wells are completed in these layers, except for a few in the Mountain Home and Bruneau–Grand View areas. Even these measured water levels are a composite of heads in several aquifers. Potentiometric surfaces shown in figures 22 and 23 are largely hypothetical, but they may provide information about water movement in these rock units.

Simulated preirrigation potentiometric surfaces for each model layer are shown in figures 24–26. The configurations of these potentiometric surfaces are similar to those shown in figures 21–23 for 1980 steady-state conditions.

The calibrated model reasonably simulated ground-water discharge to rivers and drains. Simulated and estimated river-reach gains in 1980 are listed in table 9.

Assigned transmissivity values for model layers 2 and 3 are summarized in table 10. These values were not adjusted during calibration because insufficient head data precluded verification of initial transmissivity estimates. However, changes in transmissivity in subareas 5, 6, 8, 9, and 10 for layer 2 effected large head changes in layer 1. Transmissivity in subarea 6 was increased to 12,100 ft^2/d to lower heads in layer 1 and to account for the presence of basalt in this area. Transmissivity values in subareas 5, 8, 9, and 10 were reduced to raise heads in layer 1 and to account for the presence of clay and shale.

Assigned vertical hydraulic conductivity between model layers is shown in table 11. Vertical hydraulic conductivity is highest in subareas 8 and 9, where the principal rock unit is basalt, which may be vertically fractured and faulted. Although assigned values for subareas 8 and 9 are one to two orders of magnitude higher than for other subareas, these high values seemed to be justified on the basis of model-simulated heads.

TRANSIENT ANALYSIS

The objectives of transient calibration were to (1) verify data used in the steady-state calibration, (2) evaluate specific yield and storage coefficients, and (3) evaluate historical changes in hydrologic conditions and their effects on the ground-water-flow system.

Transient conditions were simulated for the period 1881 to 1980 to test specific-yield and storage-coefficient values that are not involved in steady-state simulation.

Calibrated steady-state hydraulic conductivity, transmissivity, and leakance were used in the transient simulation. Ground-water pumpage was distributed over time on the basis of Idaho Power Company records of the number of irrigation pumps added each year since 1949 (fig. 12).

Prior to 1949, little ground water was pumped for irrigation because surface water was readily available.

Recharge from irrigation in the Boise River valley from 1880 to 1980 was estimated on the basis of watermaster records of irrigation diversions. Recharge in other areas was estimated on the basis of diversion records obtained from various governmental agencies. Diversions and recharge for each year were assumed to be in the same proportion as they were in 1980.

MODEL INPUT

Transient simulation requires values of specific yield and storage coefficient. Although neither of these values is accurately known for the western

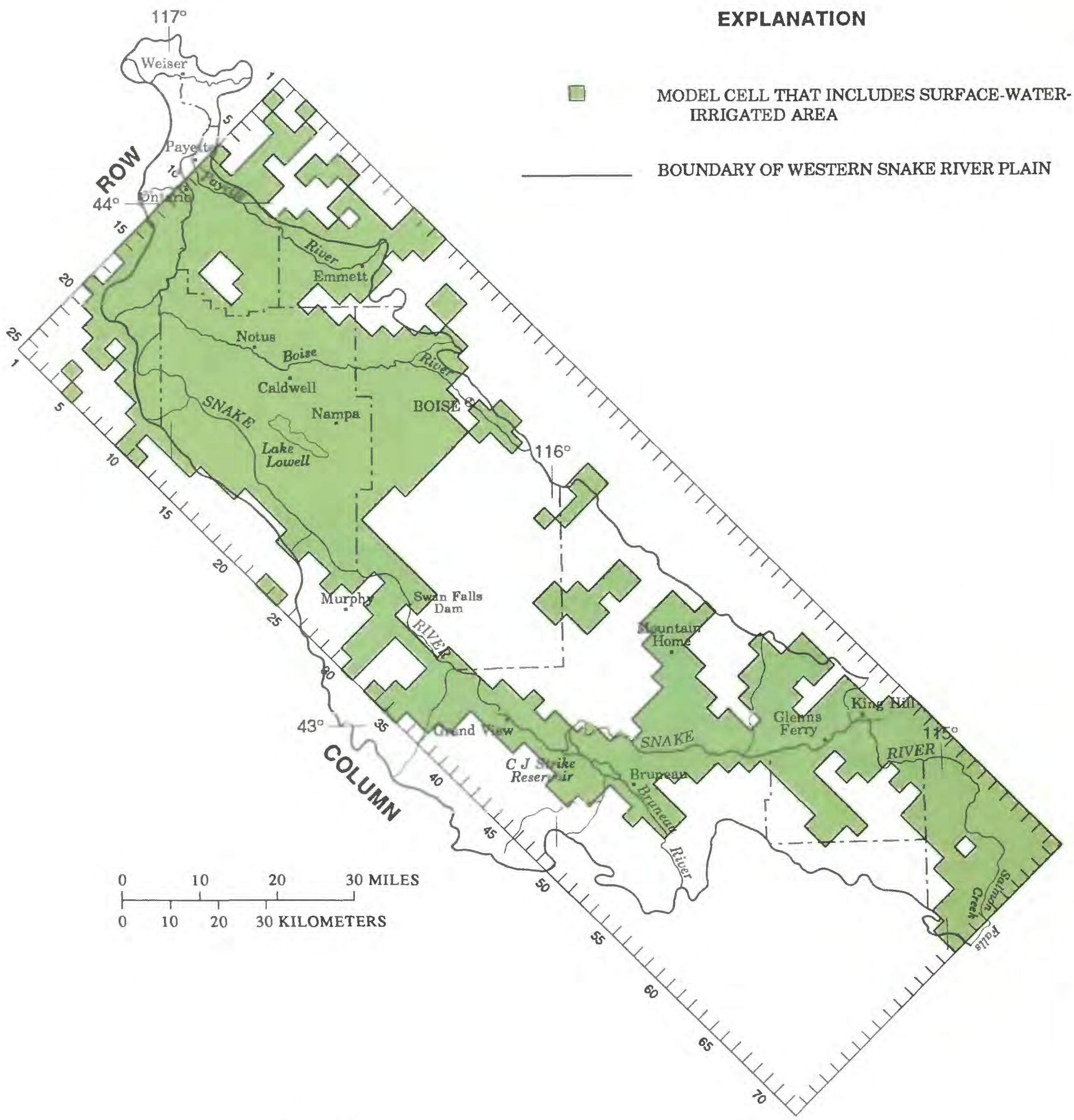


FIGURE 18.—Model cells that include surface-water-irrigated areas, 1980.

Snake River Plain, the specific yield was estimated to range from about 0.025 to 0.43 and the storage coefficient from 0.00007 to 0.006 (see the section "Hydraulic Properties").

A uniform specific yield of 10 percent was assigned to model layer 1. Inadequate information made it impractical to vary specific yield areally. Long-term water-level records were available only for the Boise River valley; therefore, more emphasis was placed on simulating water levels in that area.

Storage coefficients of layers 2 and 3 were assumed to be 4×10^{-3} and 7×10^{-3}, respectively. The values were computed by multiplying aquifer thickness times an assumed specific storage value of 1×10^{-6} (Walton, 1984, p. 22). Because information on storage coefficients for layers 2 and 3 was lack-

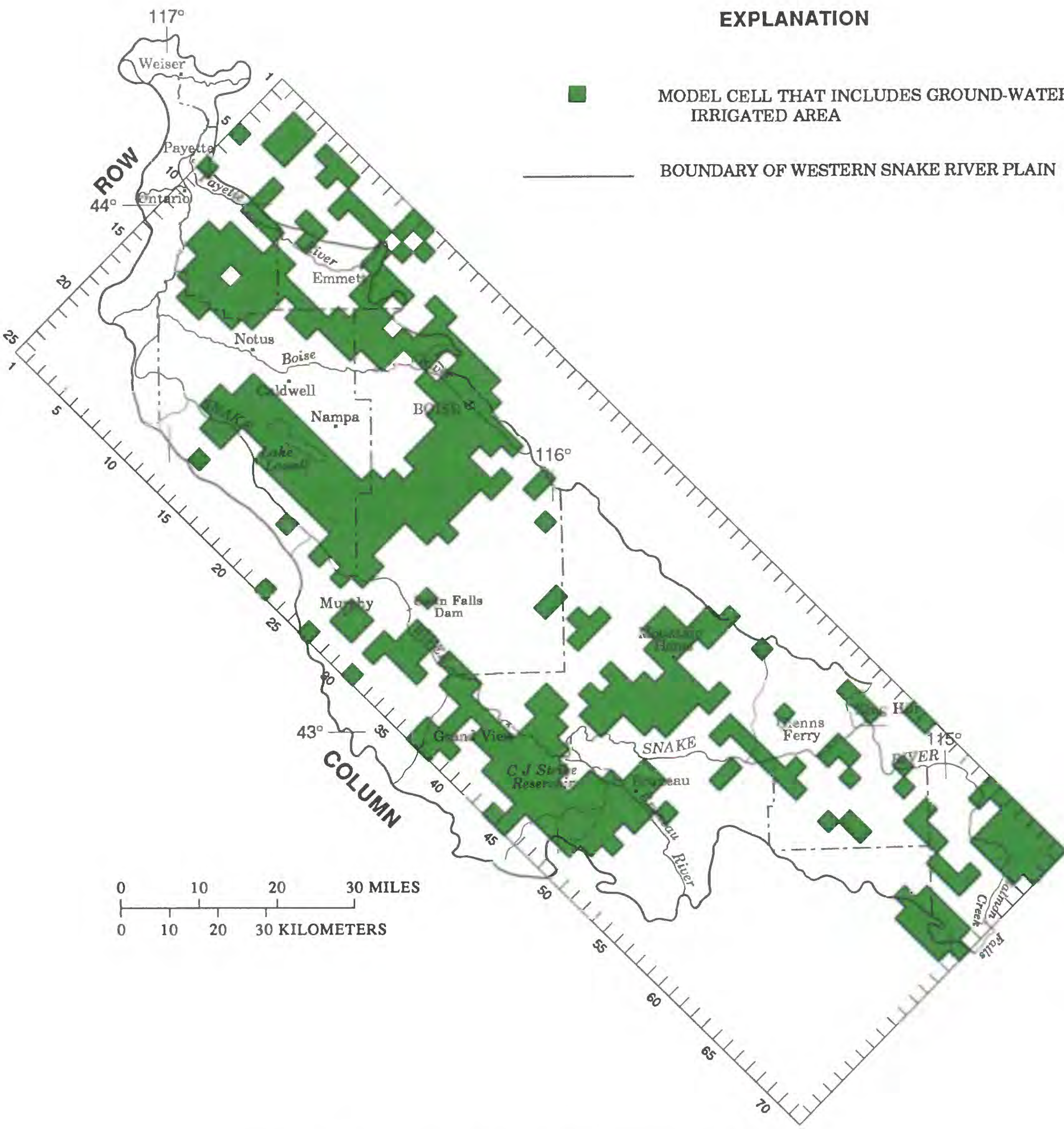


FIGURE 19.—Model cells that include ground-water-irrigated areas, 1980.

TABLE 4.—*Irrigated acreage in each model subarea, 1980*

[Values are in acres]

Model subarea (see fig. 14)	Irrigated with ground water	Irrigated with surface water	Total
1	3,000	106,000	109,000
2	10,800	87,600	98,400
3	13,700	126,500	140,200
4	2,800	8,700	11,500
5	1,400	10,400	11,800
6	25,600	124,600	150,200
7	14,300	103,500	117,800
8	2,900	100	3,000
9	12,400	20,500	32,900
10	23,800	48,400	72,200
11	18,800	60,200	79,000
Total	129,500	696,500	826,000

TABLE 5.—*Estimated ground-water recharge from surface-water irrigation in each model subarea, 1980*

Model subarea (see fig. 14)	Estimated recharge (acre-feet)	Irrigated area (acres)
1	210,000	106,000
2	180,000	87,600
3	320,000	126,500
4	10,000	8,700
5	10,000	10,400
6	190,000	124,600
7	260,000	103,500
8	0	100
9	30,000	20,500
10	70,000	48,400
11	120,000	60,200
Total	1,400,000	696,500

ing, these values were not changed during the calibration process. The effect of changing storage values on simulated water levels is described in the section "Sensitivity Analysis."

RESULTS

Transient simulation indicated that ground-water levels rose significantly as a result of irrigation in other areas as well as in the Boise River valley. The model-simulated value of 0.10 for the specific yield in layer 1 is higher than most values obtained from aquifer tests in the Boise River valley (table 1). The lower values for the aquifer tests may be a reflection of test problems and are probably not valid. The higher model-assigned value is probably a more valid estimate of storage for the range of water-level change simulated from preirrigation to 1980.

Ground-water budgets for transient simulation for 10-year periods from 1881 through 1980 are shown in table 12.

Simulated and measured water levels in selected wells are compared in figure 27. Most hydrographs based on simulated water levels are similar in shape to hydrographs based on measured water levels, although water-level altitudes differ. The best match is for well 2N-2W-34DAA1; the hydrograph is a typical one for shallow wells in the Boise River valley. Historical water levels in wells 4S-10E-30BBA1, 5S-4E-5CAA1, and others compared poorly with simulated heads. The main reason for the differences may be incorrectly assigned values of aquifer properties. Although many different combinations of values were tried, none produced the desired results. Another reason may be that the hydrographs represent local influences on water levels that the model cannot duplicate because of the relatively large grid size and other simplifying assumptions that were made. Pumped wells nearby may have caused measured water levels to be lower than simulated levels, whereas seepage from canals may have caused measured water levels to be higher.

SENSITIVITY ANALYSIS

The accuracy of model results depends on the conceptualization of the hydrologic system and the validity of simplifying assumptions. Although the values of model parameters used during the calibration process were reasonable, other values also would have been suitable.

The range of uncertainty of model results was evaluated by changing, in turn, the value of a single model parameter and holding the other values constant for a simulation. A range of values that bracketed each calibrated model value was used. Hydrographs for selected cells were used to demonstrate differences in water levels from the calibrated conditions that resulted from changes in underflow, specific yield, storage coefficient, vertical conductance, and transmissivity (figs. 28–32).

Model sensitivity to changes in underflow was tested by varying underflow from 0.5 to 2.0 times the calibrated value (fig. 28). Varying underflow values resulted in small hydraulic-head differences

TABLE 6.—*Estimated ground-water recharge from precipitation in each model subarea, 1980*

[Values are in acre-feet]

Model subarea (see fig. 14)	Total precipitation	Estimated recharge
1	110,000	4,000
2	170,000	5,000
3	210,000	6,400
4	180,000	5,400
5	120,000	3,500
6	510,000	6,900
7	80,000	3,000
8	140,000	1,800
9	130,000	400
10	270,000	2,000
11	100,000	1,900
Total	2,000,000 (rounded)	40,000 (rounded)

TABLE 7.—*Estimated ground-water pumpage and application rates in each model subarea, 1980*

Model subarea (see fig. 14)	Estimated pumpage (acre-feet)	Irrigated area (acres)	Application rate (acre-foot per acre)
1	10,600	3,000	3.5
2	18,000	10,800	1.7
3	58,800	13,700	4.3
4	7,100	2,800	2.5
5	400	1,400	.3
6	46,300	25,600	1.8
7	38,300	14,300	2.7
8	2,800	2,900	1.0
9	33,400	12,400	2.7
10	41,100	23,800	1.7
11	43,200	18,800	2.3
Total	300,000	129,000	
Average rate			2.2

in the Boise River valley, cells (10,20) and (18,17), locations of which are shown in figure 14. Head in these cells is affected by the specified head for drains in these cells. Changes in head are reduced by the capacity of drains to remove large amounts of water from the aquifer. Varying underflow values resulted in large head differences in cells (12,45), (22,44), (18,42), and (17,49), which represent areas of low transmissivity, several miles from gaining streams. Hydraulic heads increased when underflow was increased and decreased when underflow was decreased.

Values of specific yield were varied from 0.05 to 0.20; the calibrated value was 0.10 (fig. 29). As with underflow, changing specific yield in the Boise River valley resulted in little difference in water levels from the calibrated condition, partly because of the presence of drains in that area. The largest difference was in cells (12,45), (18,42), and (22,44). These cells are in areas of low transmissivity away from rivers or drains, where larger head changes are required to accommodate the low flow rates.

Model sensitivity to changes in storage coefficient was tested simultaneously for layers 2 and 3; values ranged from 0.10 to 10.0 times the calibrated value (fig. 30). The maximum difference in heads was about 20 ft, much less than for other properties tested. In general, a larger value of storage reduced head differences, and a smaller value increased head differences.

Vertical conductance was tested over a range from 0.10 to 10.0 times the calibrated values (fig.

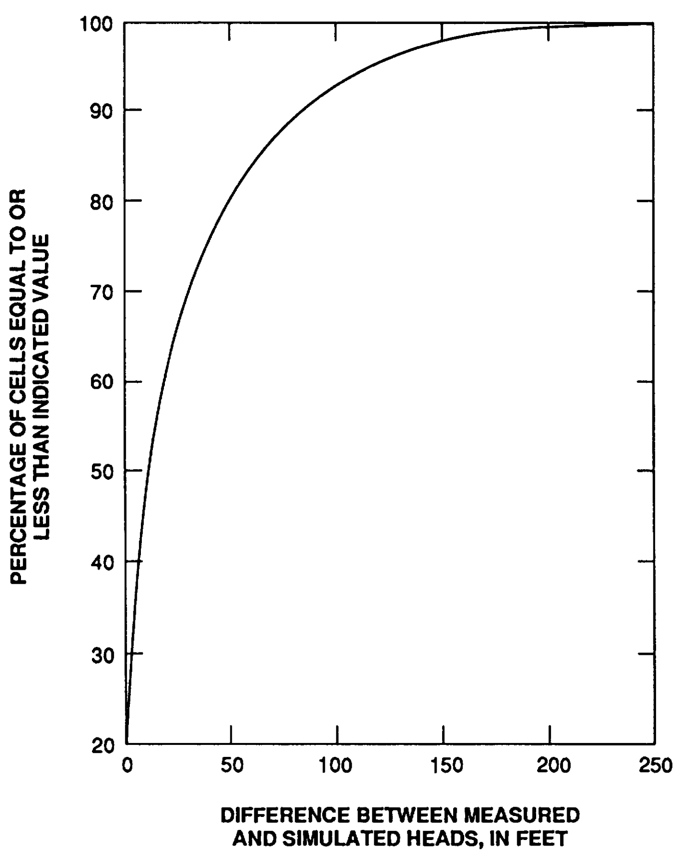


FIGURE 20.—Relation between measured and simulated heads for steady-state conditions, model layer 1. (Model layers shown in fig. 15.)

TABLE 8.—*Comparison of hydraulic conductivity estimated from steady-state model analysis with estimates from specific-capacity data, model layer 1*

Model subarea (see fig. 14)	Hydraulic conductivity (feet per day)	
	From steady-state model analysis	From specific-capacity data
1	14	8
2	17	16
3	39	9
4	4	40
5	6	14
6	3	7
7	4	11
8	43	79
9	17	63
10	43	54
11	17	25

31). The model was more sensitive to changes in vertical conductance than to any other parameter tested. The sensitivity was due to changes in hydrologic connections between aquifer layers that affect water flowing between the layers. As a result of several factors, changes in vertical conductance resulted in head differences of more than 200 ft in places—for example, cell (22,44) (fig. 31). Cells (10,20) and (18,17) are in the Boise River valley and contain drains. Heads in these cells are insensitive to changes in vertical conductance because water flowing from lower layers to the upper layer (layer 1) discharges to drains. Head changes in cells (22,44) and (18,42) indicate that in some areas, water in layer 3 flows upward, so that head in layer 3 is lowered greatly if the vertical conductance is increased. If vertical conductance is decreased, upward flow from layer 3 is reduced and the head in layer 3 is increased.

If vertical conductance is high, water in lower layers easily flows into layer 1 and initially increases the head; however, as pumping in lower layers is increased, heads in lower layers decrease accordingly and water in layer 1 flows downward, thus decreasing the heads in layer 1.

When vertical conductance is low, the heads in layer 1 initially are low because of less upward flow from lower layers. As pumping increases in the lower layers, heads in layer 1 are affected little because of the low vertical conductance. However, heads in layer 1 gradually are increased by recharge from irrigation water, which is held in layer 1 by the lower vertical conductance.

TABLE 9.—*Simulated steady-state and estimated river-reach gains, 1980*

[Values are in cubic feet per second]

River	Simulated gains	Estimated gains
Snake River:		
King Hill to Murphy	461	390
Murphy to Weiser	450	489
Boise River and drains	675	639
Payette River	263	300
Total	1,849	1,818

Transmissivity for layer 2 was tested over a range from 0.10 to 10.0 times the calibrated value (fig. 32). The largest changes are in an area near Mountain Home, cell (12,45). In general, the range in heads was small; the maximum variation was only about 20 ft. Again, in cells with drains, head increases were reduced because of the capacity of the drains to remove water.

During calibration, simulated ground-water discharge to rivers was compared with estimated or measured discharge. Model values of underflow, specific yield, storage coefficient, vertical conductance, and transmissivity were varied to show the sensitivity of ground-water discharge to these properties (fig. 33). Ground-water discharge to rivers is most sensitive to changes in underflow and vertical conductance.

Underflow is recharge to the model; therefore, an increase or decrease would be expected to increase or decrease outflow. Underflow has a cumulative effect on discharge, as does storage; these effects become more apparent with time.

Changes in leakance for rivers and drains alter the discharge relation between the river or drain and the aquifer. Because changes in leakance do not change total inflow to the model, changes in ground-water discharge to rivers or drains must be balanced by an increase or decrease in storage—that is, a rise or fall in the water table.

Changes in transmissivity for layers 1 and 2 had little effect on ground-water discharge to rivers, especially near the end of the simulation period, when conditions approached steady state.

SUMMARY AND CONCLUSIONS

The western Snake River Plain is part of a deep structural depression that extends across southern

TABLE 10.—*Assigned transmissivity values, model layers 2 and 3*

[Values are in feet squared per day]

Model subarea (see fig. 14)	Transmissivity	
	Layer 2	Layer 3
1	8,600	8,600
2	8,600	8,600
3	8,600	8,600
4	8,600	8,600
5	3,300	8,600
6	12,100	8,600
7	8,600	8,600
8	2,600	8,600
9	3,500	8,600
10	900	8,600
11	8,600	8,600

TABLE 11.—*Assigned vertical hydraulic conductivity between model layers*

[Values are in feet per day]

Model subarea (see fig. 14)	Vertical hydraulic conductivity	
	Layer 2	Layer 3
1	9	22
2	9	22
3	9	22
4	9	22
5	9	22
6	9	22
7	9	22
8	900	22
9	68	22
10	9	22
11	9	22

Idaho. Three major rock units fill the depression. The upper unit consists of about 500 ft of sedimentary and volcanic rocks; the middle unit, about 4,000 ft of fine-grained sedimentary rocks and volcanic rocks; and the lower unit, about 7,000 ft of volcanic rocks.

In general, aquifers in the upper unit are unconfined, although locally they are confined. Water is present in perched systems near Mountain Home and in the Boise River valley. Aquifers in the middle and lower rock units are confined, and water in some wells completed in these units flows at land surface.

Ground-water levels measured in the spring of 1980 indicated that the principal direction of flow in the upper rock unit is toward rivers and streams. In much of the western plain, water in the middle and lower rock units moves vertically into the upper rock unit.

Water levels are not always an accurate measure of the hydraulic head in individual aquifers. Many wells are not cased, and the water levels are a composite of hydraulic heads in multiple aquifers. Few wells are completed in the middle and lower rock units.

Sources of ground-water recharge are infiltration of surface water used for irrigation, underflow across study-area boundaries, and precipitation on the plain. In 1980, recharge from application of irrigation water, including precipitation on irrigated lands, was estimated to be about 1,400,000 acre-ft; recharge from precipitation on nonirrigated lands, about 40,000 acre-ft; and underflow, about 310,000 acre-ft.

Ground-water discharge is largely seepage to rivers and drains (canals) and ground-water pumpage. Seepage to rivers and drains was estimated to be about 1,450,000 acre-ft in 1980; ground-water pumpage was about 300,000 acre-ft.

Aquifer tests indicate that the transmissivity of the upper rock unit ranges from 5,000 to 230,000 ft^2/d, and that the specific yield in this unit ranges from 0.025 to 0.43. Specific-capacity data obtained from drillers' logs also were used to estimate transmissivity of the upper rock unit. Transmissivity thus estimated was less than 8,000 ft^2/d in 80 percent of the wells. Transmissivities are lowest in fine-grained sedimentary rocks in the middle unit and are between the middle-unit and upper-unit values in the volcanic rocks that compose the lower unit. Estimates of storage coefficient for the middle and lower rock units were not available from aquifer tests. No field estimates of vertical hydraulic conductivity were available.

A computer model was developed to simulate steady-state and transient hydrologic conditions. During steady-state calibration, vertical hydraulic conductivity and transmissivity were adjusted to achieve a match between simulated and spring 1980 measured water levels and discharge. Ground-water levels, recharge, and discharge were reasonably stable in 1980 and were assumed to represent steady-state conditions. Water levels for the period 1881 through 1980 were simulated during the transient analysis.

Hydraulic properties, such as vertical hydraulic conductivity and transmissivity, probably are the least known. They vary over a range of several or-

TABLE 12.—*Ground-water budgets based on transient simulations, 1881–1980*

[Values are in thousands of acre-feet]

	Inflow			Outflow			
10-year period ending year	Average annual under-flow (U)	Average annual recharge from irrigation (I)	River losses (R_L)	Pumpage (P)	Drains (D)	River gains (R_G)	Net change in storage (ΔS)[1]
1890	310	41	17	0	28	341	–1
1900	310	185	16	0	51	382	78
1910	310	262	14	0	79	412	95
1920	310	651	10	0	201	531	239
1930	310	819	8	0	300	609	228
1940	310	1,011	7	7	403	694	224
1950	310	1,221	4	10	502	780	143
1960	310	1,362	4	56	579	849	192
1970	310	1,400	6	144	589	869	114
1980	310	1,442	9	300	575	865	21

[1] $\Delta S = (U + I + R_L) - (P + D + R_G)$. A positive change in storage indicates a net rise in water levels; a negative change in storage indicates a net decline in water levels.

ders of magnitude, both within the same rock type and between different rock types. Storage coefficient also varies over a wide range of values. Estimates of recharge and discharge generally are accurate to within less than a single order of magnitude. However, the distribution of recharge, such as underflow, is critical to proper simulation; thus, the model may be sensitive to even small changes.

Model-assigned transmissivity values for layer 1 (upper rock unit) ranged from 1,500 to 21,500 ft^2/d and, for layer 2 (middle rock unit), from 900 to 12,100 ft^2/d. A uniform transmissivity value of 8,600 ft^2/d was assigned to layer 3 (lower rock unit). On the basis of model analysis, the vertical hydraulic conductivity ranged from 9 to 900 ft/d. Storage coefficients of layers 2 and 3 were estimated to be 4×10^{-3} and 7×10^{-3}, respectively. Calibrated model values of transmissivity, vertical hydraulic conductivity, storage coefficient, and specific yield are probably of the same order of magnitude as the estimated values that are based on field data; therefore, the general concept of the ground-water-flow system presented in this report is considered reasonable.

This model is considered calibrated only for the upper unit (layer 1), where simulated hydraulic heads approximate measured heads. Lack of hydrogeologic data for the middle and lower rock units prevented an acceptable calibration of model layers 2 and 3. The model is useful for understanding the ground-water-flow system but not for detailed management evaluations.

Model calibration indicates that the most needed data are vertical hydraulic-head distribution in the upper and middle rock units, hydraulic properties of aquifers and confining beds, underflow, and refined knowledge of subsurface hydrogeology.

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FIGURES 21–33

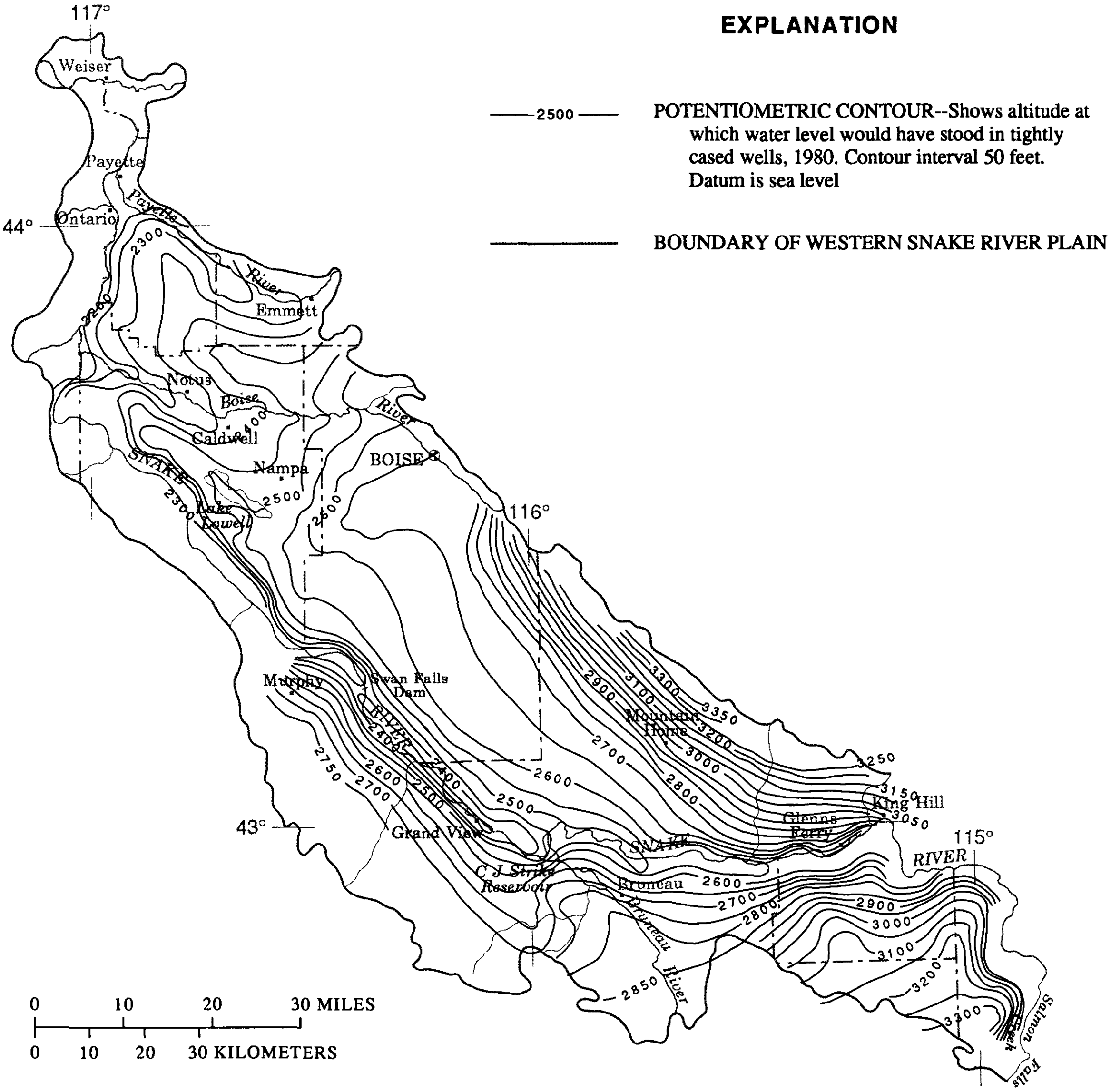


FIGURE 21.—Simulated 1980 steady-state potentiometric surface, model layer 1. (Model layers shown in fig. 15.)

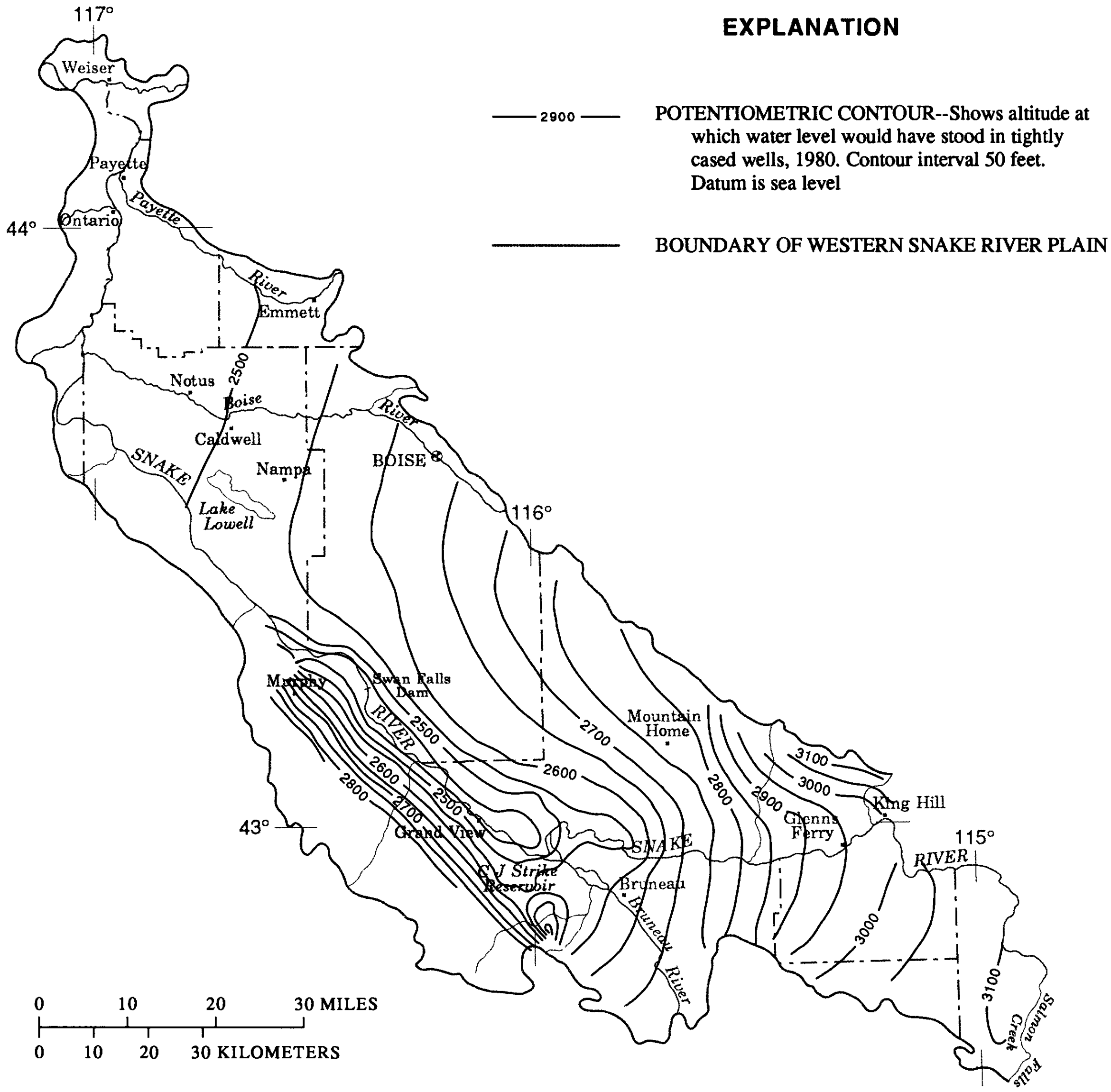


FIGURE 22.—Simulated 1980 steady-state potentiometric surface, model layer 2. (Model layers shown in fig. 15.)

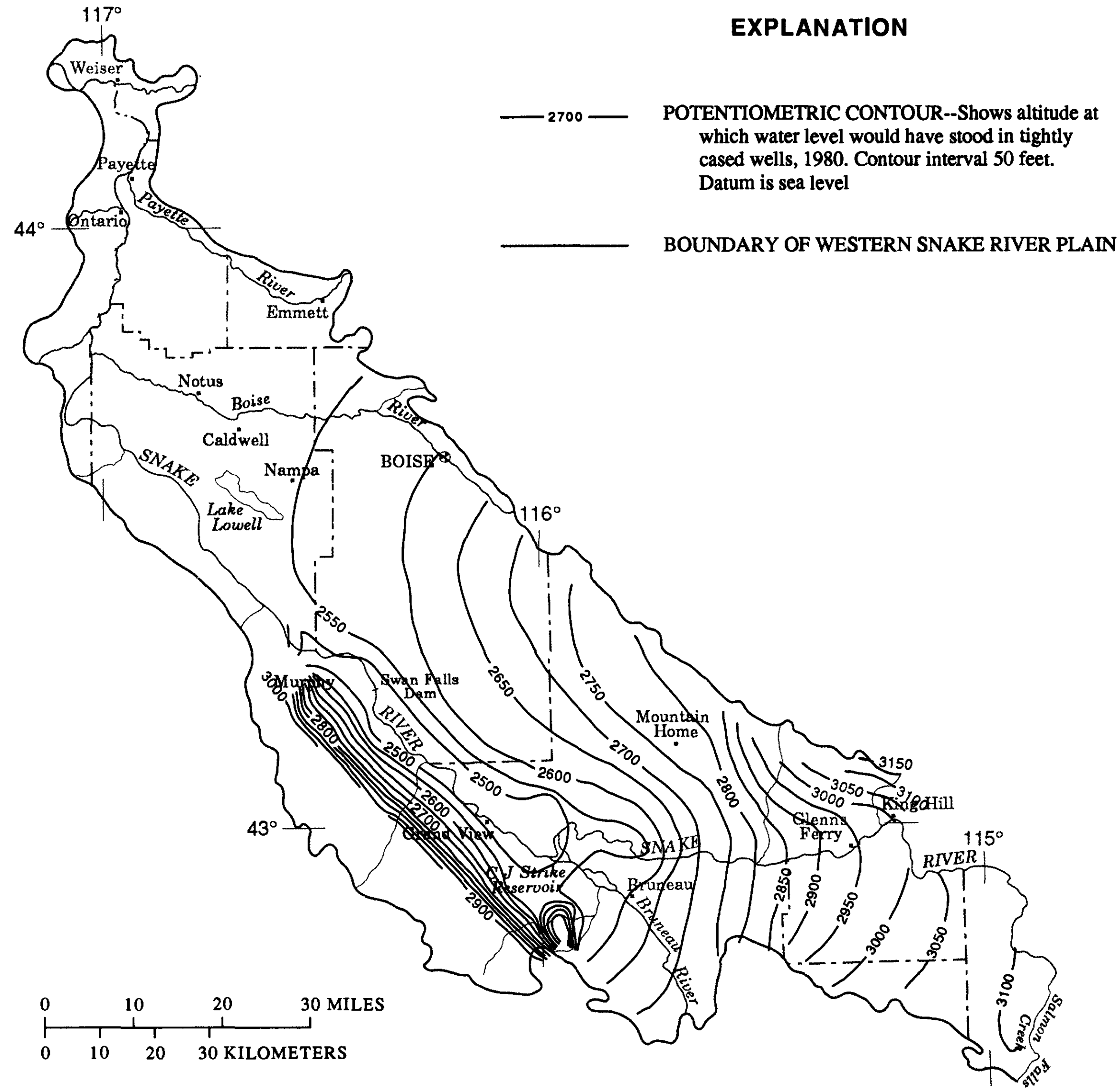
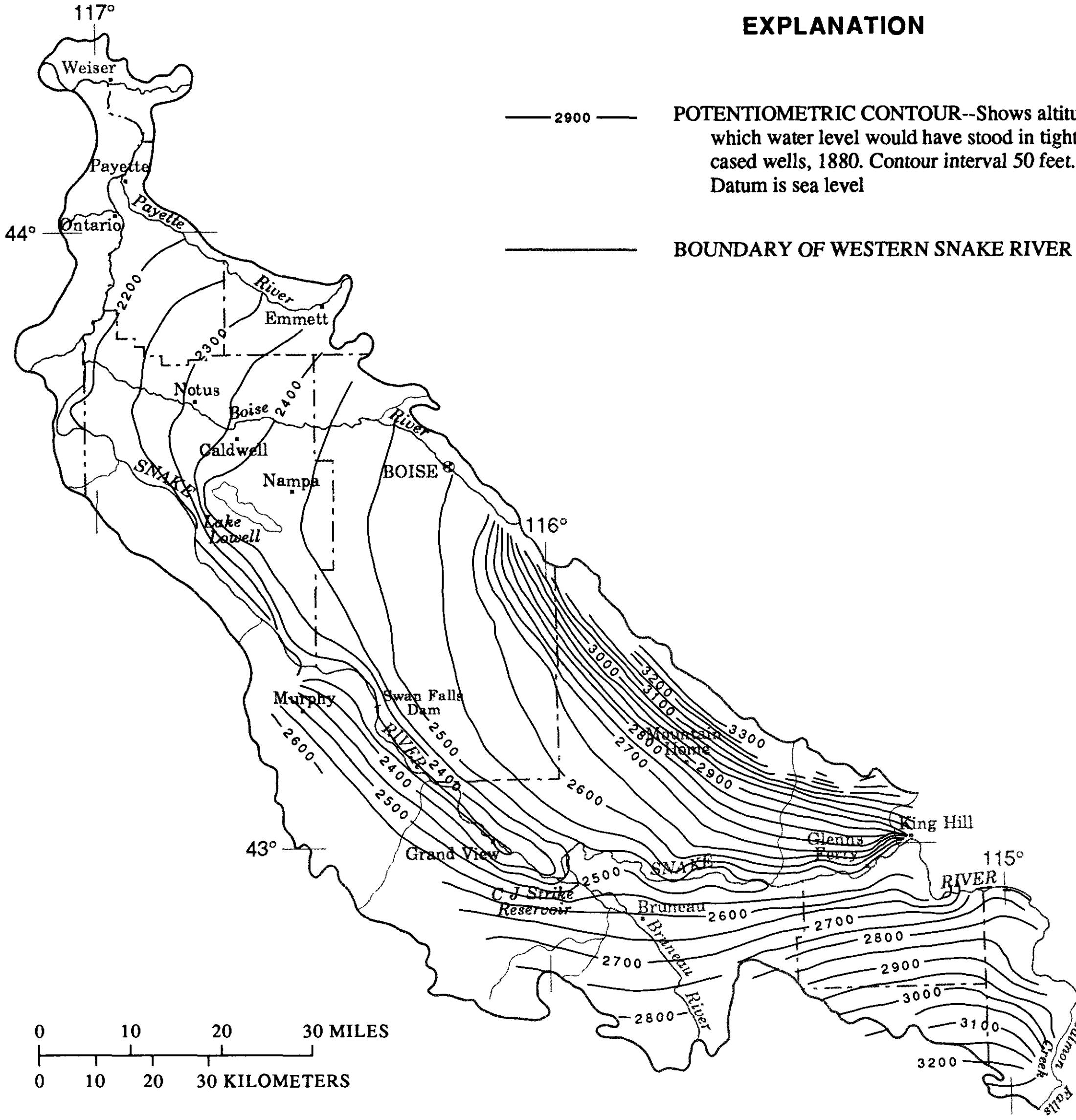


FIGURE 23.—Simulated 1980 steady-state potentiometric surface, model layer 3. (Model layers shown in fig. 15.)



EXPLANATION

—— 2900 —— POTENTIOMETRIC CONTOUR--Shows altitude at which water level would have stood in tightly cased wells, 1880. Contour interval 50 feet. Datum is sea level

———— BOUNDARY OF WESTERN SNAKE RIVER PLAIN

FIGURE 24.—Simulated 1880 preirrigation potentiometric surface, model layer 1. (Model layers shown in fig. 15.)

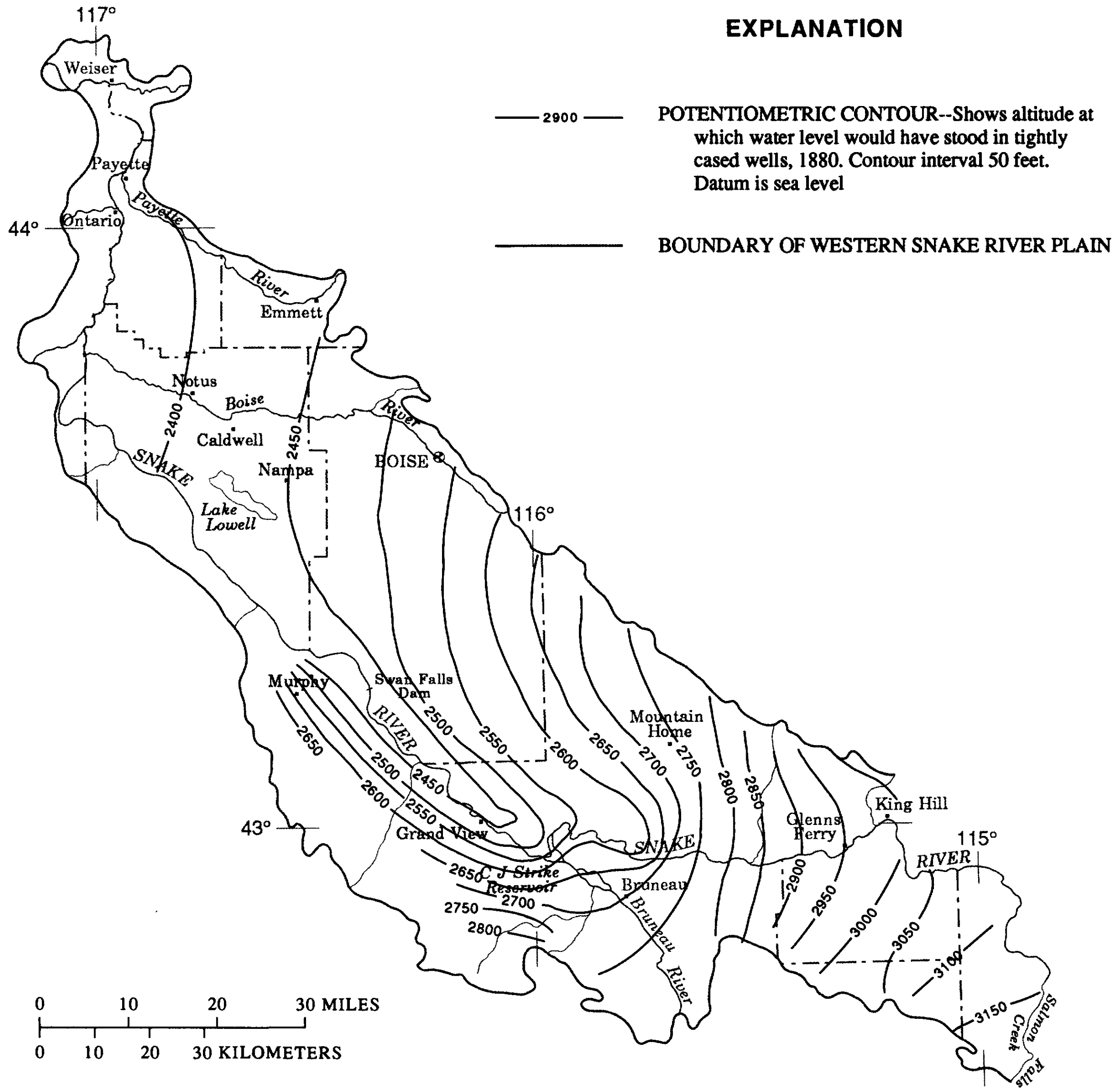


FIGURE 25.—Simulated 1880 preirrigation potentiometric surface, model layer 2. (Model layers shown in fig. 15.)

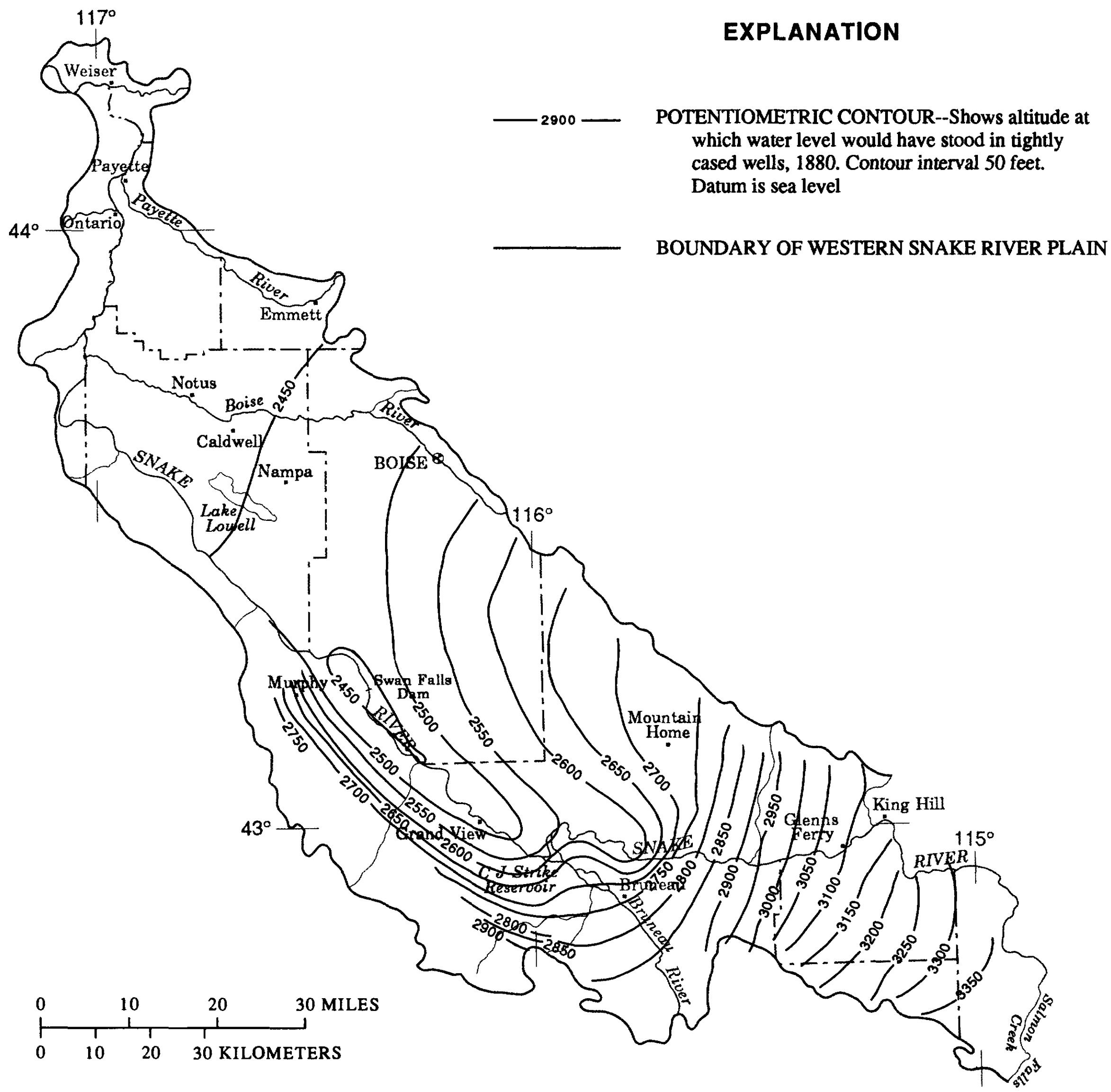


FIGURE 26.—Simulated 1880 preirrigation potentiometric surface, model layer 3. (Model layers shown in fig. 15.)

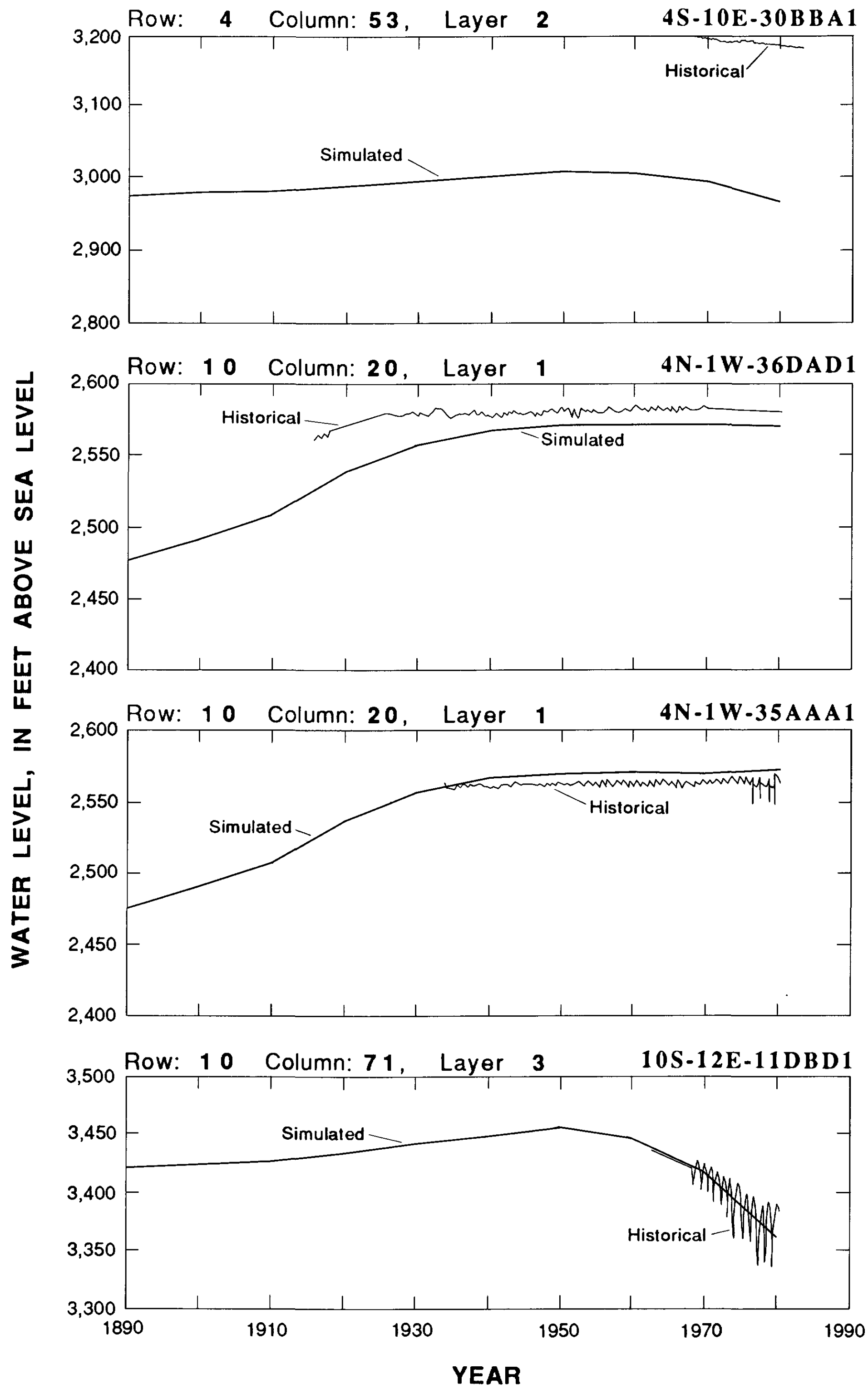


FIGURE 27.—Simulated and measured water levels, 1890–1980. (Cell locations shown in fig. 14; model layers shown in fig. 15.)

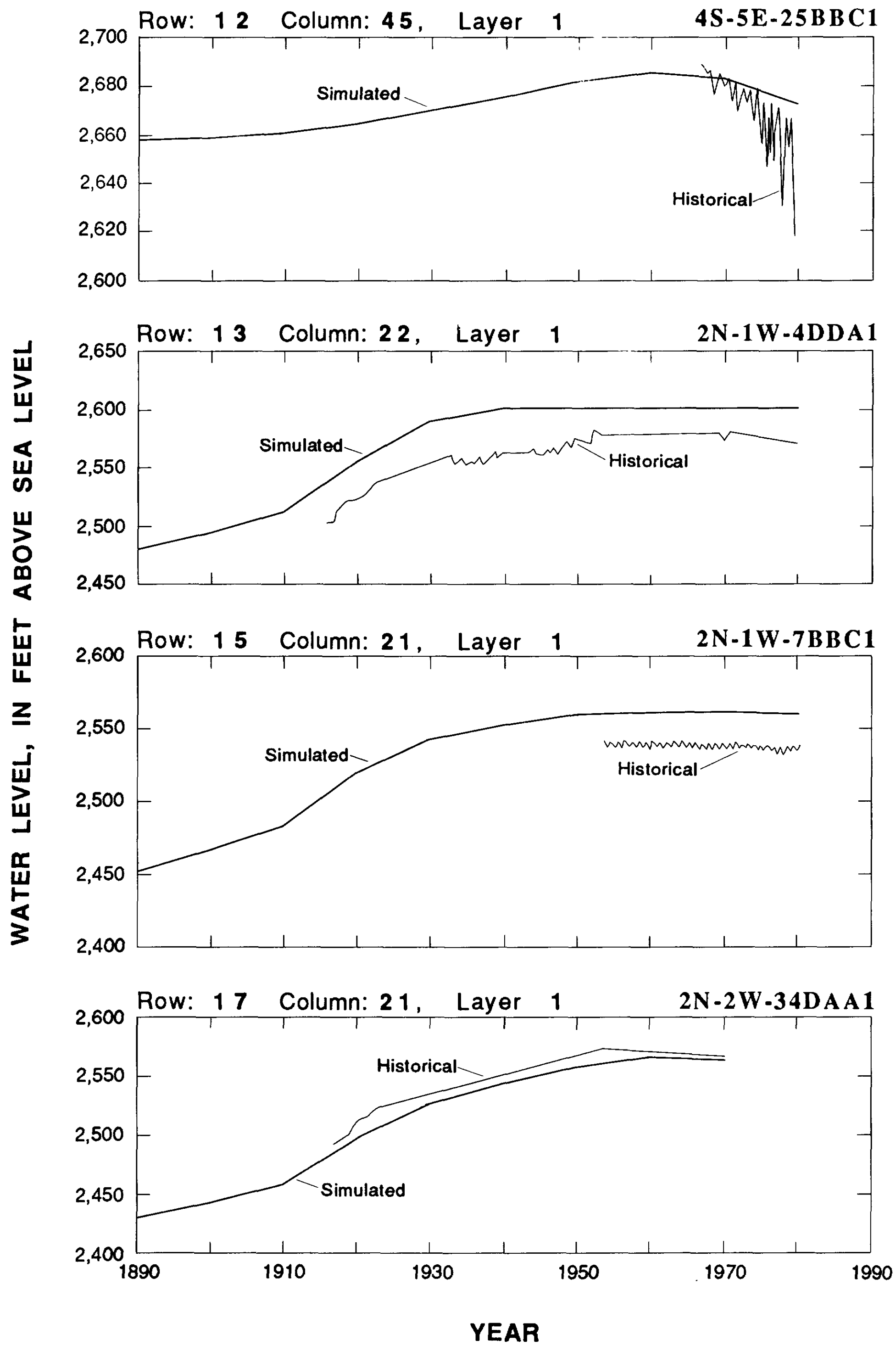


FIGURE 27.—*Continued.*

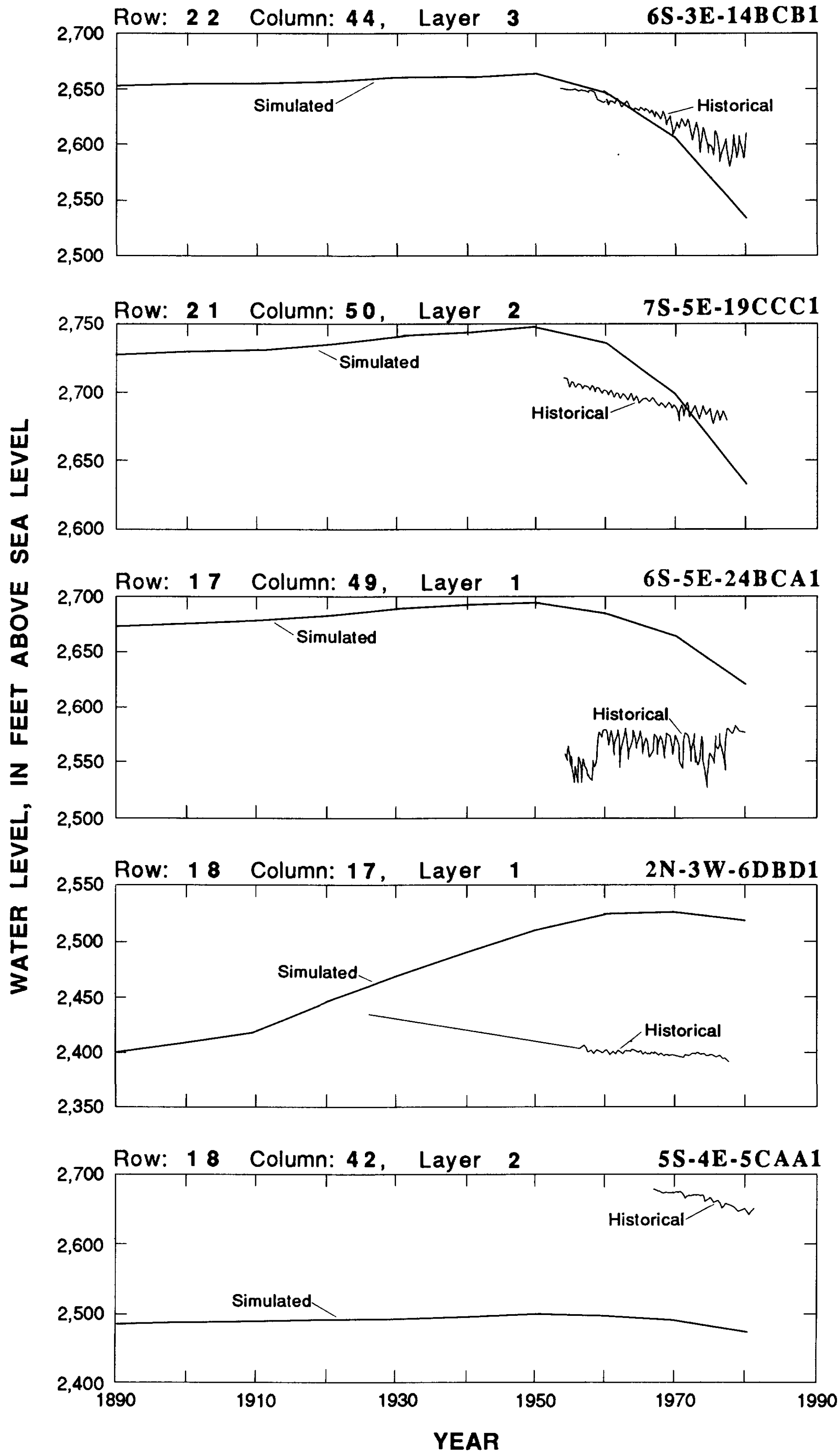


FIGURE 27.—*Continued.*

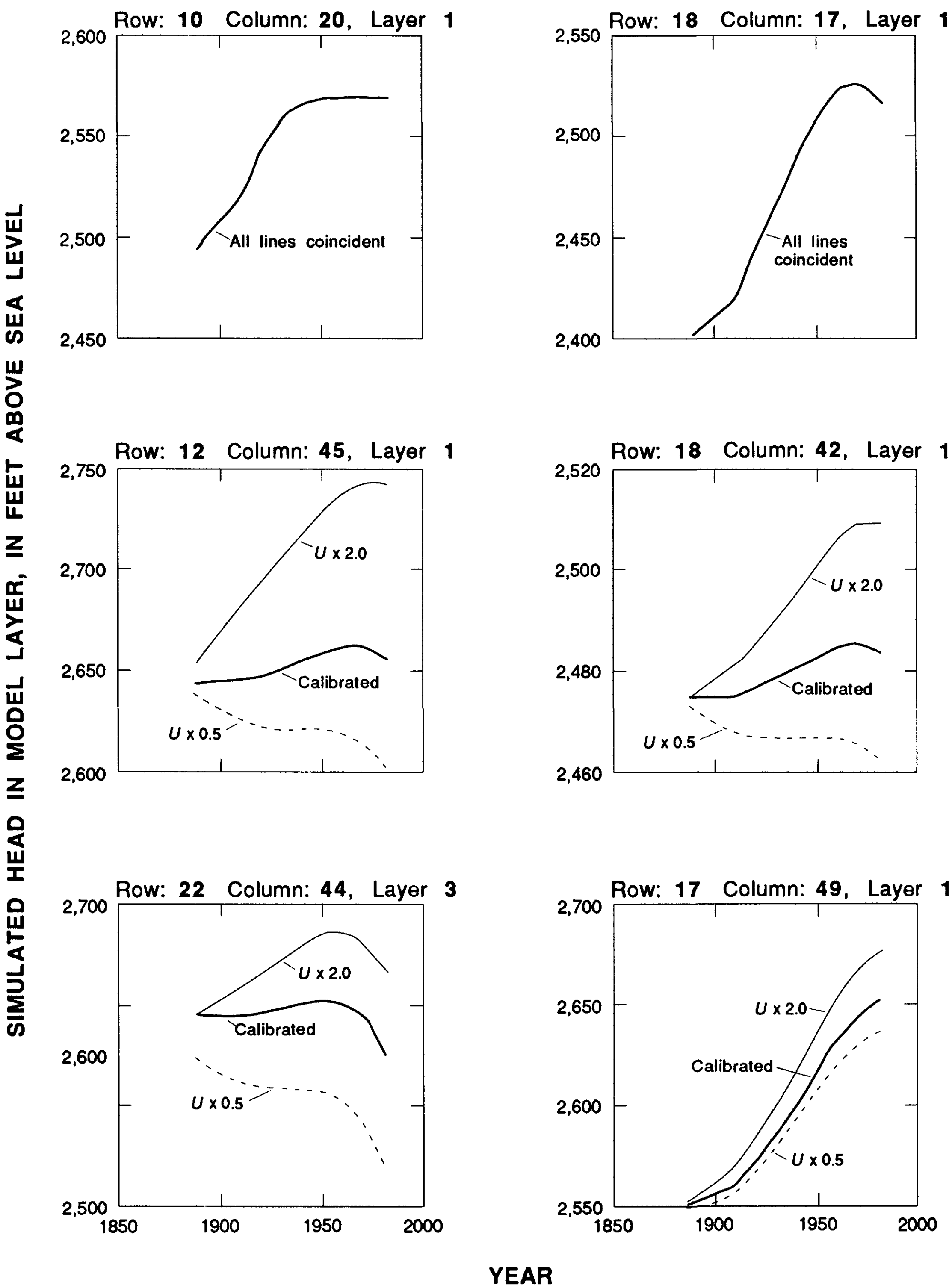


FIGURE 28.—Model sensitivity to changes in U (underflow). (Cell locations shown in fig. 14; model layers shown in fig. 15.)

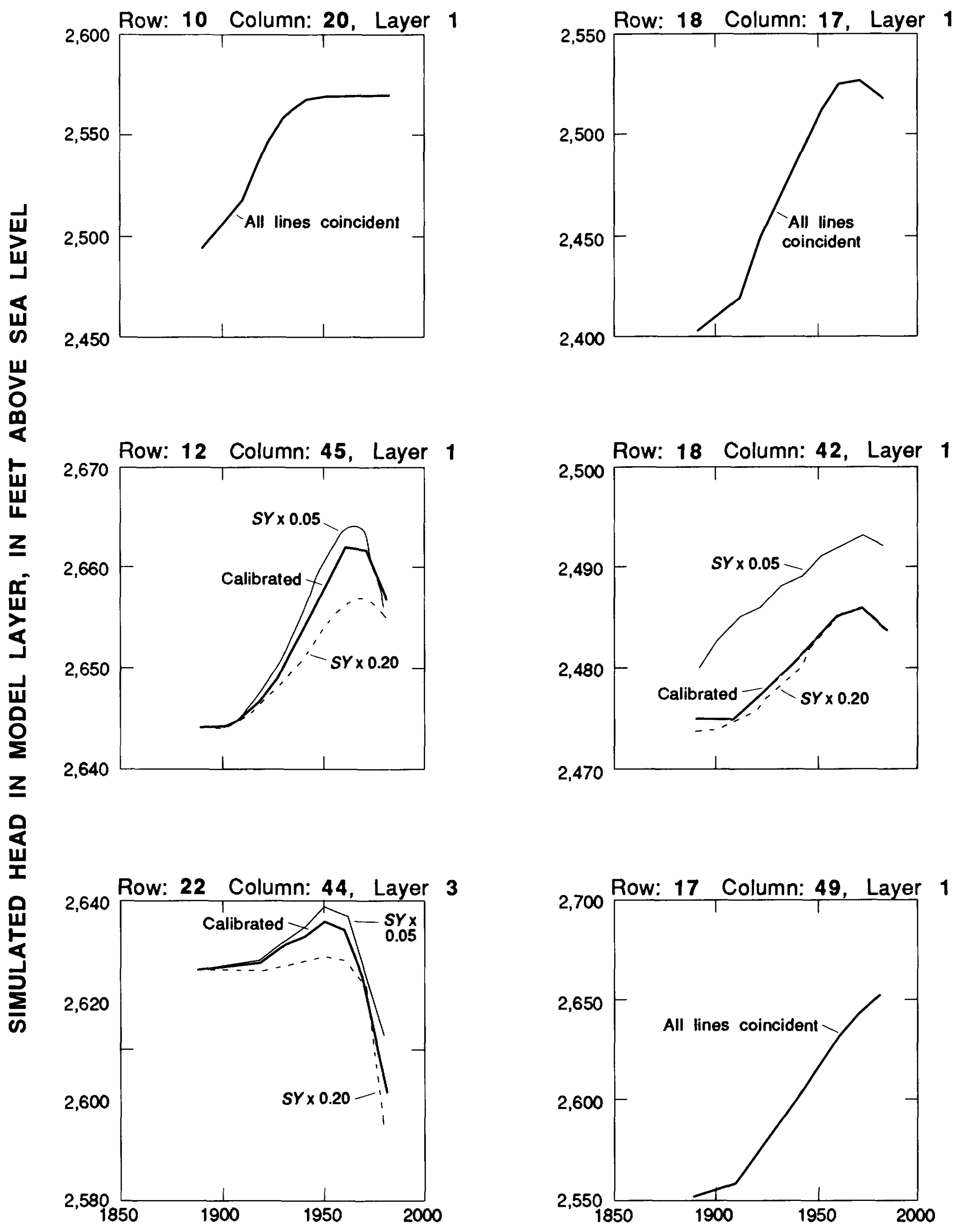


FIGURE 29.—Model sensitivity to changes in *SY* (specific yield), model layer 1. (Cell locations shown in fig. 14; model layers shown in fig. 15.)

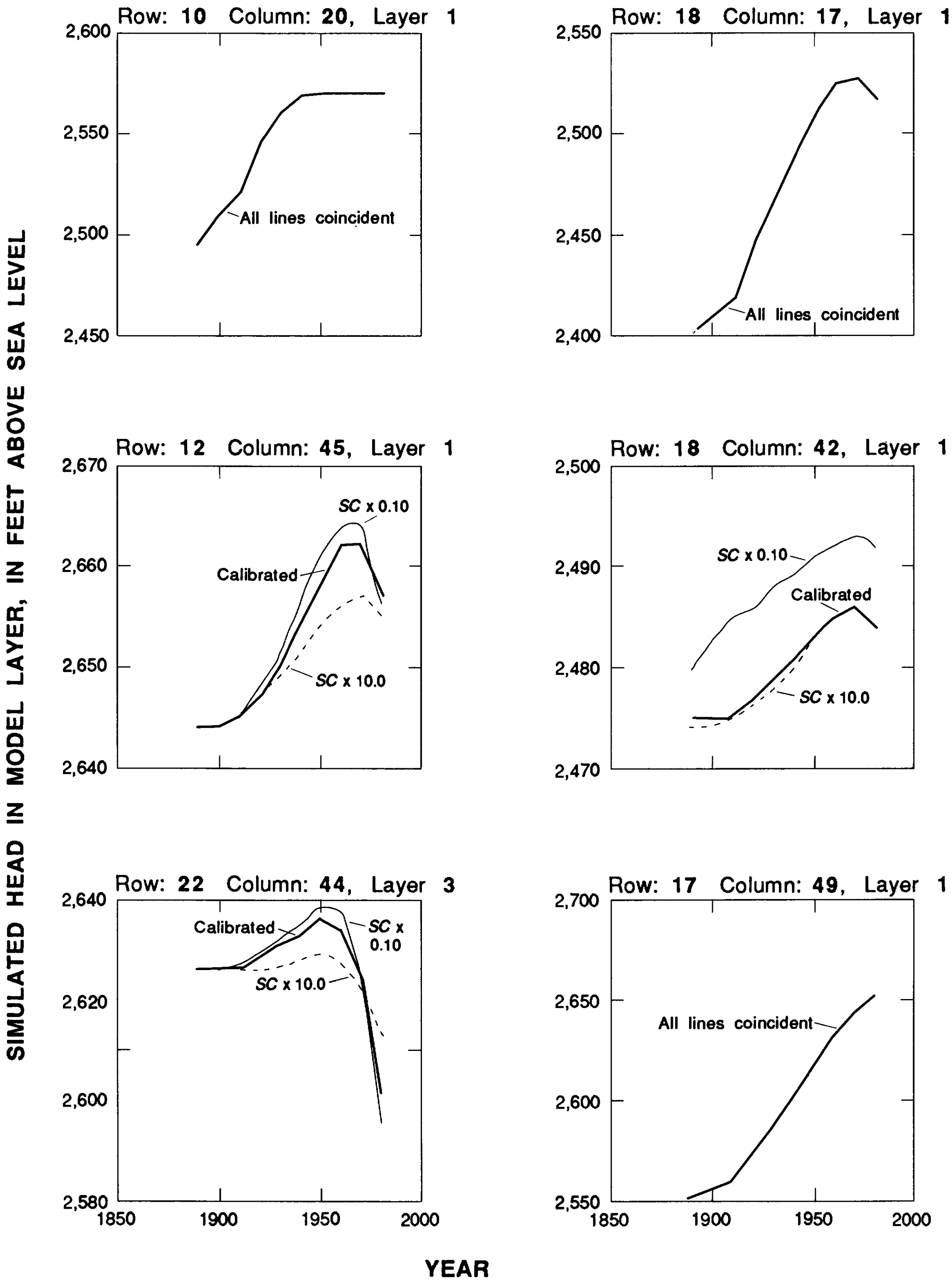


FIGURE 30.—Model sensitivity to changes in *SC* (storage coefficient), model layers 2 and 3. (Cell locations shown in fig. 14; model layers shown in fig. 15.)

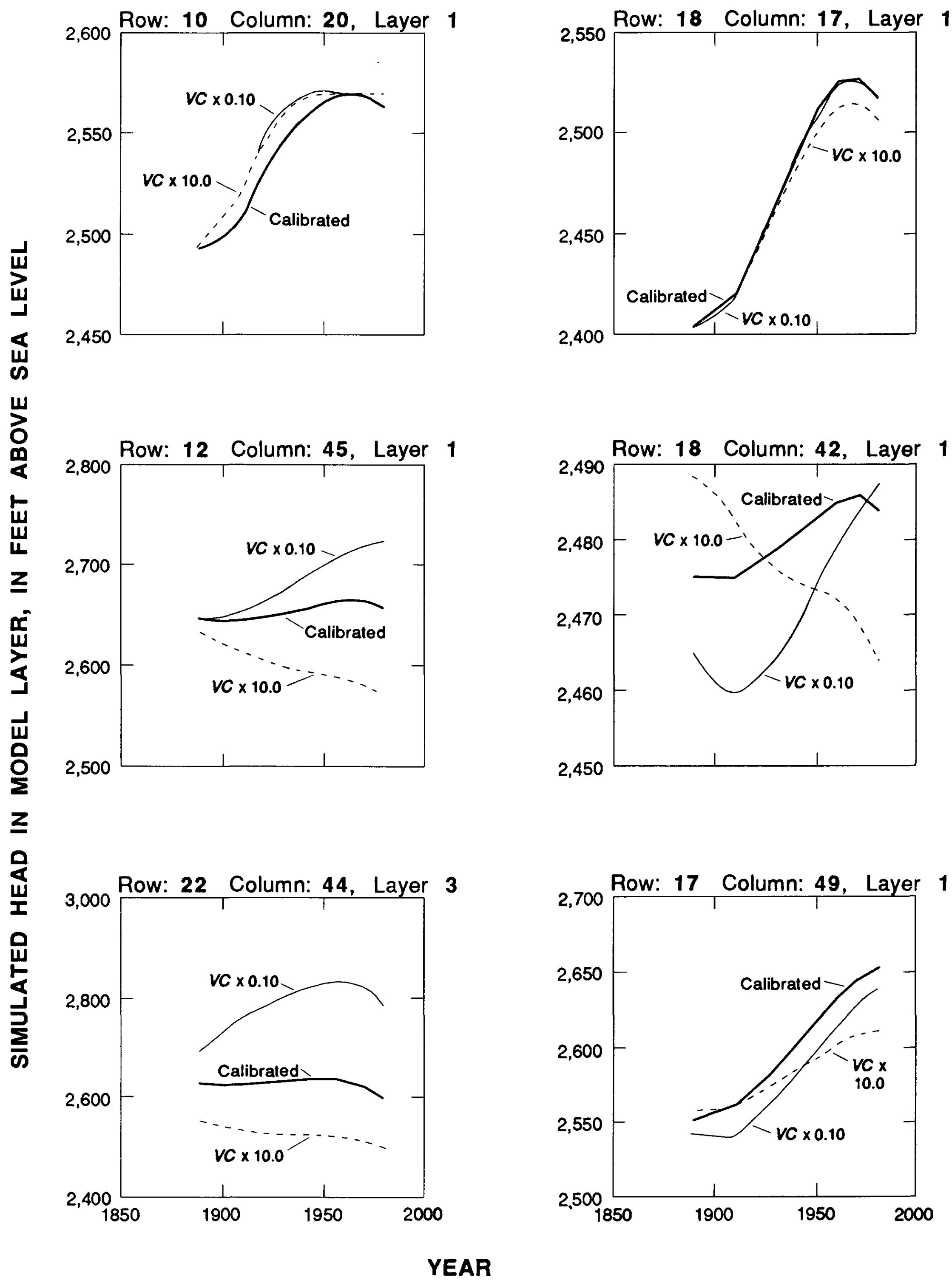


FIGURE 31.—Model sensitivity to changes in *VC* (vertical conductance). (Cell locations shown in fig. 14; model layers shown in fig. 15.)

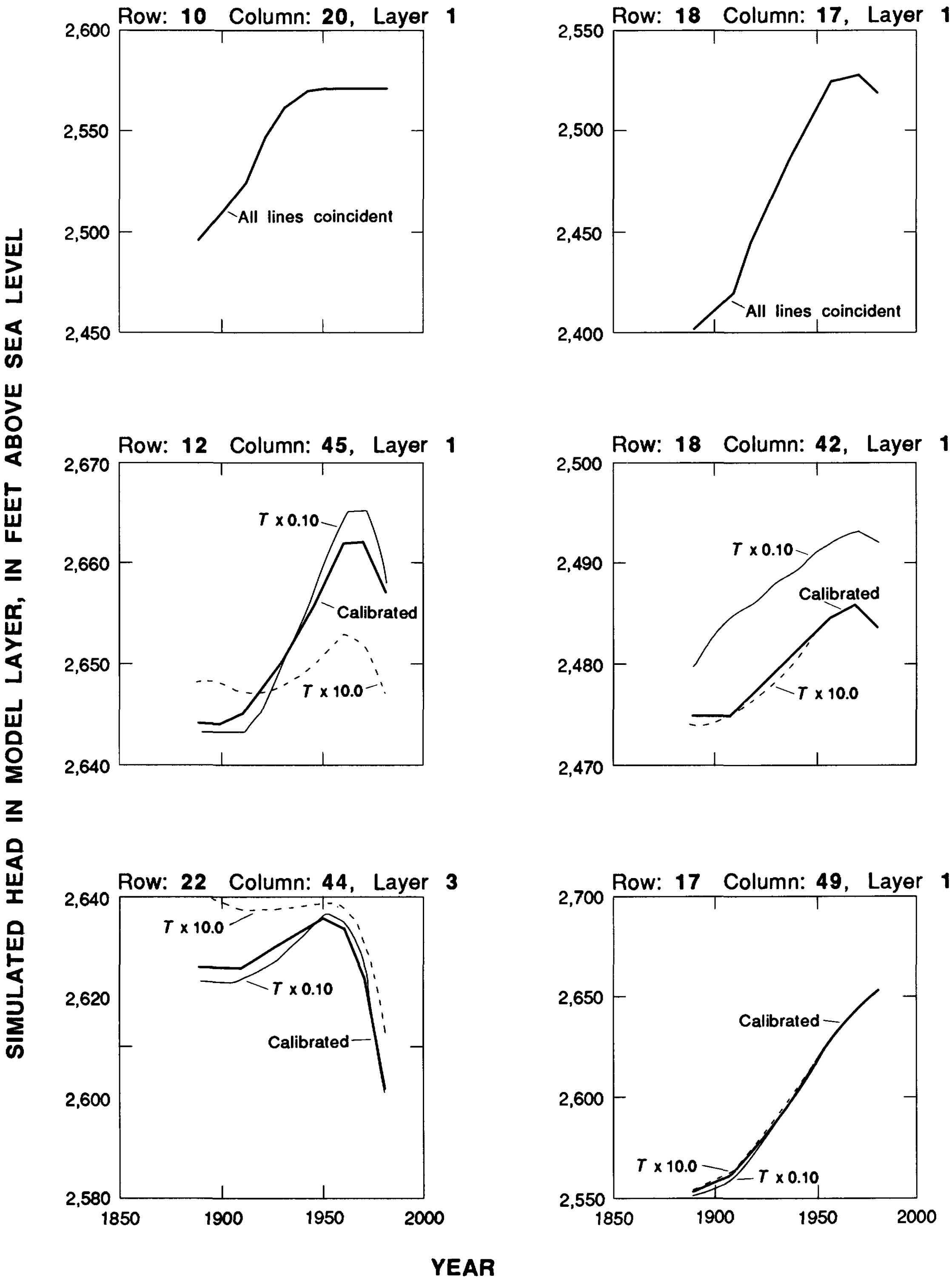


FIGURE 32.—Model sensitivity to changes in T (transmissivity), model layer 2. (Cell locations shown in fig. 14; model layers shown in fig. 15.)

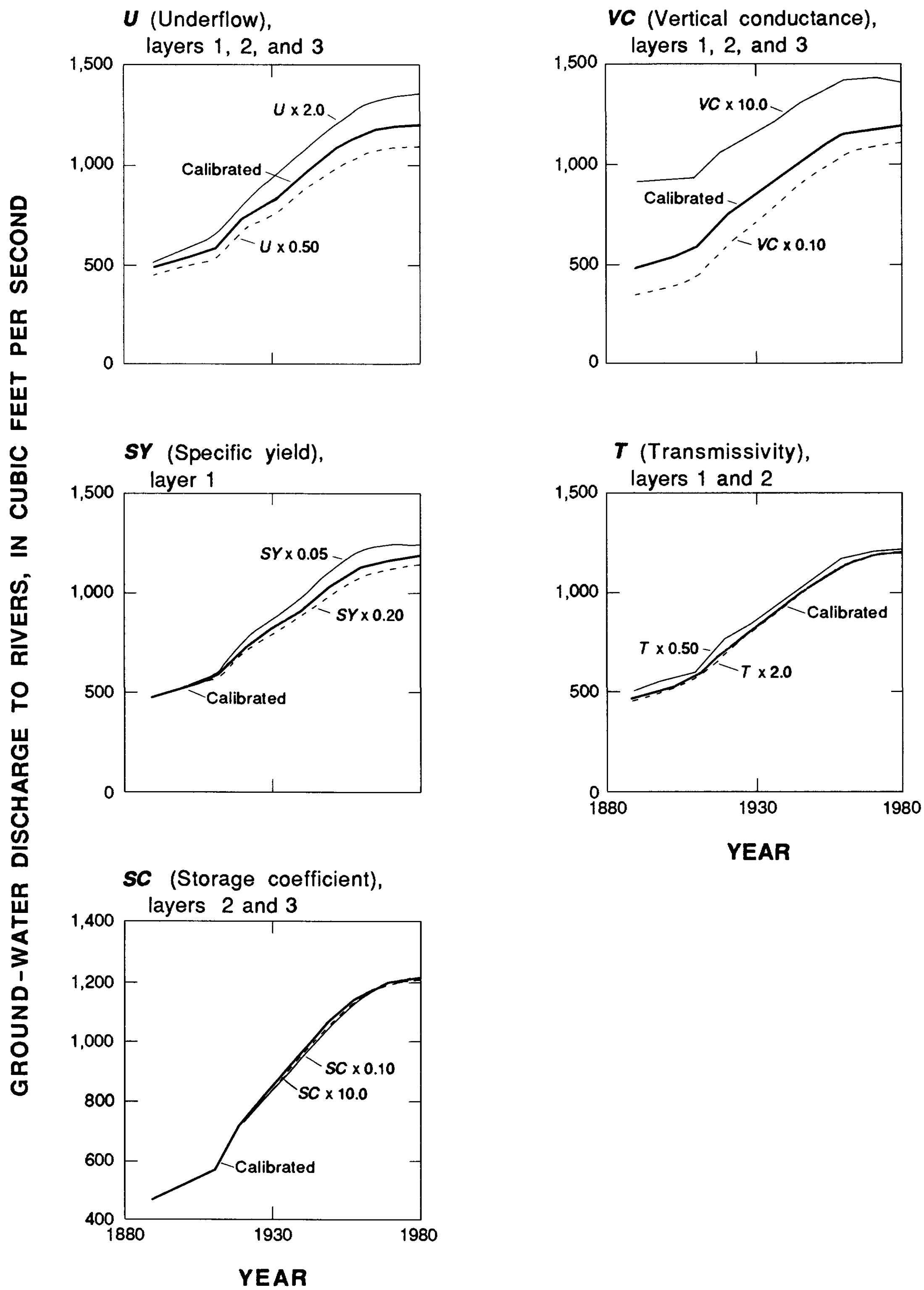


FIGURE 33.—Model sensitivity in terms of ground-water discharge to rivers owing to changes in underflow, specific yield, storage coefficient, vertical conductance, and transmissivity.

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