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HYDROLOGY AND DIGITAL SIMULATION OF THE REGIONAL AQUIFER SYSTEM, EASTERN SNAKE RIVER PLAIN, IDAHO

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1408-F

Hydrology and Digital Simulation of the Regional Aquifer System, Eastern Snake River Plain, Idaho

By S.P. GARABEDIAN

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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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ILLUSTRATIONS

[Plates in separate case]

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CONVERSION FACTORS AND ABBREVIATIONS

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 (acre-ft)	1,233	cubic meter (m^3)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
foot (ft)	0.3048	meter (m)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
inch (in.)	25.4	millimeter (mm)
mile (mi)	1.609	kilometer (km)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
square foot (ft^2)	0.0929	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)

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 net of both the United States and Canada, formerly called "Sea Level Datum of 1929."

HYDROLOGY AND DIGITAL SIMULATION OF THE REGIONAL AQUIFER SYSTEM, EASTERN SNAKE RIVER PLAIN, IDAHO

By S.P. Garabedian

ABSTRACT

The occurrence and movement of water in the regional aquifer system that underlies the eastern Snake River Plain, Idaho, depend on the transmissivity and storage capacity of rocks that compose the geologic framework and on the distribution and amount of recharge and discharge of water within that framework. On a regional scale, most water moves horizontally through interflow zones in Quaternary basalt of the Snake River Group. In recharge and discharge areas, water also moves vertically along joints and interfingering edges of basalt flows. Aquifer thickness is largely unknown, but geophysical studies suggest that locally the Quaternary basalt may exceed several thousand feet. Along the margins of the plain, sand and gravel several hundred feet thick transmit large volumes of water.

Regional ground-water movement is generally from northeast to southwest, from areas of recharge to areas of discharge. Recharge is from seepage of surface water used for irrigation, stream and canal losses, underflow from tributary drainage basins, and infiltration of precipitation. Aquifer discharge is largely spring flow to the Snake River and water pumped for irrigation. Major springs are near American Falls Reservoir and along the Snake River from Milner Dam to King Hill.

Regional ground-water flow was simulated with numerical models. Initially, a two-dimensional steady-state model that included a nonlinear, least-squares regression technique was used to estimate aquifer properties. Later, a three-dimensional steady-state and transient model was used to replace the two-dimensional model. Three-dimensional model results indicated that average total transmissivity ranged from about 0.05 to 120 feet squared per second and vertical leakance ranged from about 3×10^{-10} to 5×10^{-6} feet per second per foot of aquifer thickness.

The three-dimensional transient model was used to compare measured and estimated long-term changes in ground-water discharge and water levels with simulated values. Initial head conditions used in transient simulations were derived from a steady-state solution of estimated preirrigation hydrologic conditions. Transient simulations were 5-year stress periods beginning in 1891 and ending in 1980. Recharge for each stress period from 1926 to 1980 was estimated from surface-water irrigation, precipitation, and streamflow records. Recharge for stress periods from 1891 to 1925 was based on the average value for stress periods from 1926 to 1980 and was indexed to estimated irrigated acreages. Average annual tributary drainage-basin underflow for stress periods from 1891 to 1910 was calculated by using basin-yield equations. Underflow for stress periods from 1911 to 1980 was varied by use of streamflow records.

Transient simulations reasonably approximated measured changes in aquifer head and ground-water discharge that resulted from use of surface water for irrigation. Irrigation with surface water peaked in about 1950; subsequent increases in irrigation have been supplied largely by ground water. The three-dimensional model simulated water-level declines and reduced ground-water discharge caused in part by increases in ground-water pumping.

The transient model was used to simulate aquifer changes from 1981 to 2010 in response to three hypothetical development alternatives: (1) Continuation of 1980 hydrologic conditions, (2) increased pumpage, and (3) increased recharge. Simulation of continued 1980 hydrologic conditions for 30 years indicated that head declines of 2 to 8 feet might be expected in the central part of the plain. The magnitude of simulated head declines was consistent with head declines measured during the 1980 water year. Larger declines were calculated along model boundaries, but these changes may have resulted from underestimation of tributary drainage-basin underflow and inadequate aquifer definition. Simulation of increased ground-water pumpage (an additional 2,400 cubic feet per second) for 30 years indicated head declines of 10 to 50 feet in the central part of the plain. These relatively large head declines were accompanied by increased simulated river leakage of 50 percent and decreased spring discharge of 20 percent. The effect of increased recharge (800 cubic feet per second) for 30 years was a rise in simulated heads of 0 to 5 feet in the central part of the plain.

INTRODUCTION

The Snake River Plain regional aquifer study is one of the studies undertaken in the U.S. Geological Survey's Regional Aquifer-System Analysis (RASA) program. As stated by Lindholm (1981), the purposes of the study were to (1) refine knowledge of the regional ground-water-flow system, (2) determine effects of conjunctive use of ground and surface water, and (3) describe the chemistry of ground water. Preliminary interpretive reports generated by the Snake River Plain RASA study as of 1988 include (1) a regional water-table map and description of the ground-water-flow system (Lindholm and others, 1983, 1988); (2) results of geohydrologic test drilling in the eastern Snake River Plain (Whitehead and Lindholm, 1985); (3) water withdrawals for irrigation (Bigelow and others, 1986); (4) a ground-water-flow model of the eastern Snake River Plain (Garabedian, 1986); (5) water budgets and flow in the Snake River (Kjelstrom, 1986); (6) a map of land use showing irrigated acreage (Lindholm and Goodell, 1986); (7) a description of the geohydrologic framework

(Whitehead, 1986); and (8) a description of surface- and ground-water quality (Low, 1987).

Final interpretive results of the Snake River Plain RASA study are presented in 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 systems.

Chapter E describes water use.

Chapter F (this report) describes results of ground-water-flow modeling of the eastern Snake River Plain.

Chapter G describes results of ground-water-flow modeling of the western Snake River Plain.

PURPOSE AND SCOPE

This report describes the use of a ground-water-flow model to refine and extend knowledge of the regional ground-water-flow system in the eastern Snake River Plain (fig. 1). Two-dimensional ground-water-flow models were used in previous studies to simulate a hydrologic system that is largely three dimensional. Therefore, in this study, a three-dimensional model was used to (1) evaluate the significance of vertical variations in hydraulic conductivity and changes in head with depth, (2) evaluate the effect of sediment interbeds on regional ground-water flow, (3) simulate historical changes in the hydrologic system as a result of irrigation, and (4) estimate future hydrologic changes that might result from implementing various management alternatives.

LOCATION AND DESCRIPTION OF STUDY AREA

The eastern Snake River Plain extends across southern Idaho (fig. 1) and is about 170 mi long, 60 mi wide, and 10,800 mi² in area. Altitudes range from about 2,500 ft above sea level near King Hill (pl. 1) on the west to more than 6,000 ft in the northeastern part of the plain. Mountains bordering the plain are 7,000 to 12,000 ft in altitude.

The eastern plain is entirely within the Snake River drainage basin. Major tributaries that contribute flow directly to the Snake River upstream from King Hill are the Henrys Fork of the Snake River (hereafter referred to as Henrys Fork); the Blackfoot, Portneuf, and Big Wood Rivers; and the Salmon Falls Creek (pl. 1). Tributary streams along the northwestern edge of the plain, with the exception of the Big Wood River, lose all their flow to infiltration and evapotranspiration after reaching the plain; these streams include the Big Lost River, Little Lost River, Birch Creek, Medicine Lodge Creek, Beaver Creek, and Camas Creek. Most tributary streams originate in intermontane valleys that are generally perpendicular to the longitudinal axis of the eastern plain. Peak flows in unregulated streams are primarily from snowmelt during spring and early summer. Most regulated streams have reduced peak flows and higher average summer flows when stored surface water is released and diverted for irrigated agriculture.

Annual precipitation on much of the plain ranges from 8 to 10 in. (fig. 2), whereas precipitation on higher mountains within the Snake River basin exceeds 60 in. Most precipitation on the mountains is winter snowfall; precipitation on the plain is more uniformly distributed throughout the year (fig. 3). Cumulative departures from mean monthly precipitation at four stations on the eastern plain are shown in figure 4. General trends in precipitation for the eastern plain during the past 50 years include widespread drought from 1930 to 1935, 1952 to 1962, and

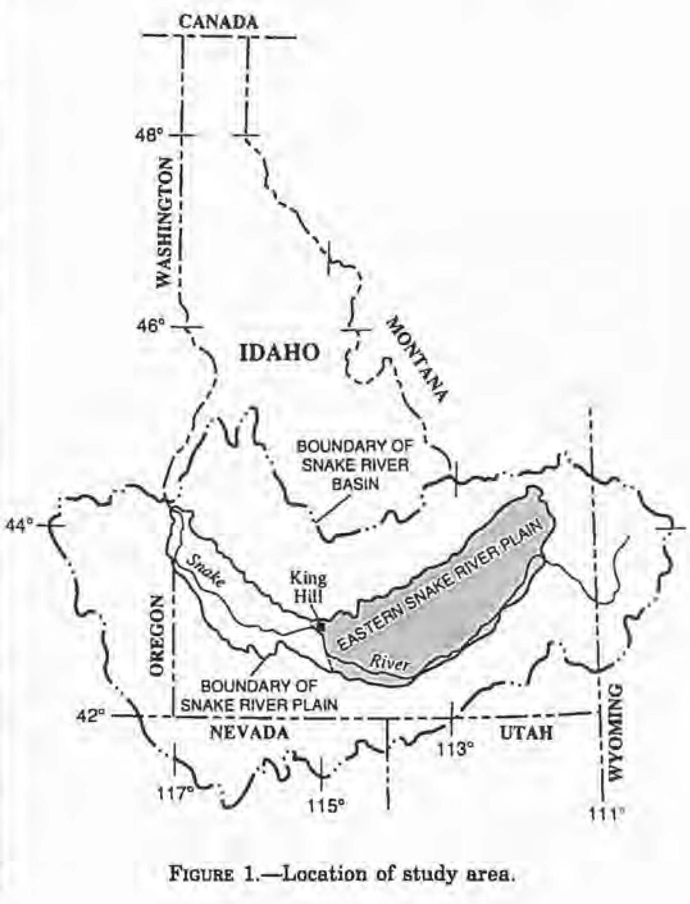


FIGURE 1.—Location of study area.

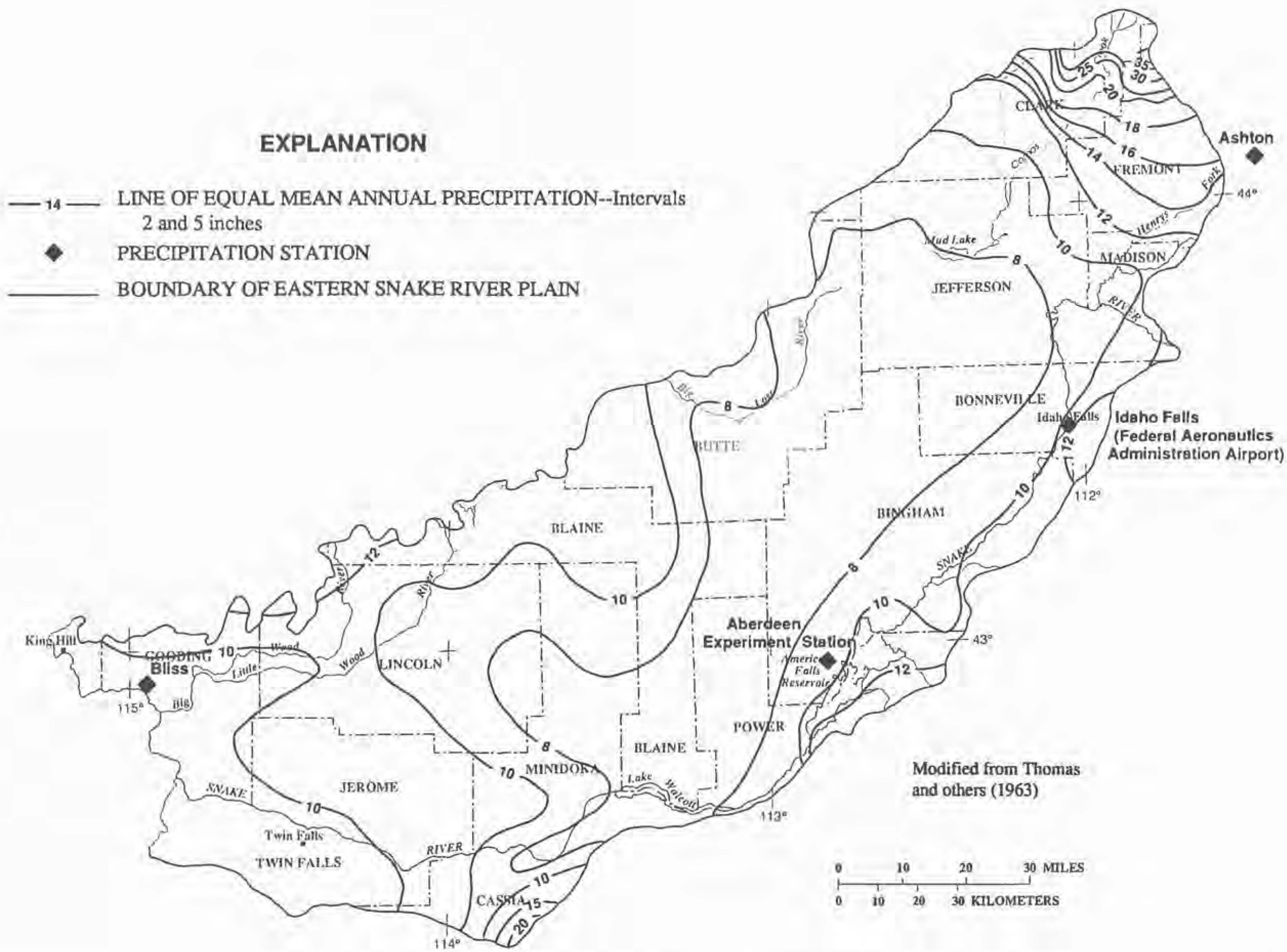


FIGURE 2.—Mean annual precipitation, 1930–57.

1977 to 1978, and wet periods from 1936 to 1942 and 1963 to 1976.

Natural vegetation on the plain is sparse because of the semiarid climate; sagebrush and bunchgrasses predominate. Most agricultural crops are irrigated, although dryland farming is moderately successful during wet years. Major crops are potatoes, small grains, sugar beets, beans, alfalfa seed, and hay. About 25 percent of the Nation's potatoes are produced on the eastern plain (Idaho Department of Agriculture, 1980, p. 5). Irrigated crop production on the eastern plain introduces about $600 million annually into Idaho's economy (Idaho Department of Agriculture, 1980, p. 24).

PREVIOUS INVESTIGATIONS

Numerous investigators have studied and reported on the geology and ground-water resources of the eastern Snake River Plain. Notable early studies were by Russell (1902) and Stearns and others (1938). In a quantitative hydrologic study, Mundorff and others (1964) used a flow-net analysis to estimate transmissivity. Skibitzke and da Costa (1962), Norvitch and others (1969), and Mantei (1974) used electric analog models to study the regional aquifer system in the eastern plain, whereas deSonneville (1974) and Newton (1978) used numerical models to

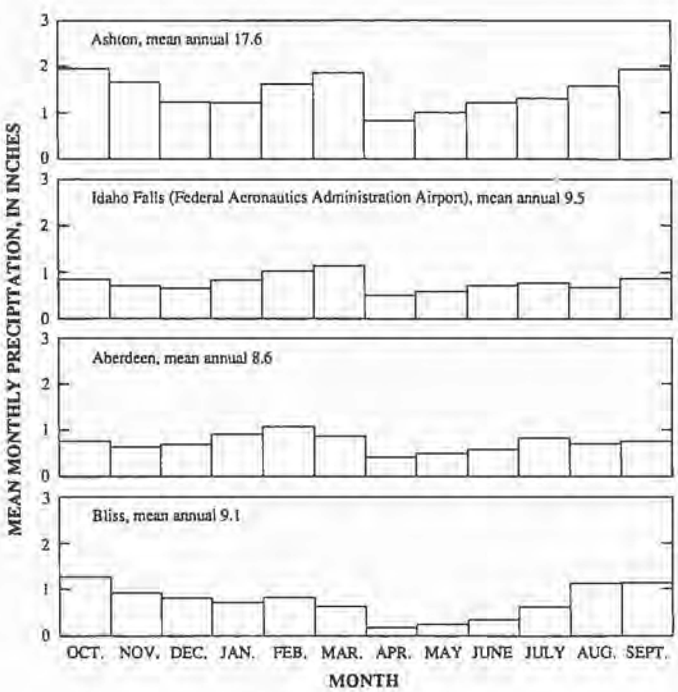


FIGURE 3.—Mean monthly distribution of precipitation, 1918–80. (Locations of precipitation stations shown in fig. 2.)

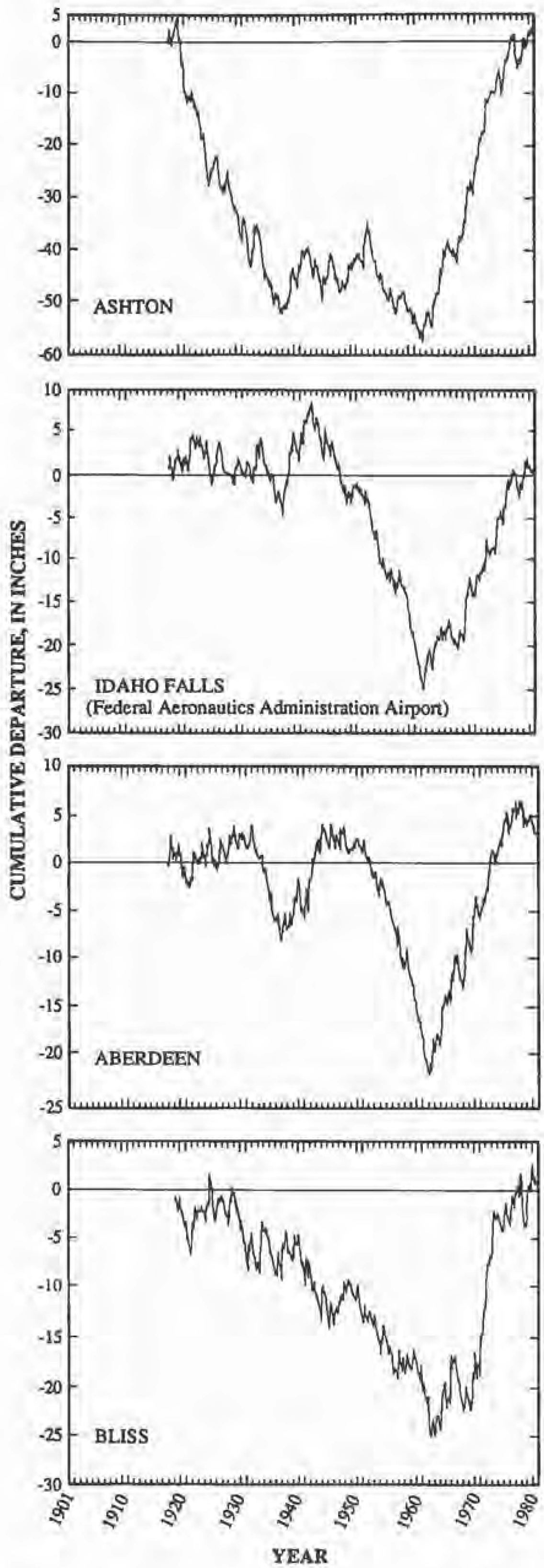


FIGURE 4.—Cumulative departure from mean monthly precipitation. (Locations of precipitation stations shown in fig. 2.)

study the regional system. Wytzes (1980) modeled the alluvial aquifer in the Henrys Fork and Rigby Fan area, and Johnson and others (1984) modeled the alluvial and basalt aquifers in the Mud Lake area. Solute transport of radioactive wastes at the Idaho National Engineering Laboratory was simulated by Robertson (1974, 1977); this simulation was updated by Lewis and Goldstein (1982).

WELL-NUMBERING SYSTEM

The well-numbering system (fig. 5) used by the U.S. Geological Survey in Idaho indicates the location of wells within the official rectangular subdivision of public lands. The first two numbers designate the township (north or south) and range (east or west) with reference to the Boise base line and Meridian. The third number designates the section and is followed by up to three letters, which indicate the ¼ section (160-acre tract), ¼-¼ section (40-acre tract), and ¼-¼-¼ section (10-acre tract). The last number is the order in which the well within the tract was inventoried.

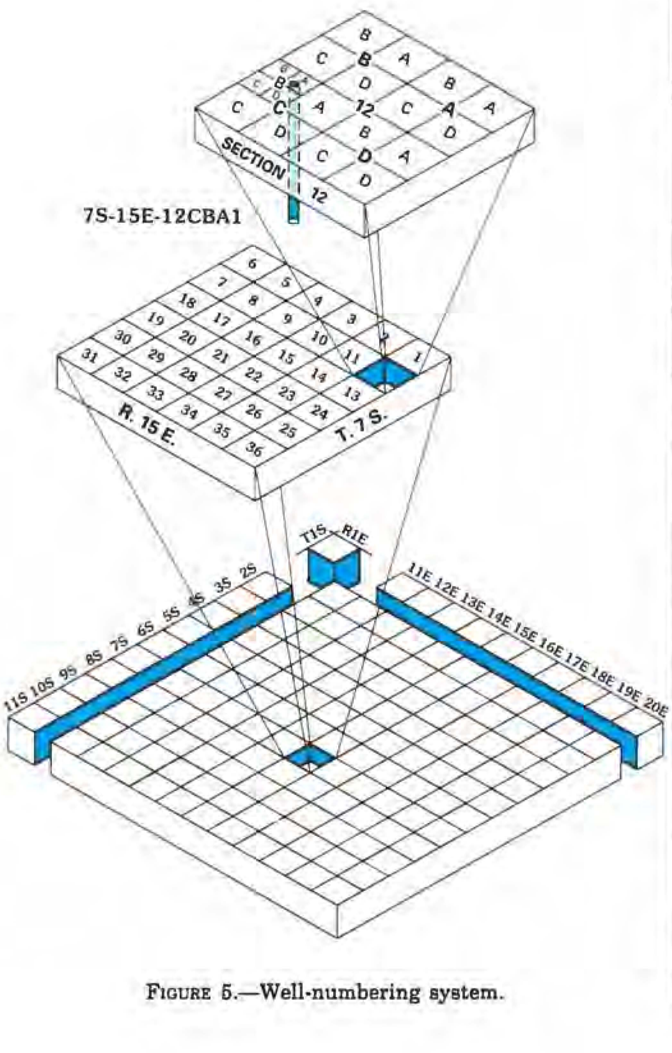


FIGURE 5.—Well-numbering system.

Quarter sections are lettered A, B, C, and D in counterclockwise order from the northeast quarter of each section. Within the quarter sections, 40-acre and 10-acre tracts are lettered in the same manner. For example, well 7S-15E-12CBA1 is in the NE¼NW¼SW¼ sec. 12, T. 7 S., R. 15 E., and was the first well inventoried in that tract.

GEOLOGY

The predominant rock type in the eastern Snake River Plain is Quaternary basalt of the Snake River Group (included in QTb on pl. 2). Basalt, interbedded with terrestrial and lacustrine sediments, along the margins of the plain fills a structural basin defined by faulting on the northwest and downwarping and faulting on the southeast (Whitehead, 1986). Electrical-resistivity soundings and other geophysical evidence indicate that aggregate basalt thickness may, in places, exceed several thousand feet (Whitehead, 1986). The structural basin was formed as the result of Cenozoic tectonic stresses and is a transition zone between the Basin and Range province to the southeast and the Northern Rocky Mountain province to the north and east.

Silicic volcanic rocks, including rhyolite, latite, and andesite, are present near the margins of the plain as thick flows of welded tuff, ash, and pumice. The northeastern end of the plain is delimited by rocks of the Yellowstone Group (mainly rhyolite). Idavada Volcanics are present southwest of the plain and may underlie the entire eastern plain. Underlying the Quaternary basalt in the southwestern part of the eastern plain are Tertiary sedimentary rocks of the Glenns Ferry Formation and Tertiary Banbury Basalt, both of which are part of the Idaho Group (pl. 2). Granitic rocks of the Idaho batholith, along with pre-Cretaceous sedimentary and metamorphic rocks, border the plain to the northwest. Adjacent to the plain on the southeast and perpendicular to its axis are several intermontane valleys and block-faulted mountain ranges.

Kuntz (1978, p. 9) noted that volcanism on the eastern plain was localized along rift zones (pl. 2). Rifts appear to be extensions of basin-and-range structures (faults) that are present northwest and southeast of the plain. Kuntz (1978, p. 13) indicated that faults are abundant owing to northeast-southwest extension along the axis of the eastern plain. In some places, this extension has caused open fissures at land surface.

Quaternary basalt of the Snake River Group was extruded from individual vents and from series of vents. A typical flow is 20 to 25 ft thick and 50 to 100 mi^2 in areal extent. Consequently, individual basalt flows cannot be correlated over great distances. Rubble and clinker zones usually form at the top of a basalt flow as cooling lava solidifies and then is broken by continued movement of underlying lava. Basalt vesicles are formed by the escape of entrapped gases. The centers of individual flows are typically less vesicular and more massive than flow tops. They are characterized by vertical fractures that, in places, form columnar basalt. Subsequent flows or fine-grained sedimentary deposits may partially fill fractures and vesicles.

Lava tubes are unique cooling features that form when a lava conduit drains, leaving a solid roof intact. Tubes may be continuous for a few feet to thousands of feet in length. Lava tubes have been penetrated in the subsurface, as evidenced by drill stems suddenly dropping as much as several tens of feet.

Sediments interbedded with basalt along the margins of the plain were deposited by the Snake River and tributary streams. In some areas, particularly in alluvial fans, sand and gravel predominate. In other areas, particularly where streams were dammed by basalt flows, fine-grained lacustrine sediments predominate. Soil cover on the plain is minimal over younger basalt and consists primarily of windblown material. Most agricultural soils are in areas of fluvial and lacustrine sediments near the margins of the eastern plain.

HYDROLOGY

SURFACE WATER

The eastern Snake River Plain is drained by the Snake River and its tributaries, which receive most ground-water discharge. The Snake River, which flows onto the plain near Heise, contributes about 49 percent of total tributary drainage-basin yield to the eastern plain. Another 23 percent of tributary drainage-basin yield is from Henrys Fork, and 10 percent is from northern tributaries. Most of the remaining yield from tributary drainage basins is from the Blackfoot, Portneuf, and Raft Rivers and Salmon Falls Creek (pl. 1).

The Snake River descends 2,524 ft from Heise (altitude 5,019 ft) to King Hill (altitude 2,495 ft), 307 river miles downstream, and is entrenched as much as 700 ft in the reach from Milner to King Hill.

Surface water is used extensively for irrigation on the eastern plain; more than 9 million acre-ft are diverted annually. Reservoir storage capacity in the Snake River basin above King Hill increased from about 1 million acre-ft in 1910 to about 5 million acre-ft in 1980 (Kjelstrom, 1986). Because of upstream storage, Snake River peak flows have been reduced, and more water is available during the irrigation season (May to October). Although flow in the Snake River is low during winter months, it is lowest in the summer owing to diversions for irrigation.

IRRIGATION

Surface water diverted for irrigation is presently the largest source of ground-water recharge in the eastern Snake River Plain. Consequently, changes in the amount of water used for irrigation must be known to model the ground-water-flow system. Use of surface water for irrigation increased rapidly after 1880. Decreed surface-water rights on the eastern plain increased from 204 ft^3/s in 1880 to 25,527 ft^3/s in 1905 (Idaho Department of Reclamation, 1921, pl. XXV). Earliest irrigation was concentrated along Henrys Fork, the upper Snake River, and the Big Wood and Little Wood Rivers (pl. 3); about 330,000 acres were irrigated in 1899. Irrigated lands expanded rapidly along the Snake River following construction of storage reservoirs and canals (table 1).

By 1929 about 1,540,000 acres were irrigated on the eastern plain, and by 1945 acreage had increased to about 1,770,000. Use of ground water for irrigation increased rapidly after 1945, and in some areas ground water replaced or supplemented surface water as a source. About 1,830,000 acres were irrigated in 1959—1,430,000 acres with surface water and 400,000 acres with ground water. Most of the land irrigated with ground water in 1959 was near land irrigated with surface water. However, some land shown as irrigated with surface water in 1945 is shown as irrigated with ground water in 1959, particularly near Mud Lake (pl. 3). In the Mud Lake area, both ground and surface water are used for irrigation. Some areas reported as irrigated with surface water in 1945 are actually irrigated with ground water that is transported to place of application via canals. Ground-water-irrigated acreage continued to increase and by 1966 totaled 640,000 acres; at this same time, surface-water-irrigated acreage totaled 1,560,000 acres. In 1979, a total of about 2,270,000 acres were irrigated: 1,230,000 with surface water, 930,000 with ground water, and 110,000 with com-

TABLE 1.—*Irrigated acreage, 1890–1945, and dates surface reservoir storage was added, Snake River drainage basin upstream from King Hill*

Year	Irrigated acres along Snake River from Heise to Neeley	Irrigated acres along Henrys Fork	Reservoir name (pl. 1)	Storage (acre-feet)
1890	47,000	2,000		
1900	221,000	30,000		
1905	299,000	49,000	Milner Lake	14,200
1906			Lake Walcott	107,240
1906			Jackson Lake	300,000
1909			Magic Reservoir	191,500
1910	372,000	58,000	Salmon River Canal Co. Reservoir	182,650
1910			Blackfoot Reservoir	413,000
1910			Jackson Lake (expanded)	380,000
1911			Oakley Reservoir	74,350
1915	423,000	62,000		
1916			Jackson Lake (expanded)	847,000
1919			Mackay Reservoir	44,370
1920	451,000	65,000		
1921			Mud Lake	61,660
1922			Henrys Lake	90,420
1923			Fish Creek Reservoir	13,500
1924			Grays Lake	40,000
1926			American Falls Reservoir	1,700,000
1930	471,000	68,000		
1935	462,000	56,000		
1938			Island Park Reservoir	127,300
1939			Grassy Lake	15,200
1939			Little Wood River Reservoir	29,960
1940	483,000	70,000		
1945	497,000	71,000		
1949			Lower Salmon Falls Reservoir	18,500
1951			Portneuf Reservoir	23,700
1956			Palisades Reservoir	1,400,000
1975			Ririe Lake	100,000
Total storage				5,494,550

bined surface and ground water. Lindholm and Goodell (1986), as part of the RASA study, used Landsat data to determine irrigated acreage on the Snake River Plain in 1980.

Canals shown in figure 6 supply the irrigated areas shown on plate 3. The Aberdeen–Springfield Canal was completed in 1900, the Twin Falls South Side Canal in 1907, the Twin Falls North Side Canal in 1911, and the Milner–Gooding Canal in 1930. Most diversions are by gravity feed and most canals are unlined; canal seepage losses range from 3 to 40 percent of diverted flow (Kjelstrom, 1986).

As surface-reservoir storage increased and water supply became more reliable, irrigation practices changed; in particular, winter diversions to maintain soil moisture were discontinued. Use of sprinklers to distribute water increased irrigation efficiency. About 20 percent of surface-water-supplied lands and 90

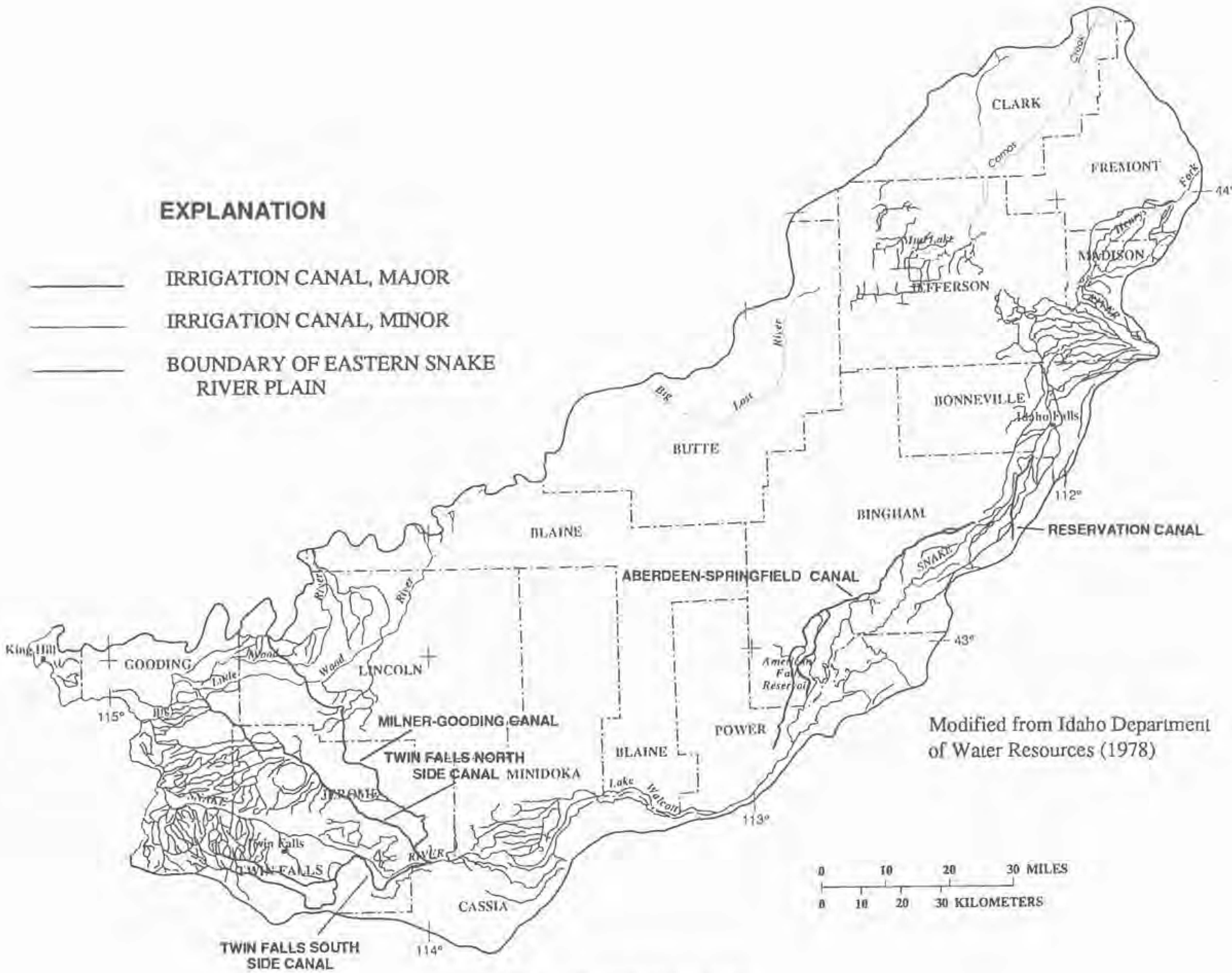


FIGURE 6.—Irrigation canals.

TABLE 2.—*Comparison of average annual crop consumptive irrigation requirements*

[Values in feet per year]

Location (pl. 1)	Simons (1953) Lowry–Johnson (1942)	Norvitch (1966)[1] Jensen–Criddle (1952)	Sutter and Corey (1970) Modified Blaney–Criddle (1950)	Moffatt (1980)[2] Jensen–Criddle (1952)	Value used in present study
	Reporting source and method				
Aberdeen	1.8	1.23	1.48	1.3	1.5
American Falls	1.8	1.18			1.5
Arco	1.3	1.15			1.5
Blackfoot	1.8	1.27	1.44		1.3
Carey		1.27			1.6
Dubois		1.25	1.34		1.3
Idaho Falls		1.18		1.3	1.3
Jerome	1.7	1.64	1.78	1.6	1.6
Mud Lake	1.6	1.16			1.3
Pocatello	1.8	1.31	1.68		1.5
Rupert	1.8	1.48	1.83	1.6	1.6
St. Anthony	1.2	.99	1.03		1.0
Shoshone	1.7	1.39	1.80		1.6

[1]U.S. Geological Survey, unpublished data on file in Boise, Idaho, office.
[2]U.S. Geological Survey, written commun.

percent of ground-water-supplied lands are now irrigated with sprinkler systems (Kjelstrom, 1986). A ditch-and-furrow system is used to distribute irrigation water in most other areas.

EVAPOTRANSPIRATION

Evapotranspiration (*ET*) rates used in this study are based on crop consumptive irrigation requirements determined by Sutter and Corey (1970). These rates represent plant-growth requirements minus growing-season precipitation.

A comparison of results using different methods to calculate *ET* rates at different locations in the Snake River drainage basin is shown in table 2. Simons (1953) used the Lowry–Johnson (1942) method, which is based on daily maximum air temperatures above the freezing point during the growing season. R.F. Norvitch (U.S. Geological Survey, unpubl. data on file in Boise, Idaho, office, 1966) and R.L. Moffatt (U.S. Geological Survey, written commun., 1980) used the Jensen and Criddle (1952) method, which is based on mean monthly temperature, length of growing season, monthly percent of annual daytime hours, amount of precipitation, and crop type.

Sutter and Corey (1970) calculated *ET* rates using a modified Blaney and Criddle method. Input was similar to that used in the Jensen and Criddle (1952) method with the addition of a crop-growth stage coefficient. Concurrent with the RASA study, Allen and Brockway (1983) adapted the FAO–Blaney–Criddle method to Idaho. The primary data requirement is mean air temperature. Although results obtained from the different methods are similar, rates calculated by the Lowry–Johnson method are consistently higher than those calculated by other methods. Differences between results range from about 20 to 40 percent, a reasonable range of error in *ET* estimates.

GROUND WATER

The occurrence and movement of ground water in the eastern Snake River Plain depend on the hydraulic characteristics of rocks that compose the geohydrologic framework and on the distribution and amount of aquifer recharge and discharge. A general description of hydrologic characteristics of major rock units in the eastern plain is presented in the explanation for plate 2.

Sand and gravel aquifers are located chiefly along the margins of the plain, in alluvial fans, and near present streams. Hydraulic conductivity of sand and gravel generally ranges from 3×10^{-5} to 3 ft/s (Freeze and Cherry, 1979, p. 29). Lacustrine silt and clay were deposited in lava-dammed streams, in local surface depressions, and by eolian processes. Because hydraulic conductivity of silt and clay is low (from 3×10^{-11} to 3×10^{-4} ft/s), vertical and horizontal flow is impeded. In some areas, such as Mud Lake and the Big Lost River valley, fine-grained sediments cause perched water zones and significant head changes with depth.

Largest well yields in the eastern plain are from basalt of Quaternary and late Tertiary age. Freeze and Cherry (1979, p. 29) indicated that the hydraulic conductivity of permeable basalt ranges from 3×10^{-6} to 3×10^{-1} ft/s. On the basis of transmissivity estimates from aquifer tests (Mundorff and others, 1964, p. 146, 147, 153–155), hydraulic conductivity of basalt in the eastern plain was estimated to range from 4×10^{-4} to 4×10^{-1} ft/s (G.F. Lindholm, U.S. Geological Survey, written commun., 1987). Horizontal water movement in basalt is primarily through rubble and clinker zones at the tops of flows and between successive flows. Water moves between flow tops along joints and at interfingering edges of rubble zones.

Davis (1969) indicated that hydraulic conductivity of an individual basalt flow is anisotropic; highest values are along the direction of original lava flow, parallel to rubble zones, lava tubes, and other cooling features. Individual flows in the central part of the eastern plain appear to have random directions, and anisotropy from the alignment of many basalt flows is unlikely. However, large-scale fractures in rift zones perpendicular to the axis of the plain (pl. 2) may cause anisotropy over broad areas.

Tertiary basalt generally yields less water to wells than younger basalt because individual flows are thicker and secondary minerals (calcite, clays, zeolites) fill many voids, reducing hydraulic conductivity. Tertiary basalt in a test hole drilled during this study is more massive and contains more secondary minerals than Quaternary and late Tertiary basalts (Whitehead and Lindholm, 1985).

The upper 2,445 ft in a 10,365-ft deep test hole at the INEL (Idaho National Engineering Laboratory, pl. 1) consists of basaltic lava flows and interbedded sediments of alluvial, lacustrine, and volcanic origin (Doherty and others, 1979). Basalt above a depth of 1,600 ft is typically tholeiitic olivine basalt of the Snake River Group. Lost circulation above 1,600 ft prevented return of drill cuttings and supported the hypothesis that the highly porous basalts are of the Snake River Group. Doherty and others (1979) noted that secondary mineralization is common in the basalt from 1,600 to 2,445 ft, and that porosity and hydraulic conductivity are reduced accordingly.

Fractured silicic volcanic rocks yield moderate amounts of water; if the rocks are tightly welded, well yields are low. In many locations, particularly in fault zones along the margins of the plain, volcanic rocks contain thermal water under confined conditions. In the INEL test hole, several thousand feet of silicic volcanic rocks below the basalt are hydrothermally altered, and nearly all fractures are sealed by secondary mineralization.

Consolidated sedimentary, metamorphic, and igneous rocks that compose mountains surrounding the eastern plain probably have low hydraulic conductivities, but their hydraulic properties are poorly known. Highest well yields are from fractures, faults, and weathered zones.

WELL YIELDS, SPECIFIC CAPACITIES, AND AQUIFER TESTS

Basalt of Quaternary and late Tertiary age that underlies the eastern Snake River Plain yields large quantities of water to wells. Data on 336 irrigation wells completed in basalt indicate that about 75 percent are pumped seasonally at 900 to 3,300 gal/min. Pumping drawdown below static water level in 68 percent of the wells was 20 ft or less. Maximum reported yield from a single well completed in basalt was about 7,250 gal/min. Along the margins of the plain where sedimentary rocks are interlayered with basalt, about 50 of 60 irrigation wells are pumped at 300 to 2,700 gal/min. Pumping drawdown in 45 percent of the wells completed in sedimentary rocks was 20 ft or less. Maximum reported yield from a single well was 3,000 gal/min. These data indicate that wells completed solely in basalt generally yield more water with less drawdown than wells completed in sedimentary rocks.

Median specific capacities (yield, in gallons per minute, per foot of drawdown) indicate the relative water-yielding capabilities of different aquifers. Specific-capacity data from 176 irrigation wells across the eastern plain are presented by county in table 3. Largest median specific capacities are from counties in the central part of the plain (Jefferson, Minidoka, Lincoln, Bonneville), where Quaternary basalts are thick and transmissivities high. Lowest median specific capacities are from counties along the southern margin of the plain (Cassia, Twin Falls), where Tertiary basalts and sediments predominate. The in-

TABLE 3.—*Specific capacities reported by drillers*

[Values in gallons per minute per foot of drawdown; QTs, Quaternary-Tertiary sediment; QTb, Quaternary-Tertiary basalt]

County	Aquifer	Number of wells	Mean	Standard deviation	Minimum	Maximum	Median
Bingham	QTs/QTb	16	940	1,710	27	6,400	120
Bonneville	QTb	5	340	280	33	680	360
Butte	QTb	10	710	1,220	3	3,600	130
Cassia	QTs/QTb	21	1,100	2,910	3	10,000	40
Gooding	QTb	6	1,500	2,920	9	7,450	340
Jefferson	QTb	29	2,120	2,540	18	9,000	950
Jerome	QTb	38	480	550	8	1,850	200
Lincoln	QTb	3	320	230	57	460	450
Minidoka	QTb	19	840	870	28	3,980	710
Power	QTs/QTb	21	180	220	1	750	80
Twin Falls	QTs/QTb	8	190	310	1	760	4

dicated maximums, minimums, and standard deviations show large areal variability in specific capacity because of differences in well construction, degree of development, and heterogeneity of the geologic framework. This heterogeneity is due to the discontinuity of highly productive zones of rubbly basalt and sand and gravel layers and indicates that aquifer properties change abruptly over short distances.

Transmissivity values from aquifer tests are indicative of relative areal differences in transmissivity but do not generally represent total aquifer transmissivity. Transmissivity and storage-coefficient data from 31 aquifer tests reported by Mundorff and others (1964, p. 146, 147, 153–155) and by Haskett and Hampton (1979, p. 26, 29) are presented in table 4. Transmissivities calculated from these tests typically represent local conditions around a partially penetrating well. Test data indicate that the upper 100 to 200 ft of the Snake River Plain aquifer has a range of transmissivity from less than 1.0 to 56 ft^2/s and an average unconfined storage coefficient of about 0.05. The data also show a large variation in transmissivity. For example, test data from Butte County in the central part of the plain show more than a hundredfold difference between low-and high-transmissivity values for basalt of the Snake River Group.

RECHARGE

Recharge to the eastern Snake River Plain ground-water system is from seepage of surface water used for irrigation, stream and canal losses, underflow from tributary drainage basins, and infiltration of precipitation. Recharge from each source was calculated separately. Pumped ground water in excess of crop consumptive irrigation requirements (*ET* minus growing-season precipitation) was assumed to return directly to the aquifer and therefore was not considered a source of recharge. The average recharge rate for each surface-water-irrigated area shown in figure 7 was determined using the equation

$$\text{Irrigation recharge (ft/yr)} = \frac{\text{Diversions–Return flows (acre-ft/yr)}}{\text{Area (acres)}} - ET \text{ (ft/yr)} \qquad (1)$$

Assuming that the ratio of recharge to surface water diverted is reasonably constant with time, recharge from surface-water irrigation in 1980 was about equal to the average annual recharge from 1928 to 1980 (fig. 8).

If diversion and return-flow data for irrigation districts were not available from watermaster reports (Idaho Department of Water Resources, 1980; Water Districts 37, 37M, 1980), they were calculated from U.S. Geological Survey records (1980) and from U.S. Bureau of Reclamation data (Roger Larson, written commun., 1981). To simplify the estimation of irrigation recharge, irrigation districts were grouped into areas similar to those used by Norvitch and others (1969), as shown in figure 7 and listed in appendix A.

Crop consumptive irrigation requirements (table 2) were adjusted for precipitation during the growing season to calculate recharge. Estimated recharge rates and volume of recharge for that part of each irrigation area (fig. 7) within the modeled area are shown in table 5. The model is discussed in the

TABLE 4.—*Transmissivities and storage coefficients determined by aquifer tests*

[Sources of data: Haskett and Hampton, 1979, p. 26, 29; Mundorff and others, 1964, p. 146, 147, 153–155. Aquifer: QTb, Quaternary–Tertiary basalt; QTs, Quaternary–Tertiary sediment; —, no data available]

County	Aquifer	Well location	Depth (feet)	Specific capacity (gallons per minute per foot of drawdown)	Transmissivity (feet squared per second)	Storage coefficient
Blaine	QTb	8S-26E- 3DCC2		185	7.4	—
Bonneville	QTb	3N-37E-12BD2	550	1,615	11.6	—
	QTb	1N-36E- 1CC1	218	4,570	23.2	0.075
Butte	QTb	6N-31E-13AC1	345	61	1.1	.01
	QTb	-13AC2	365	141	1.2	.03
	QTb	5N-31E-10CD1	—	—	.9	—
	QTb	4N-26E-32CB1	253	25	1.1	.024
	QTb	4N-30E- 7AD1	687	—	2.6	—
	QTb	4N-30E-30AA1	546	147	2.3	—
	QTb	-30AA2	—	—	1.7	—
	QTb	-30AD1	529	—	5.7	—
	QTb	3N-29E-14AC1	596	2,175	21.7	.02
	QTb	-14AD1	—	—	27.9	.06
	QTb	3N-29E-24AD1	605	—	5.1	.06
	QTs/QTb	3N-30E-34BA1	653	18	.2	—
	QTs/QTb	2N-29E- 1DB1	681	15	.2	—
Cassia	QTb	10S-21E-34DD1	473	860	9.7	.22
Fremont	QTb	7N-39E-16DBB4	—	1,740	55.7	—
Gooding	QTb	8S-15E-33CC1	107	—	15.5	.045
Jefferson	QTb	8N-34E-11DC1	116	2,060	12.4	.055
	QTb	7N-34E-24AA1	106	2,500	7.1	.10
	QTb	6N-35E-26CC1	300	—	7.0	.034
Jerome	QTb	7S-19E-19AA1	280	2,150	13.3	—
	QTb	8S-19E- 5DA1	329	88	7.7	—
	QTb	9S-19E-25BB1	208	1,470	4.3	—
	QTb	10S-21E-26AAA2	—	7	1.2	—
Lincoln	QTb	5S-17E-26AC1	254	1,610	5.6	—
	QTb	6S-18E- 7BC1	224	457	5.3	—
Madison	QTb	7N-38E-23DB1	236	1,130	18.6	.000017
	QTs/QTb	6N-38E-25ACB1	685	1,305	23.2	—
Minidoka	QTb	8S-24E- 8AD2	258	695	13.5	.014

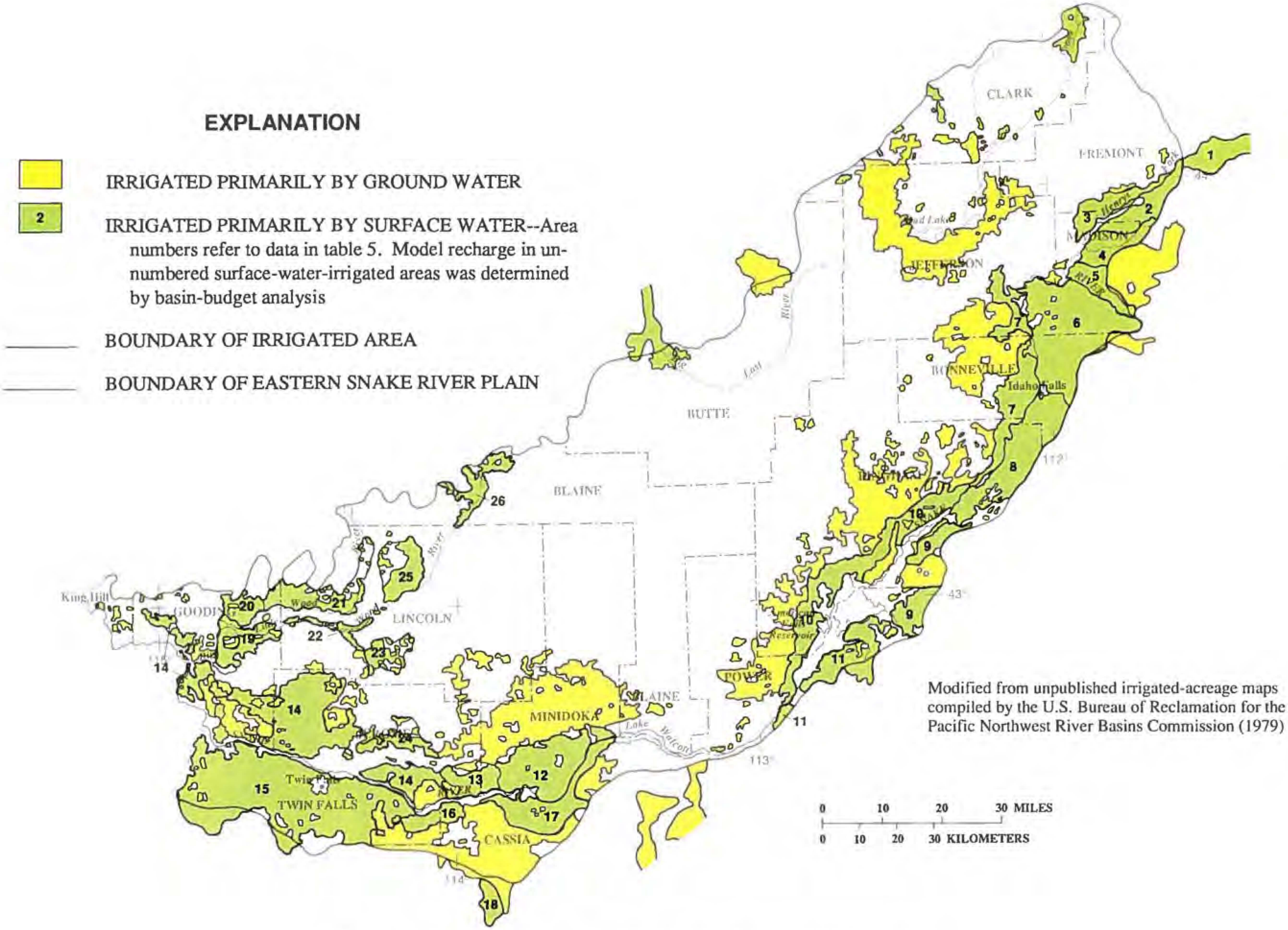


FIGURE 7.—Irrigated lands, 1979, and numbered surface-water-irrigated areas.

section "Ground-Water-Flow Modeling." Total volume of ground-water recharge in the modeled area from surface-water irrigation in 1980 was estimated to be about 4,800,000 acre-ft.

Irrigation-diversion data from the Idaho Department of Water Resources (written commun., 1981) were used to calculate 5-year average ground-water-recharge rates from 1928 to 1980 (appendix B). These records are the basis for recharge rates listed in table 6; calculations are shown in appendix B. Recharge rates for most irrigated areas changed slightly over the period of record; some fluctuations were noted during dry and wet periods. However, recharge rates calculated for areas 1 and 16 changed substantially owing to large changes in irrigated area, as shown on the irrigated-area maps (pl. 3). Mapped differences in irrigated area are assumed to be due largely to actual changes in irrigated acreage but may, in part, reflect mapping errors. Estimation errors of local scale are presumed to have minimal effect on regional analysis of ground-water hydrology.

Rates for total recharge from infiltration of surface water used for irrigation during 5-year periods from 1891 to 1980 are listed in table 7. Calculations for the period 1891–1925 were based on the average recharge rates shown in table 6, along with 1899 and 1929 irrigated-acreage maps in various combinations, as listed in table 7. Variations in irrigated acreage from 1891 to 1920 were estimated on the basis of information in table 1, which indicates that about 50 percent of presently irrigated land above Neeley was put into production between 1890 and 1900. Opening dates of major canals, such as the Twin Falls North Side and South Side Canals, also were considered in estimating other increases in irrigated acreage. Return-flow estimates were based on data collected by the U.S. Bureau of Reclamation in 1979–80 and by the U.S. Geological Survey in 1980 (Kjelstrom, 1986). Few return-flow data are available for past years.

Stream and canal losses also provide significant amounts of recharge. Snake River losses to ground water totaled about 700,000 acre-ft in 1980, and gains from ground water totaled about 7,100,000 acre-ft (Kjelstrom, 1986). Snake River reach losses and gains in 1980 are listed in table 8; average annual losses and gains for 5-year periods from 1912 to 1980 are listed in table 9. Losses decreased in the reach from Heise to near Blackfoot from 1912 to 1980 as a result of a rise in ground-water levels under surface-water-irrigated lands near the river. Raised ground-water levels reduced head differences between the river and the aquifer and subsequently reduced river losses to the aquifer. Annual ground-water discharge to the Snake River between Blackfoot and Neeley was consistently about 1,800,000 acre-ft, but annual discharge between Neeley and Milner fluctuated from about 90,000 to 480,000 acre-ft. Variations in discharge between Neeley and Milner probably result from wet and dry climatic cycles and from error in the water-budget analysis. Average annual ground-water discharge to the Snake River between Milner and King Hill increased from about 3,800,000 acre-ft during 1912–15 to a maximum of about 5,300,000 acre-ft during 1951–55 in response to increased diversions of surface water for irrigation. Since 1955, ground-water discharge to the reach has declined to about 4,800,000 acre-ft/yr.

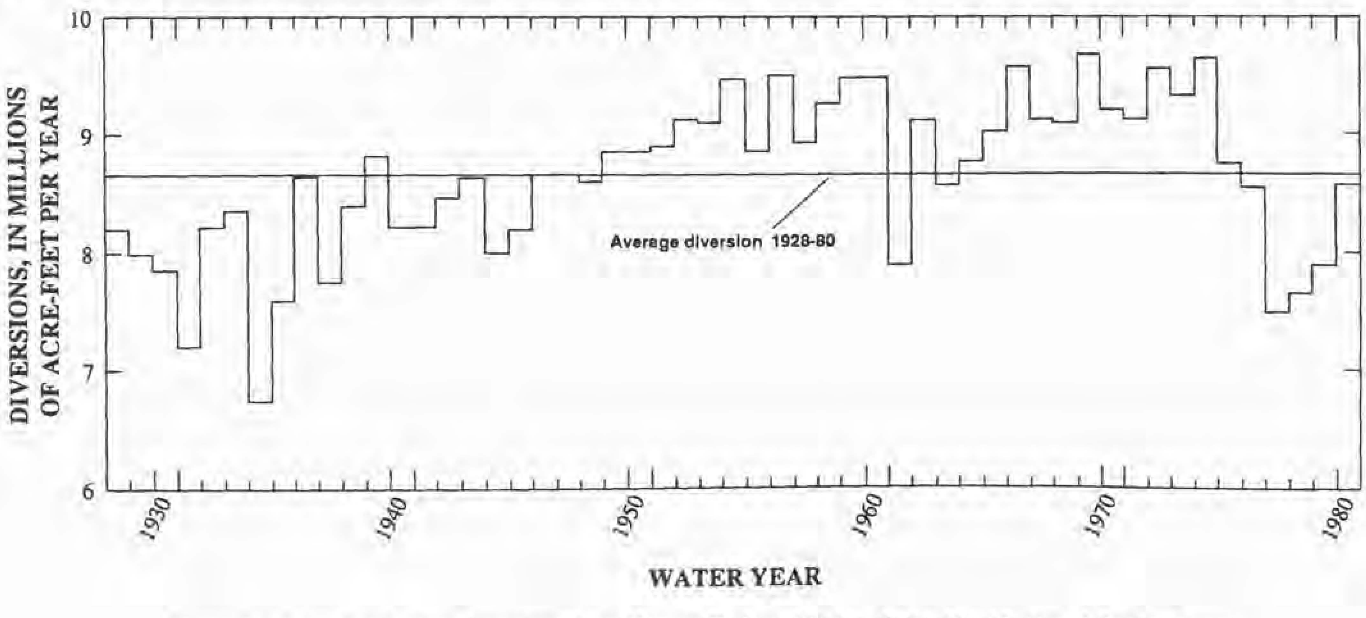


FIGURE 8.—Annual irrigation diversions from Henrys Fork and Teton, Falls, Blackfoot, and Snake Rivers.

TABLE 5.—*Estimated recharge from surface-water irrigation, water year 1980*

[Values are rounded]

Area (fig. 7)	Diversion–return flow (acre-feet per year)	Area (acres)	Evapotranspiration (feet per year)	Recharge (feet per year)	Area in model[1] (acres)	Recharge to modeled area (acre-feet per year)[1]
1	38,800	31,300	1.0	0.24	1,100	300
2	154,300	21,800	1.1	5.97	19,600	117,300
3	379,100	26,300	1.2	13.24	26,300	347,600
4	246,600	32,800	1.3	6.22	32,200	200,500
5	128,700	30,500	1.3	2.92	28,400	82,800
6	1,388,600	146,200	1.3	8.20	137,500	1,127,900
7	256,400	62,700	1.3	2.79	62,700	174,800
8	527,900	82,800	1.3	5.08	78,200	397,100
9	227,300	29,300	1.5	6.25	29,300	183,400
10	487,900	97,100	1.5	3.53	96,600	341,100
11	73,900	49,300	1.5	0	22,000	0
12	338,500	82,000	1.6	2.53	82,000	207,600
13	48,800	17,600	1.6	1.18	17,600	20,700
14	1,022,100	179,600	1.6	4.09	179,600	734,400
15	571,900	258,500	1.6	.61	229,300	139,900
16	60,600	11,200	1.6	3.80	11,200	42,600
17	245,800	49,300	1.6	3.38	48,100	162,500
18	44,900	12,000	1.6	2.16	0	0
19	67,100	18,000	1.6	2.13	18,000	38,300
20	62,600	17,000	1.6	2.08	17,000	35,400
21	226,100	29,000	1.6	6.19	29,000	179,700
22	106,800	6,200	1.6	15.50	6,200	96,800
23	69,200	15,900	1.6	2.74	15,900	43,600
24	36,000	11,600	1.6	1.50	11,600	17,400
25	94,700	27,500	1.6	1.85	27,500	50,900
26	129,200	17,100	1.6	5.96	17,100	101,800
Totals	7,033,800	1,362,600			1,244,000	4,844,400

[1]See section "Ground-Water Flow Modeling."

Most canal losses were added to the total recharge from each irrigation area. However, because the Milner–Gooding, Aberdeen–Springfield, and Reservation Canals lose water by seepage before reaching points of delivery, they were treated separately as distributed losses and are listed in table 10. Tributary streams listed in table 10 also were treated as distributed losses because they lose all their flow to seepage or to *ET* on the plain. Losses from tributary streams that reach the Snake River were included with irrigation recharge in the areas supplied by those streams.

Kjelstrom (1986) used basin-yield equations to calculate average annual underflow rates from tributary drainage basins (table 11). Equations incorporate drainage area, mean annual precipitation, and percentage of forest cover as independent variables. Coefficients for independent variables were determined from a regression analysis by using basins for which streamflow records were available and from which, on the basis of geologic conditions, underflow was assumed to be relatively small (Kjelstrom, 1986). Underflow estimated using rates determined from basin-yield equations is about 8 percent of the average water yield from tributary drainage basins.

Average annual recharge from infiltration of precipitation on the plain was assumed to vary according to the amount of precipitation, soil thickness, and

TABLE 6.—*Recharge from Henrys Fork and Snake, Big Wood, and Little Wood River diversions, 1928–80*

[Values in feet per year; —, no data available]

Area (fig. 7)	Year 1928–30	1931–35	1936–40	1941–45	1946–50	1951–55	1956–60	1961–65	1966–70	1971–75	1976–80	1928–80 average
1	4.33	2.45	4.29	4.76	6.43	0.44	0.55	0.71	0.80	([1])	([1])	2.75
2	3.75	2.90	3.61	3.77	4.00	4.00	4.64	4.38	4.68	5.19	4.55	4.13
3	8.05	7.55	8.19	7.65	7.92	12.39	13.12	8.05	7.93	10.60	8.96	9.13
4	2.80	2.53	3.11	3.10	3.19	4.31	4.63	2.04	2.14	6.21	3.77	3.44
5	4.06	1.95	2.41	1.40	1.63	3.30	3.78	.85	1.20	3.79	3.18	2.50
6	8.63	7.66	8.22	7.91	8.21	9.61	10.03	8.05	8.76	11.21	8.42	8.79
7	2.64	2.05	2.58	2.54	3.12	3.06	3.31	2.34	2.42	2.93	2.48	2.68
8	3.80	3.56	4.27	4.03	4.40	4.49	4.69	4.02	4.30	4.38	3.96	4.17
9	—	—	—	—	—	—	—	—	—	—	3.86	3.86
10	1.76	1.43	1.64	1.51	1.71	2.45	2.61	2.02	2.33	4.20	3.57	2.29
11	—	—	—	—	—	—	.97	.38	.97	.66	.78	.41
12	4.11	2.78	3.36	2.70	2.76	3.32	3.43	2.29	2.77	2.95	2.41	2.99
13	—	—	—	—	—	—	.15	0	0	.53	.56	.25
14	2.34	1.73	2.11	2.10	2.28	4.83	4.71	3.02	3.63	4.96	3.79	3.23
15	1.31	1.51	1.54	1.84	1.92	1.76	1.63	1.27	1.44	1.87	1.72	1.62
16	3.25	1.27	1.59	.79	1.18	12.06	13.56	17.88	20.54	1.58	1.60	6.85
17	1.62	1.75	2.14	2.12	2.50	3.30	3.34	3.33	3.68	4.31	3.39	2.86
18	.57	.11	.53	1.45	1.20	.64	.41	3.43	2.45	—	—	1.20
19	—	—	1.14	.74	.97	1.43	1.69	1.52	1.35	.86	.66	1.15
20	—	0	.65	.34	.50	1.82	1.77	1.29	1.43	2.10	1.62	1.15
21	—	—	5.99	3.14	3.65	6.44	7.79	10.86	12.24	7.61	4.67	6.93
22	—	—	12.42	25.86	22.59	6.83	5.94	19.54	12.87	10.99	8.14	14.10
23	—	1.33	2.10	3.35	3.74	1.56	1.32	2.73	2.28	3.48	2.34	2.42
24	—	—	—	—	—	—	—	.28	.22	.93	.37	.45
25	—	1.39	2.07	2.26	2.75	2.14	2.05	2.67	3.64	2.26	1.26	2.25
26	1.50	2.08	2.11	1.96	2.92	2.30	1.34	2.21	2.73	2.90	2.77	2.33

[1]No area inside study boundary.

infiltration capacity of the soil cover. Most recharge is snowmelt that infiltrates during winter and spring months when evapotranspiration rates are low. Recharge was varied using October to March precipitation records from stations at Aberdeen, Ashton, Bliss, and Idaho Falls (fig. 2). Recharge from precipitation, shown in figure 9, was calculated by subdividing the eastern Snake River Plain into six areas (table 12), which differ in soil type (fig. 10 and appendix C) and amount of mean annual precipitation (fig. 2). Rates of recharge from precipitation were modified from rates used by Mundorff and others (1964, p. 184) by taking into account soil texture and soil depth to estimate recharge from precipitation (table 12). Total average annual recharge to the ground-water system from precipitation was estimated to be about 700,000 acre-ft.

Stephenson and Zuzel (1981) used a similar approach to estimate recharge from precipitation during a study of ground-water recharge in a small basin underlain by basalt in southwestern Idaho. They determined that ground water was recharged by infiltration in areas of low-relief rubbly basalt outcrops and shallow soils, and in bedrock channels during runoff and channel flow. Recharge took place after 0.8 to 1.2 in. of rain fell with a 24-hour period or after higher intensity cloudbursts. Stephenson and Zuzel (1981) also determined that the time from the end of precipitation to the ground-water-level peak depends only on soil depth.

Variations in tributary-stream losses, underflow from tributary drainage basins, and recharge from precipitation during 5-year intervals between 1911

TABLE 7.—*Total recharge from surface-water irrigation, 1891–1980*

Years	Irrigated acreage (pl. 3)	Recharge (table 6)	Total recharge (acre-feet per year)
1891–95	0.50 × 1899 acreage	Average 1928–80	730,000
1896–1900	1.00 × 1899 acreage	...... do.	1,450,000
1901–05	0.65 × 1929 acreage, areas 1–10; 1.00 × 1899 acreage, areas 11–26	...... do.	2,220,000
1906–10	0.80 × 1929 acreage, areas 1–10; 1.00 × 1899 acreage, areas 11–26	...... do.	2,660,000
1911–15	0.90 × 1929 acreage, areas 1–10; average of 1899 and 1929 for areas 11–26	...... do.	3,890,000
1916–20	0.95 × 1929 acreage, areas 1–10; average of 1899 and 1929 for areas 11–26	...... do.	4,040,000
1921–25	1929	...... do.	5,130,000
1926–30	1929	1928–30	4,610,000
1931–35	Average of 1929 and 1945 ...	1931–35	4,170,000
1936–40	Average of 1929 and 1945 ...	1936–40	4,650,000
1941–45	1945	1941–45	4,620,000
1946–50	1945	1946–50	4,960,000
1951–55	1959	1951–55	5,400,000
1956–60	1959	1956–60	5,560,000
1961–65	1966	1961–65	4,850,000
1966–70	1966	1966–70	5,290,000
1971–75	1979	1971–75	5,750,000
1976–80	1979	1976–80	4,600,000

and 1980 are listed in table 13. Flows in streams and underflow crossing the northern boundary of the eastern Snake River Plain were estimated using correlations with the long-term streamflow hydrograph of the Big Lost River below Mackay Reservoir (L.C. Kjelstrom, U.S. Geological Survey, written commun., 1983). Underflow from tributary drainage basins along the southern boundary of the plain was estimated using correlations with the long-term record of the Portneuf River at Pocatello.

DISCHARGE

SEEPS AND SPRINGS

Ground-water discharge from the eastern Snake River Plain aquifer system is largely seepage and spring flow to the Snake River in the reaches from Blackfoot to Neeley and from Milner to King Hill. During the 1980 water year, about 4,700,000 acre-ft of ground water were discharged to the Snake River along the reach from Milner to King Hill (Kjelstrom, 1986). This amount is about 70 percent of gaged flow at King Hill in 1980. Most springs discharge from basalt of the Snake River Group along the north side of the river. H.R. Covington (U.S. Geological Survey, written commun., 1983) determined that the altitudes of north-side springs are controlled by several factors: (1) altitude of the contact between relatively impermeable Banbury Basalt and basalt of the Snake River Group, (2) location of lake clays, and (3) location of relatively impermeable Idaho Group (Glenns Ferry Formation) sedimentary rocks.

Most major springs along the Snake River from Milner to King Hill discharge from pillow lavas and basaltic sands. H.R. Covington (U.S. Geological Survey, written commun., 1983) described the pillow lavas as basalt that was deposited in a lake behind a lava dam in an ancestral Snake River canyon. As lava continued to flow into the canyon, a sequence of dense lavas was deposited downstream from the dam, whereas pillow lavas were deposited upstream. Pillow

TABLE 8.—*Snake River losses to and gains from ground water, water year 1980*

[From Kjelstrom, 1986]

Reach (gaging-station locations shown on pl. 1)	Loss (–) or gain (cubic feet per second)	Loss (–) or gain (acre-feet per year)
Heise to Lorenzo	–145	–105,000
Lorenzo to Lewisville	289	209,000
Lewisville to Shelley	–379	–275,000
Shelley to at Blackfoot	–153	–111,000
At Blackfoot to near Blackfoot	–270	–196,000
Near Blackfoot to Neeley	2,620	1,902,000
Neeley to Minidoka	179	130,000
Minidoka to Milner	132	96,000
Milner to Kimberly (north side)	30	21,000
Milner to Kimberly (south side)	266	193,000
Kimberly to Buhl (north side)	1,112	807,000
Kimberly to Buhl (south side)	110	80,000
Buhl to Hagerman (north side)	3,456	2,509,000
Buhl to Hagerman (south side)	150	109,000
Hagerman to King Hill	1,412	1,025,000
Total loss	–947	–687,000
Total gain	9,756	7,081,000

TABLE 9.—*Snake River losses to and gains from ground water, 1912–80*

Years	Average annual loss (–) or gain between gaging stations (acre-feet per year): Heise to near Blackfoot	Near Blackfoot to Neeley	Neeley to Milner	Milner to King Hill
1912–15	–730,000	1,860,000	130,000	3,760,000
1916–20	–740,000	1,810,000	220,000	4,020,000
1921–25	–670,000	1,850,000	100,000	4,280,000
1926–30	–560,000	1,830,000	140,000	4,560,000
1931–35	–810,000	1,860,000	90,000	4,550,000
1936–40	–630,000	1,800,000	180,000	4,790,000
1941–45	–550,000	1,850,000	210,000	5,040,000
1946–50	–400,000	1,850,000	220,000	5,170,000
1951–55	–330,000	1,880,000	200,000	5,290,000
1956–60	–340,000	1,830,000	200,000	5,130,000
1961–65	–460,000	1,850,000	140,000	4,850,000
1966–70	–360,000	1,810,000	230,000	4,980,000
1971–75	–190,000	1,770,000	480,000	4,960,000
1976–80	–430,000	1,930,000	280,000	4,810,000

TABLE 10.—*Average annual tributary-stream and canal losses to the ground-water system*

[Streams shown on pl. 1; canals shown in fig. 6; NA, not applicable]

Name	Loss (acre-feet per year)	Consumptive water use upstream from the boundary of the plain[1] (acre-feet per year)
Big Lost River	51,000	35,000
Little Lost River	12,000	16,000
Medicine Lodge Creek	30,000	4,000
Beaver Creek	31,000	1,000
Camas Creek	63,000	9,000
Milner–Gooding Canal	97,000	NA
Aberdeen–Springfield Canal	95,000	NA
Reservation Canal	11,000	NA
Totals	390,000	65,000

[1]Additional streamflow available for recharge before irrigation began.

lavas generally are unsorted, coarse grained, poorly indurated, and have extremely high porosity and hydraulic conductivity. Highly permeable pillow lavas and the interconnection of ancestral canyons make the basaltic aquifer along the river reach from Kimberly to Bliss highly transmissive.

In many places, the top of the Banbury Basalt defines the lower limit of major spring emergence along the present canyon between Twin Falls and Bliss. However, not all springs discharge from in situ Quaternary basalt. Many springs discharge from talus aprons at various altitudes above the canyon floor; a few appear to discharge from older Banbury Basalt. From test drilling and examination of roadcuts along the canyon, Whitehead and Lindholm (1985, p. 17) suggested that fine-grained sediments of the Idaho Group control some spring-vent altitudes. At some locations, springs discharge from coarse-grained flood debris on the canyon floor or directly into the river.

Ground-water discharge (mainly spring flow) from the north side of the Snake River increased dramatically after 1911 as a result of surface-water irrigation in Gooding, Jerome, and Lincoln Counties (fig. 11). Spring flow continued to increase until about 1950–55 and peaked in 1951 at about 6,800 ft^3/s. Spring flow generally has declined since the 1950's and was about 6,000 ft^3/s in 1980. Generally, individual spring discharges are lowest in April before irrigation begins and highest in October just after irrigation ends. Both short-term and long-term fluctuations in spring discharges are strongly and rapidly responsive to changes in recharge from irrigation.

Ground-water discharge from springs and seeps on the south side of the Snake River from Milner to

TABLE 11.—*Estimated underflow from tributary drainage basins*

Name (pl. 1)	Underflow (cubic feet per second)	Underflow (acre-feet per year)
Camas Creek	215	155,000
Beaver Creek	85	62,000
Medicine Lodge Creek	13	9,000
Warm Springs and Deep Creeks	42	30,000
Birch Creek	108	78,000
Little Lost River	214	155,000
Big Lost River	408	295,000
Fish Creek	8	6,000
Little Wood River	25	18,000
Silver Creek	73	53,000
Big Wood River	14	10,000
Thorn Creek	8	6,000
Clover Creek	14	10,000
Salmon Falls Creek	138	100,000
Cottonwood, Rock, and Dry Creeks	20	14,000
Goose Creek	39	28,000
Raft River	116	84,000
Rockland Valley (Rock Creek)	70	51,000
Bannock Creek	30	22,000
Portneuf River	87	63,000
Lincoln and Ross Fork Creeks	6	4,000
Blackfoot River	18	13,000
Willow Creek	40	29,000
Snake River	10	7,000
Rexburg Bench	26	19,000
Teton River and Henrys Fork	4	3,000
Big Bend Ridge area	153	111,000
Totals	1,984	1,435,000

Hagerman generally is small, about 500 ft³/s in water year 1980 (table 8). Half of the total was along the Milner to Kimberly reach; from Kimberly to Buhl, an unmeasured amount of ground water is discharged to field drains and tunnels. The smaller amount of spring flow from the south side of the Snake River is likely due to a reduction in hydraulic conductivity of basalt. Mundorff and others (1964, p. 73) reported that water levels rose as much as 200 ft in the Twin Falls area after irrigation began in 1907. Many fields became waterlogged and drains were constructed. These observations indicate that the transmissivity south of the Snake River is generally lower than that of the north-side basalt aquifer.

Directly north and east of American Falls Reservoir, major springs and seeps discharge along the Snake and Portneuf Rivers. Most discharge is from the sand and gravel aquifer that underlies the Snake River flood plain from Blackfoot to American Falls Reservoir. Details of geology in the immediate springs area are poorly known, as there are no deep drill holes on the flood plain. East of the flood plain, several hundred feet of sand and gravel overlie basalt (pl. 2, section D–D'); Quaternary basalt predominates to the west. In the immediate vicinity of American Falls Reservoir, lacustrine sediments confine water that discharges as springs where confining beds are absent, such as along the Portneuf River. Fifty to eighty feet of flood deposits from the Pleistocene breakout of Lake Bonneville (Malde, 1968, p. 21) overlie the previously mentioned sand and gravel deposits in part of the area.

Since 1912, mean annual ground-water discharge to the Snake River from springs between Blackfoot and Neeley has been consistent, averaging about 2,500 ft³/s (fig. 12). River gains from ground water measured in 1902, 1905, and 1908 of less than 2,000 ft³/s indicate that some of the measured discharge in 1912 may be attributed to recharge from irrigation. Spring discharge apparently was not affected by the filling of American Falls Reservoir in 1926.

The Snake River from Lorenzo to Lewisville (pl. 1) gains ground water during the irrigation season but loses ground water the rest of the year. Gains are likely from sand and gravel that comprise the alluvial fan around Rigby. Aquifer recharge from surface-water irrigation is as much as 8 ft/yr. During the 1980 water year, the Lorenzo to Lewisville reach gained about 209,000 acre-ft of water (Kjelstrom, 1986).

GROUND-WATER PUMPAGE

Pumpage of ground water for irrigation increased rapidly after 1945. By 1959, about 400,000 acres were irrigated with ground water; by 1966, 640,000 acres; and by 1979, 930,000 acres, or 40 percent of the irrigated lands on the eastern Snake River Plain. Amounts of pumpage were estimated from acreages shown on plate 3, and consumptive irrigation requirements are given in table 2. About 630,000 acre-ft of water were pumped for irrigation in 1959, 990,000 acre-ft in 1966, and 1,430,000 acre-ft in 1979. Water pumped in excess of consumptive irrigation requirements was assumed to return to the aquifer.

Ground-water pumpage for irrigation in 1980 was estimated from electrical power-consumption data (Bigelow and others, 1986). An estimated 1,760,000

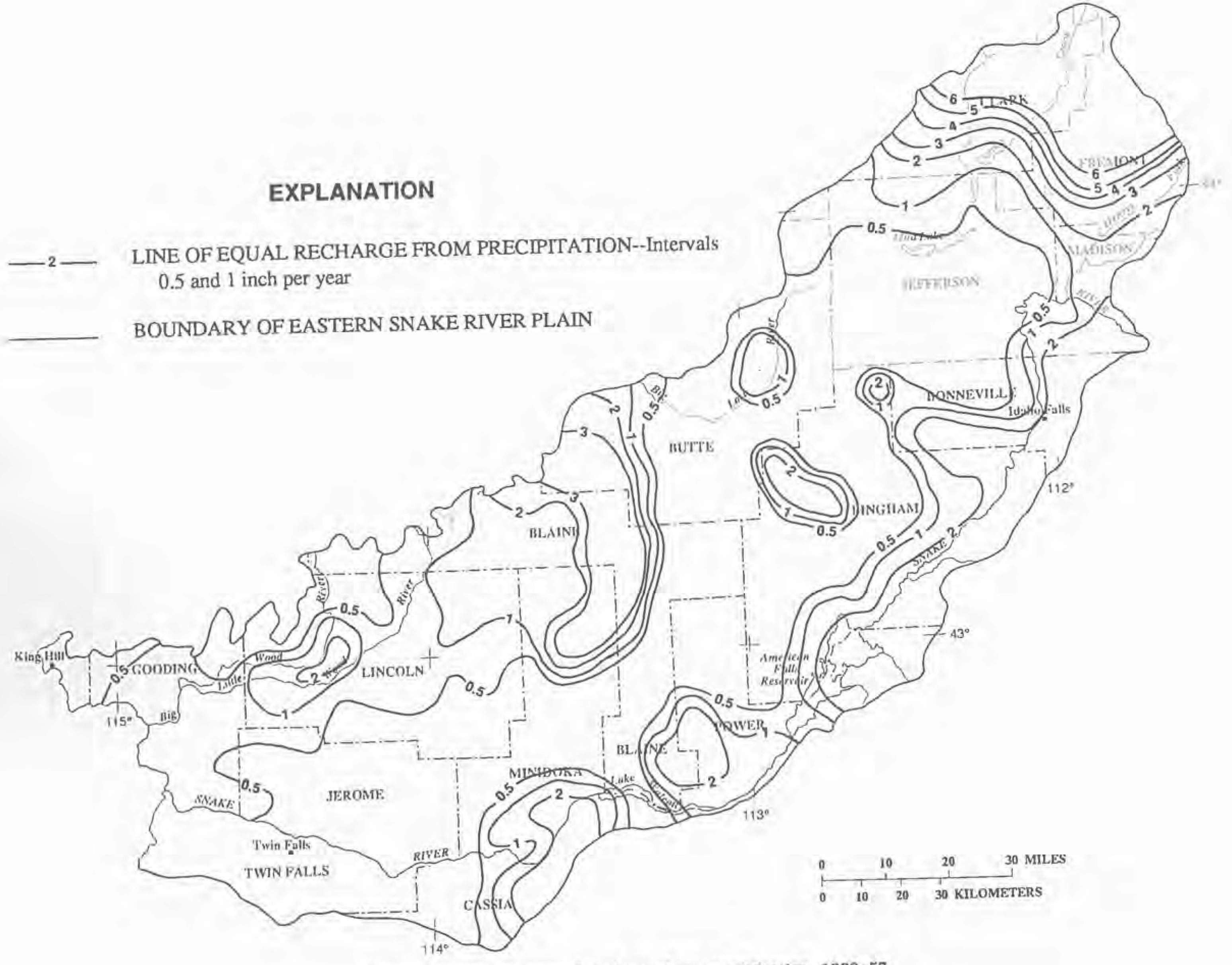


FIGURE 9.—Average annual recharge from precipitation, 1930–57.

acre-ft of ground water were withdrawn from about 4,000 wells to irrigate about 930,000 acres. Some pumped water was returned to the aquifer from canal loss and field seepage. Therefore, pumpage estimated from power-consumption data was compared with estimated consumptive irrigation requirements, and the smaller of the two estimates was used to determine net ground-water withdrawal. Net ground-water withdrawal in 1980 was estimated to be about 1,140,000 acre-ft, or about two-thirds of total pumpage estimated from power-consumption data.

Pumpage for other uses in 1969 was estimated by Young and Harenberg (1971, p. 22–24) to be about 34,000 acre-ft for municipal use, 7,000 acre-ft for rural and domestic use, and 38,000 acre-ft for industrial use. Goodell (1985) estimated that in 1980, about 40,000 acre-ft were pumped for municipal use, 9,000 acre-ft for rural and domestic use, and 44,000 acre-ft for industrial use. These estimated, nonirrigation uses of ground water are about 5 percent of estimated total withdrawals for irrigation.

REGIONAL GROUND-WATER FLOW

Ground-water flow in the regional aquifer system underlying the eastern Snake River Plain is generally perpendicular to water-table contours (pl. 4) and is from major recharge areas in the northeast to discharge areas in the southwest. Most recharge takes place along the margins of the plain and in surface-water-irrigated areas; most discharge is from springs along the Snake River near American Falls Reservoir and from Milner to King Hill. A comparison of water-table contours for 1928–30, 1956–58, and 1980 (pl. 4) indicates that regional ground-water levels and the direction of flow have been relatively stable in the central part of the eastern plain for the past 50 years. However, between 1890 and 1920, water levels rose at locations in the southwestern end of the eastern Snake River Plain (table 14), and spring flows increased (fig. 11) in response to surface-water irrigation of large tracts of land on the plain. By 1929, most surface water for irrigation was appropriated, and since 1945 amounts of ground water withdrawn for irrigation have increased. Hydrographs on plate 4 show that, despite strong seasonal variations, ground-water levels have generally declined in most areas since ground-water pumping intensified in about 1950.

In addition to showing a long-term decline in water levels, hydrographs on plate 4 show seasonal and short-term climatic effects. In surface-water-irrigated areas, water levels are usually highest from August through October and lowest in March or April. An example of this type of fluctuation is shown by the hydrograph for well 7N-38E-23DBA1 in Madison County. This well shows about 5 ft of yearly head change in response to surface-water irrigation. In ground-water-irrigated areas, water levels are usually highest from October through March and lowest in July or August. An example of this type of fluctuation is shown by the hydrograph for well 5N-34E-9BDA1 in Jefferson County. This well shows about 4 ft of yearly head change in response to ground-water pumpage. The strong influence of irrigation on ground-water levels is shown by the hydrograph for well 4S-24E-6BBC1. Although this well is more than 20 mi from irrigated areas, water levels fluctuate seasonally in response to irrigation. Of the six hydrographs shown, only the hydrograph for well 3N-29E-14ADD1 does not show seasonal fluctuations.

Water levels also rise and fall in response to climatic trends. Most of the hydrographs on plate 4 show water-level rises from about 1964 to 1976 in response to above-normal precipitation (fig. 4). Surface-water diversions also were above normal during this period because more water was available (fig. 8). Therefore, it seems likely that the water-level rises were due to an overall increase in water supply, rather than solely due to an increase in recharge from infiltration of precipitation on the plain.

In several areas on the eastern plain, shallow flow systems have developed locally in alluvium. Shallow systems are usually along losing river and canal reaches and in areas where excess water is applied for irrigation. In these areas, downward water movement is impeded by fine-grained sediments. Shallow ground-water systems have developed near the lower Henrys Fork near Rexburg; in the Rigby Fan, Mud Lake, Rupert, and Burley areas; and near the mouth of the Big Lost River near Arco. In these areas, water levels in shallow wells completed in alluvium are higher than those in nearby deeper wells. In the Rexburg area, hydraulic heads in piezometers 300 to 500 ft below land surface are 20 to 45 ft lower than in wells less than 100 ft deep. Hydraulic heads near Rigby are 80 ft higher in the upper alluvium than in deeper basalt. In the Mud Lake area, where highly permeable basalt is interfingered with less permeable sedimentary rocks, heads in shallow wells are 50 to 200 ft higher than in deeper wells. The same is true in the Rupert–Burley area, where head differences are 60 to 200 ft, and at the mouth of the Big Lost River, where head differences between shallow and deep wells are 300 to 700 ft. Some of the water in shallow systems is perched and ultimately leaks through an unsaturated zone to recharge the deeper regional system.

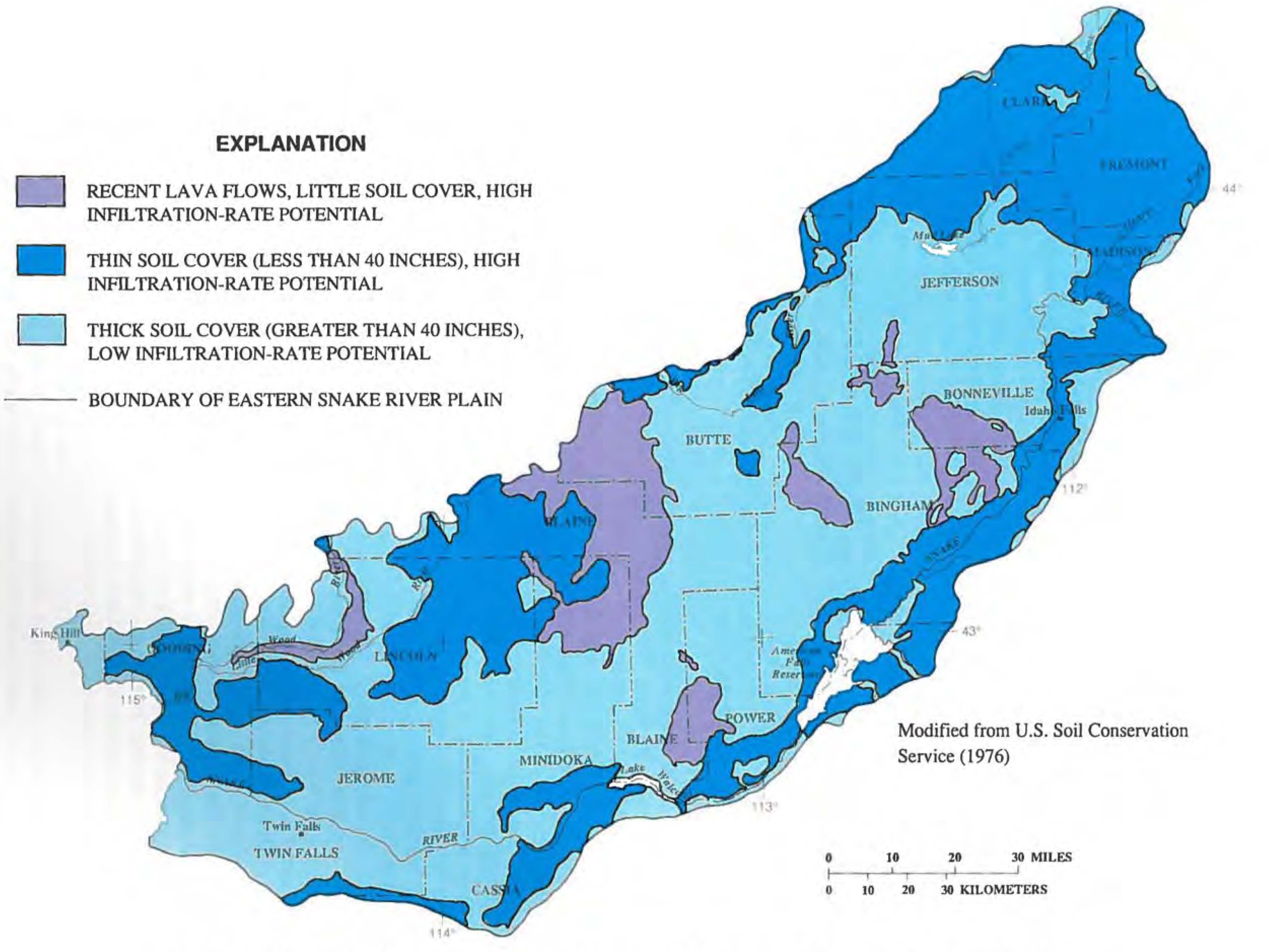


FIGURE 10.—Distribution of generalized soil types. (Description of generalized soil types given in appendix C.)

TABLE 12.—*Recharge from precipitation*

[<, less than; >, greater than]

County or area	Soil description (see fig. 10 and appendix C)	Average annual precipitation (inches per year)	Recharge (inches per year)
Gooding, Jerome, Lincoln, Jefferson	Thin soil cover (<40 in.), high infiltration rate, group 2, appendix C	10	1
Butte, Blaine, Minidoka	Recent lava flows, little soil cover, group 1, appendix C	10	3.5
Central part of plain	Thick soil cover (>40 in,), low infiltration rate, group 3, appendix C	8-10	.3
Blaine, Power, Bingham, Bonneville	Recent lava flows, little soil cover, group 1, appendix C	8	2.8
Lands adjacent to the Snake River	Thin soil cover (<40 in.), high infiltration rate, group 2, appendix C	10	2
Fremont, Clark	Thin soil cover (<40 in.), high infiltration rate, group 2, appendix C	16-20	6

TABLE 13.—*Recharge from tributary streams, underflow, and precipitation, 1911–80*

[Values in acre-feet per year; locations shown on pl. 1]

Years	North side[1] Stream loss	North side[1] Underflow	South side[2] underflow	Precipitation[3]
1911–20	200,000	1,060,000	480,000	740,000
1921–30	170,000	900,000	460,000	650,000
1931–35	120,000	650,000	280,000	620,000
1936–40	150,000	780,000	340,000	820,000
1941–45	210,000	1,110,000	420,000	670,000
1946–50	180,000	940,000	490,000	720,000
1951–55	200,000	1,050,000	400,000	610,000
1956–60	190,000	1,030,000	370,000	630,000
1961–65	200,000	1,070,000	400,000	700,000
1966–70	240,000	1,260,000	420,000	700,000
1971–75	240,000	1,300,000	700,000	970,000
1976–80	190,000	1,010,000	450,000	700,000
Average	190,000	1,000,000	440,000	700,000

[1]Includes Clover Creek, Thorn Creek, Big Wood River, Silver Creek, Little Wood River, Fish Creek, Big Lost River, Little Lost River, Birch Creek, Warm Springs Creek, Deep Creek, Medicine Lodge Creek, Beaver Creek, Camas Creek, and Big Bend Ridge area. Flows varied using the gage on the Big Lost River at Mackay as an index.

[2]Includes Salmon Falls Creek, Cottonwood Creek, Rock Creek (Twin Falls County), Dry Creek, Goose Creek, Raft River, Rock Creek (Power County), Bannock Creek, Portneuf River, Lincoln Creek, Ross Fork Creek, Blackfoot River, Willow Creek, Snake River, Rexburg Bench, Teton River, and Henrys Fork. Flows varied using the gage on the Portneuf River at Pocatello as an index.

[3]Estimated from October to March precipitation at Aberdeen, Ashton, Bliss, and Idaho Falls stations.

Water levels in piezometers at test hole 4N-38E-12BBB1,2,3,4,5 (fig. 13) in a recharge area decrease with depth. Piezometers 1, 3, and 5 are completed in major aquifer zones separated by clay units. Hydraulic-head differences between zones are several tens of feet, whereas head differences between piezometers 2 and 3 (completed in the upper basalt aquifer and lacking clay units) are small. Water-level changes in all piezometers indicate seasonal water-level rises and declines in response to surface-water irrigation. Highest water levels are in summer and early fall; lowest levels are in early spring. Water levels in piezometer 1 typically peak in June or July, whereas water levels in piezometer 5 usually peak in September or October. The time lag in head change with depth probably is due to the effect of clay units with low hydraulic conductivity. Both the 80-ft head change with depth and seasonal fluctuations of about 20 ft are, in large part, attributed to the application of surface water in excess of crop consumptive use requirements and the presence of clay units. In the vicinity of test hole 4N-38E-12BBB1,2,3,4,5, ground-water recharge is about 8 ft/yr.

In several areas on the eastern plain, deeper wells have higher heads than shallower wells. Upward gradients have been defined in ground-water-discharge areas near American Falls Reservoir and along the Snake River from Milner to King Hill. In both areas, the regional ground-water system discharges horizontally and vertically to the Snake River as spring flow and seepage. However, on a local scale, the same area also receives recharge from surface-water irrigation.

Upward gradients also are evident near the Roberts area between the Snake River and Mud Lake.

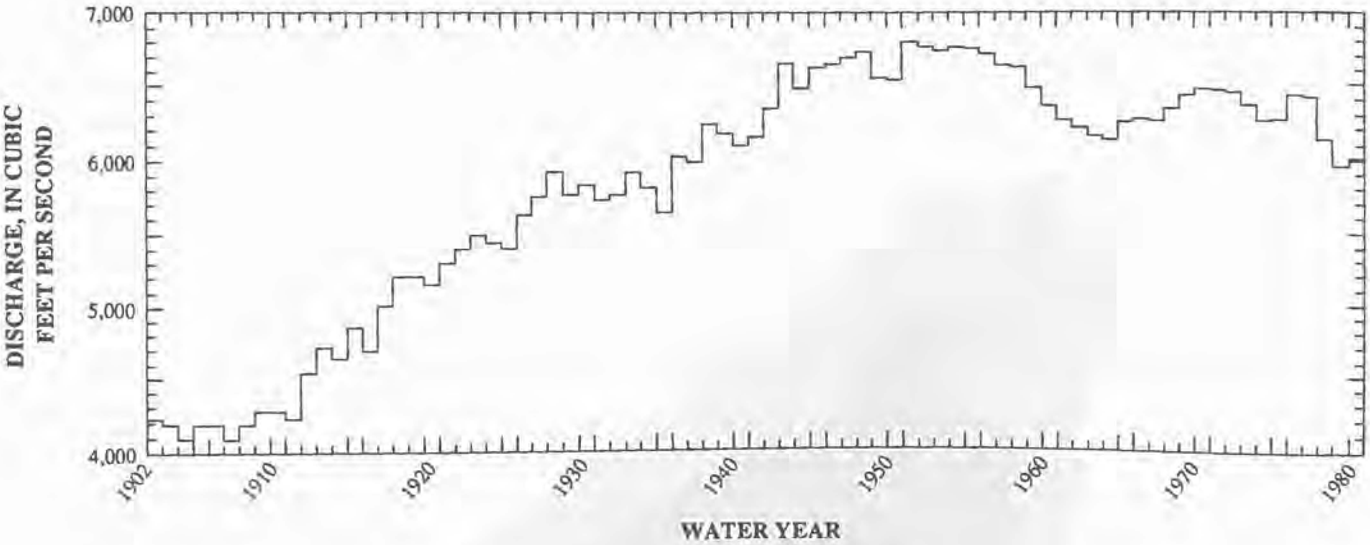


FIGURE 11.—Mean annual ground-water discharge along the north side of the Snake River from Milner to King Hill. (Modified from Kjelstrom, 1986.)

TABLE 14.—*Ground-water-level changes*

[From Mundorff and others, 1964, p. 162]

Well location	Date	Depth to water (feet below land surface)	Date	Depth to water (feet below land surface)	Water-level rise (feet)
6S-13E- 6DD	Before 1901	430	1959	350	80
8S-15E-28BA	1909	94	1959	62	32
7S-15E-33	1907	190	1959	150	40
7S-15E- 8	1907	215	1959	190	25
5S-15E-31 or 32	1907	145	1959	110	35
6S-17E- 2AB	1890	Dry at 280	1952	210	70
8S-17E-19BB1	1907	342	1954	298	44
8S-18E-15CC	1907	318	1959	200	118
4S-19E-26DA1	1913	330	1957	311	19
9S-19E-15AC	1907	252	1959	160	92
9S-19E-26	1912	189	1959	127	62
6S-20E-15DA	Before 1901	341	1959	200	141
7S-23E- 5	Before 1901	265	1959	210	55
8S-25E- 1CB1	Before 1901	375	1959	185	190
9S-24E-29AA1	1905	101	1951	59	42

Around the northeast end of American Falls Reservoir, water levels in wells completed below fine-grained lakebeds are about 20 ft higher than water levels in wells completed in the shallow alluvium. The lakebeds confine water in underlying sand, gravel, and basalt aquifers. Springs discharge to the Snake River, Spring Creek, and the Portneuf River where the streams have eroded through the confining lakebeds. In the Roberts area, deep wells have heads 50 to 130 ft higher than shallow wells. Ground water is likely confined by lakebeds similar to those in the Mud Lake area.

Water levels in piezometers at test hole 7S-15E-12CBA1,4,5 (fig. 14) in a discharge area increase with depth. This test hole was drilled as part of the present study, and results were summarized by Whitehead and Lindholm (1985). The water level in piezometer 5 is about 155 ft higher than the water level in piezometer 1. The test hole is in a regional discharge area for the eastern plain that includes Thousand Springs, about 12 mi southwest of the drill site. Silt and clay between piezometers 1 and 4 and massive basalt between piezometers 4 and 5 are confining units. The water-level rise from May to November is in response to applied irrigation water (pl. 3) and canal leakage. Although water levels in the three piezometers peak at about the same time, the smaller rise in piezometer 5 compared with that in piezometer 1 probably is due to impedance of flow through confining zones.

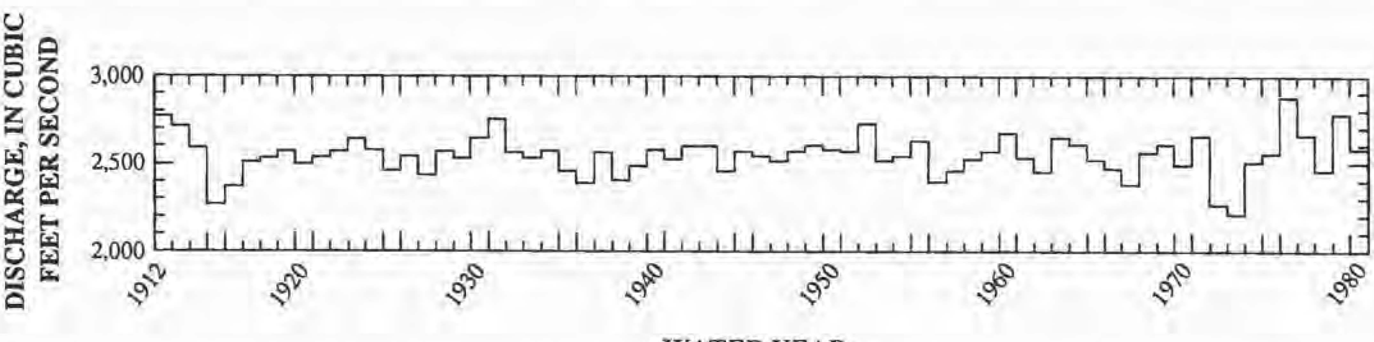


FIGURE 12.—Mean annual ground-water discharge to the Snake River from near Blackfoot to Neeley. (Modified from Kjelstrom, 1986.)

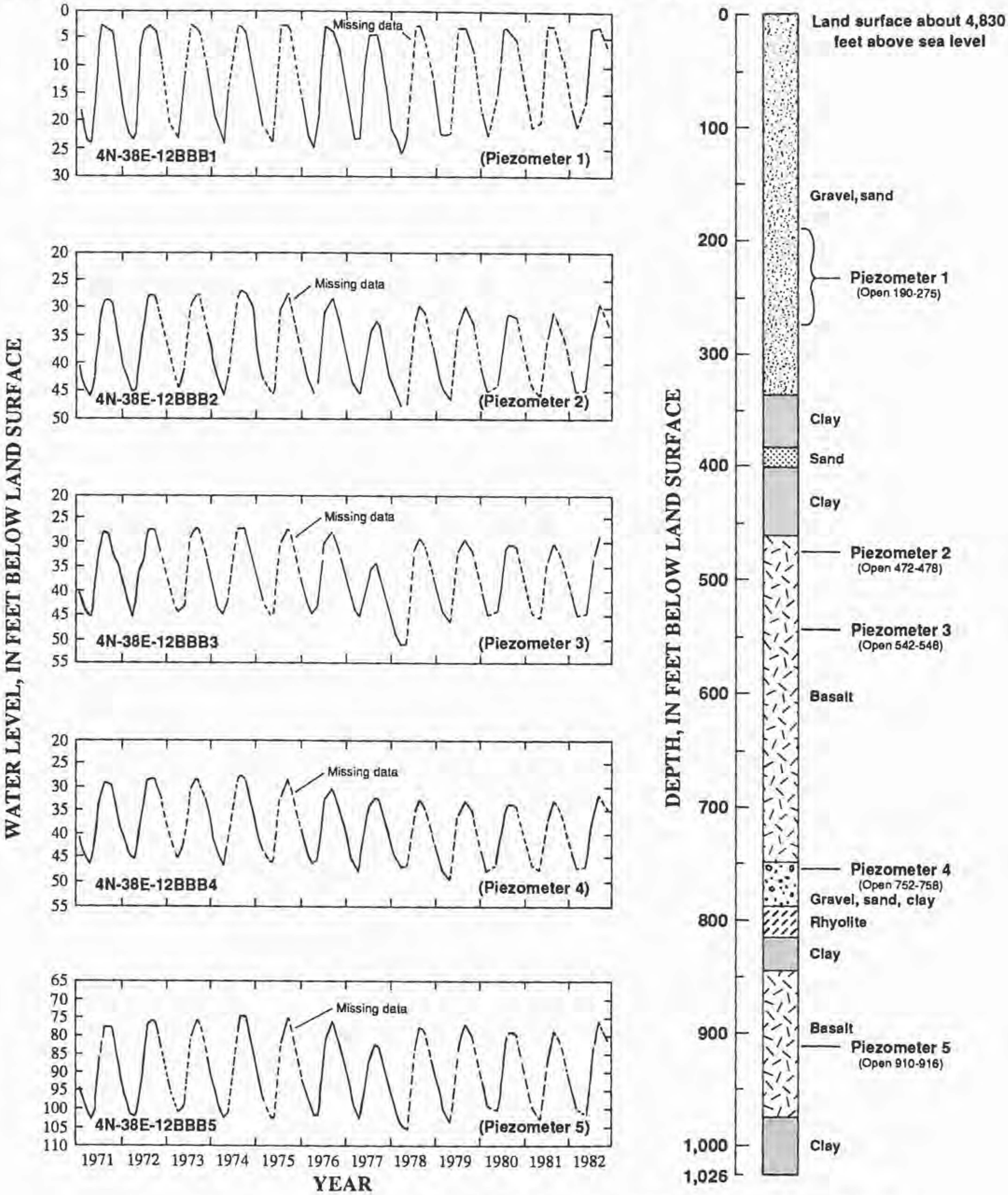


FIGURE 13.—Hydrographs and lithologic log, test hole 4N-38E-12BBB1,2,3,4,5. (Location shown on pl. 4)

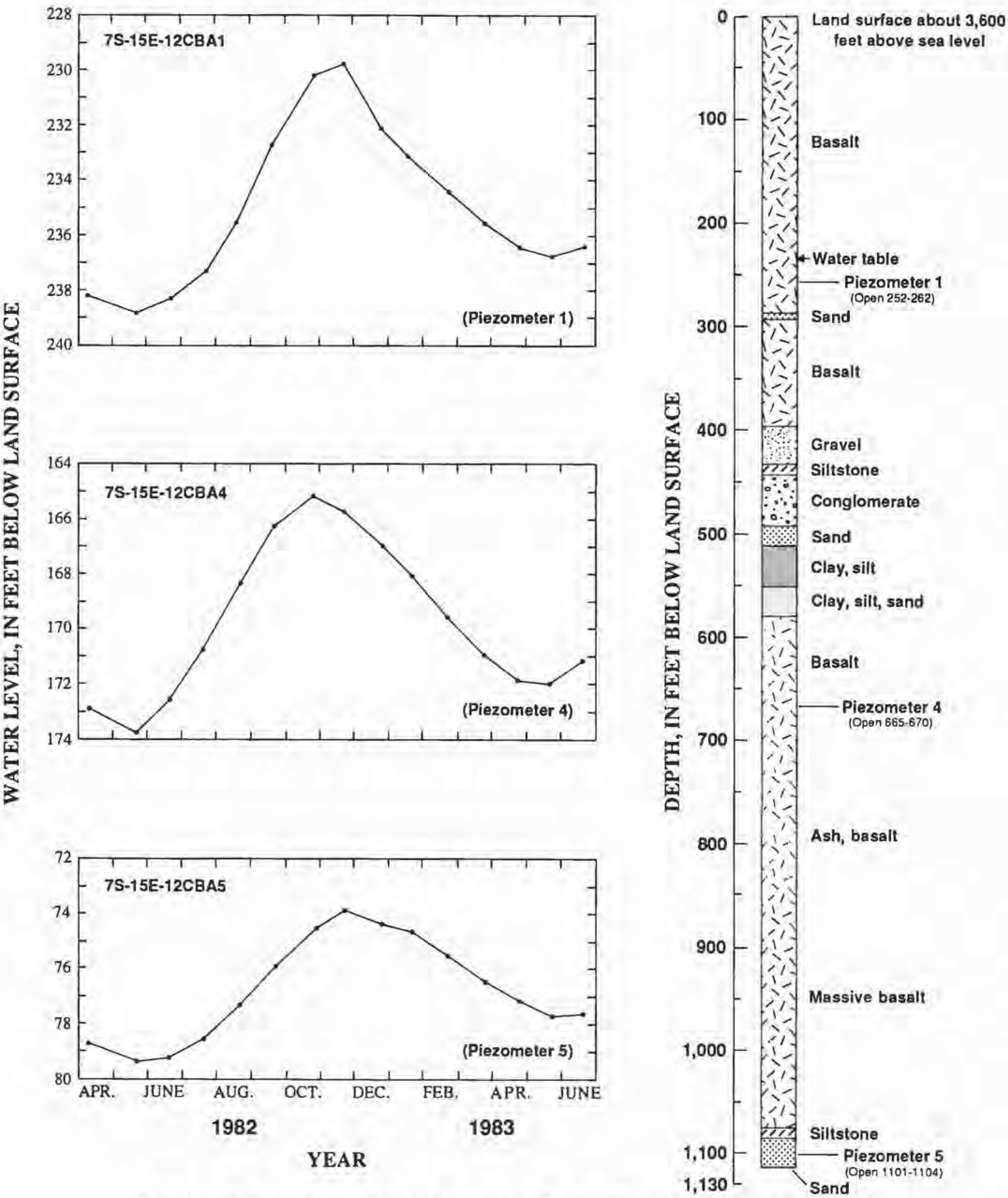


FIGURE 14.—Hydrographs and lithologic log, test hole 7S-15E-12CBA1,4,5. (Location shown on pl. 4)

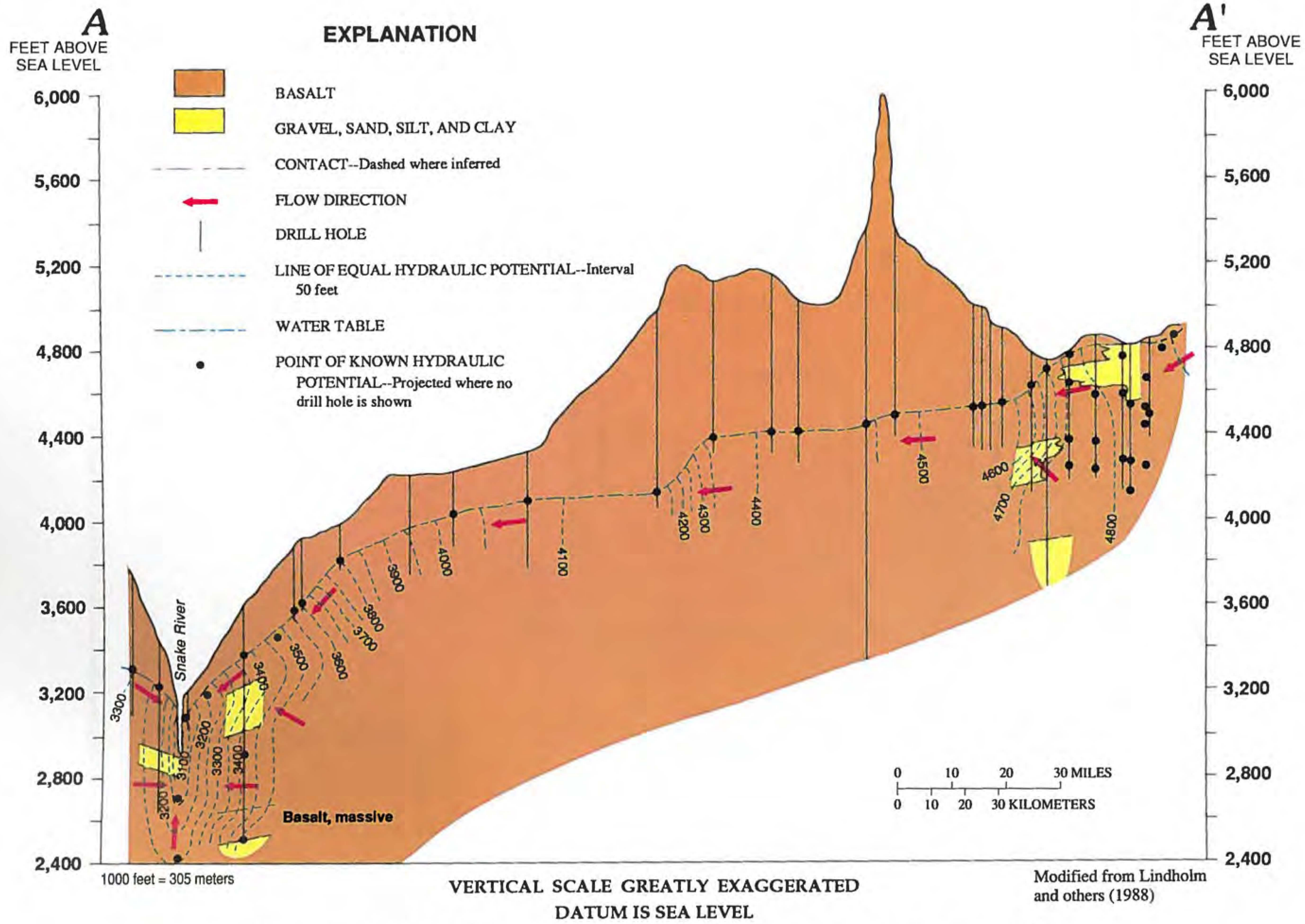


FIGURE 15.—Section *A–A´* showing general directions of ground-water flow. (Line of section shown on pl. 4)

A longitudinal flow section from a major recharge area, Henrys Fork, to a major discharge area, Thousand Springs, is shown in figure 15. The regional aquifer underlying the eastern plain generally is unconfined, but local confined conditions are apparent in the Mud Lake area, where the water-table gradient is steep (pl. 4) owing to intercalated basalt and fine- to coarse-grained sediments (Lindholm and others, 1988). The fine-grained sediments likely confine flow and reduce overall aquifer transmissivity. The steep gradient between Arco and Lake Walcott (pl. 4) may be due to decreased transmissivity along a rift zone (pl. 2) where dikes may have healed fractures perpendicular to the direction of ground-water flow (Lindholm and others, 1988).

The steep gradient near the Snake River is due to thinning of the basalt aquifer (pl. 2, section *A–A'*) and reduction in transmissivity. Little or no underflow leaves the eastern plain because the Snake River is a regional sink, as indicated by the converging flow lines in figure 15.

GROUND-WATER BUDGET

A water year 1980 ground-water budget was compiled for the eastern Snake River Plain (table 15). A net loss in aquifer storage of about 100,000 acre-ft was calculated from water-level changes measured in observation wells in water year 1980 (fig. 16). Storage coefficients used to calculate change in aquifer storage were 0.05 for basalt (aquifer-test data, Mundorff and others, 1964, p. 156) and 0.20 for sediments. The calculated change in aquifer storage compares favorably with the residual from the ground-water budget in table 15.

The most accurate estimates in the water year 1980 ground-water budget are of Snake River gains and losses; errors of these estimates range from 3 to 10 percent. Estimates of recharge from surface-water irrigation are less accurate because *ET* values used in calculations are empirical. *ET* is particularly difficult to estimate for large areas with varying climatic conditions and crop types. Estimates of recharge from tributary drainage basins (streamflow and underflow) vary in accuracy because discharge from some streams is measured directly, whereas discharge from other streams is estimated from basin-yield equations. Change in aquifer storage was calculated using data from widely scattered observation wells and estimates of the aquifer storage coefficient and is, therefore, approximate. The estimation of change in storage does, however, compare well with the residual from the ground-water budget, not only in sign (net loss), but also in magnitude.

TABLE 15.—*Ground-water budget, water year 1980*

Recharge	Acre-feet
Surface-water irrigation	4,840,000
Tributary drainage-basin underflow	1,440,000
Direct precipitation	700,000
Snake River losses	690,000
Tributary-stream and canal losses	390,000
Total	8,060,000
Discharge	
Snake River gains	7,080,000
Ground-water pumpage (net)	1,140,000
Total	8,220,000
Change in aquifer storage (budget residual)	−160,000
Estimated change in aquifer storage from water-level changes	−100,000

The least accurate estimates in the water year 1980 ground-water budget are of recharge from infiltration of precipitation. Although precipitation is measured at several sites, aquifer recharge from precipitation cannot be measured directly. Mundorff and others (1964, p. 184) estimated that recharge from precipitation is about 500,000 acre-ft annually. Given the difference in sizes between the area studied by Mundorff and others (8,400 mi²) and that used for this study (10,800 mi²), the difference in estimates (200,000 acre-ft) is reasonable.

Although individual budget-item errors may be large, the overall budget error is small. The similarity between change in aquifer storage and the budget residual (table 15) is due to compensating errors in calculations of *ET*, basin yield, and recharge from precipitation.

GROUND-WATER-FLOW MODELING

APPROACH

The approach used in this study was to develop a digital computer model of the eastern Snake River Plain regional aquifer system for testing various concepts of regional ground-water flow. Modeling progressed in stages from two-dimensional steady-state simulations to three-dimensional steady-state and transient simulations. Results and conclusions from each stage are documented; the emphasis in this report is on the final stage of modeling.

The digital computer model is a mathematical representation of the ground-water-flow system.

EXPLANATION

—2— LINE OF EQUAL WATER-LEVEL CHANGE--Intervals 1 and 5 inches

• OBSERVATION WELL

——— BOUNDARY OF EASTERN SNAKE RIVER PLAIN

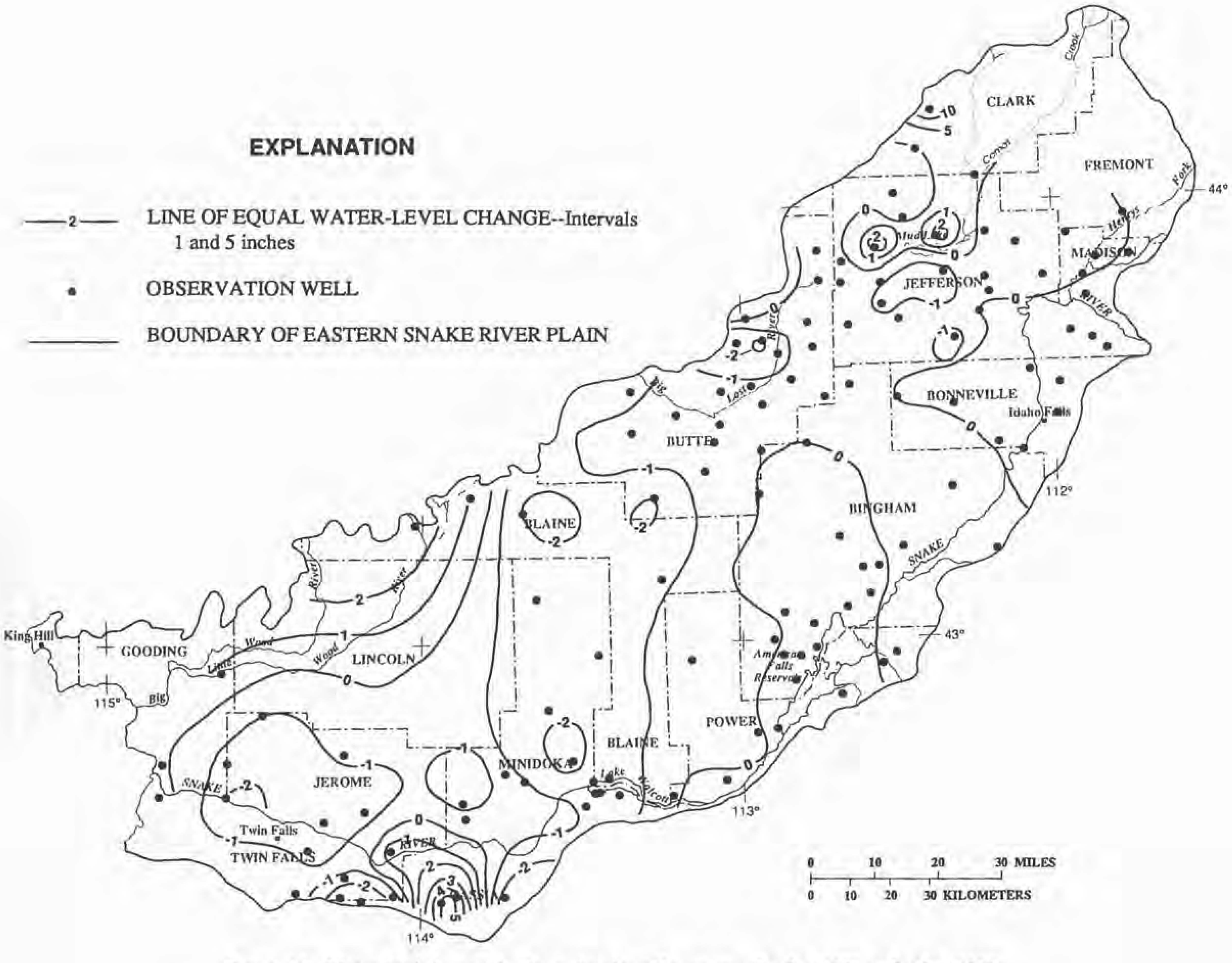


FIGURE 16.—Water-level changes in the regional aquifer system, October 1979 to October 1980.

Modeling complex aquifer systems requires several simplifying assumptions. The validity of model assumptions can be judged by how well field conditions are simulated. When model results approximate field hydrologic conditions within reasonable limits of error, the model is assumed to be calibrated and valid for hydrologic analysis.

The goal of modeling was to simulate known aquifer conditions (head and spring discharge) within reasonable ranges of values and to define the hydrologic effects of changes in model input data through sensitivity testing. Because model solutions are not unique, a model can be calibrated using physically unrealistic input data. Some adjustments of model parameters made during model calibration are justified on the basis of available evidence; other adjustments, although not justified with available data, may indicate where additional data are needed. Differences between simulated and measured head and spring discharge also indicate areas where model refinement is needed.

Model input data for this study are based on geologic and hydrologic information with varying degrees of accuracy. For example, streamflow measurements are considered accurate within about ±5 percent, whereas aquifer hydraulic conductivity, which commonly is estimated by indirect methods, may be in error by one to several orders of magnitude.

Initial values of aquifer hydraulic properties, recharge, discharge, and pumpage were estimated for model input. Model output then was compared with known aquifer conditions to determine the reasonableness of the initial estimates. The model was tested to determine its sensitivity to changes in transmissivity, storage coefficient, aquifer leakance, recharge, riverbed or spring-outlet conductance, ground-water pumpage, and boundary flux. Input data were varied within reasonable ranges to achieve a better fit to known conditions. Adjustments were made to least known and to most sensitive model parameters. Simulation results indicated how the aquifer might have responded to past stresses, such as increased pumping and reduced recharge, and how the aquifer might respond to hypothetical future stresses.

ASSUMPTIONS

To model the regional aquifer system, assumptions were made concerning aquifer properties, hydraulic fluxes, and initial conditions for transient analysis. The major assumptions are outlined in this section; other, more specific assumptions are discussed in subsequent modeling sections. Ground-water flow was assumed to be laminar and the Darcy flow equation applicable. A three-dimensional finite-difference ground-water-flow model (McDonald and Harbaugh, 1988) was used most extensively in this study. Vertical variations in head within each model layer were assumed to be negligible, and head losses between layers were assumed to be controlled by confining beds near the base of each layer. Therefore, model-simulated heads are an approximate average of heads within that aquifer layer.

Local aquifers perched above the regional aquifer system were not simulated, although recharge to perched aquifers was assumed to ultimately reach the regional aquifer. Vertical hydraulic conductivity was assumed to be anisotropic owing to low-hydraulic-conductivity basalt between permeable flow tops and fine-grained layers within sand and gravel zones. Horizontal hydraulic conductivity was assumed to be isotropic because two-dimensional simulations indicated small differences in modeling results between isotropic and anisotropic conditions.

For steady-state model analysis, calculated 1980 water-year fluxes (table 16) were assumed to approximate the average annual flux for the period 1950–80. During that period, hydrologic conditions were stable relative to conditions from 1880 to 1950. Ground-water recharge was assumed to equal discharge (steady-state conditions) for the period 1950–80 because irrigation diversions and ground-water levels were relatively stable (fig. 8 and pl. 4). Ground-water-level declines due to pumping, climatic variations, and decreased surface-water diversions during that period were generally small, and changes in storage were accordingly small (about 1 percent, table 16). Approximate steady-state fluxes were computed by including these small changes in storage as part of the recharge term. Recharge from irrigation was assumed to take place directly below surface-water-irrigated areas.

For 1891 to 1980 transient calculations, 5-year averages were assumed to adequately represent long-term variations in flux. It was also assumed that initial (preirrigation) conditions could be approximated by removing recharge due to surface-water irrigation and ground-water pumping from the calibrated steady-state model. Surface-water altitudes, used in the model for river-leakage simulations, were corrected for prereservoir conditions. The estimated preirrigation steady-state condition was a stable initial condition for transient simulation; therefore, simulated changes were assumed to result from changes in model input, not from nonequilibrium initial conditions.

TABLE 16.—*Steady-state model mass balance, water year 1980*

[Values in cubic feet per second]

Inflow		Outflow	
A. Specified flow	2,740		
C. Recharge	7,800	B. Wells	1,790
D. Snake River losses[1]	1,140	D. Snake River gains[2]	9,890
Total	11,680	Total	11,680

A. Specified flow includes the following:	
Tributary drainage-basin underflow simulated as recharge wells (table 11)	1,980
Stream and canal losses (table 10)	540
Irrigation-return flow in Mud Lake area	220
Total	2,740
B. Ground-water discharge includes the following:	
Total ground-water pumpage	1,790
Irrigation-return flow in Mud Lake area	−220
Net ground-water pumpage (table 15)	1,570
C. Ground-water recharge includes the following:	
Surface-water irrigation (table 5)	6,690
Precipitation (table 15)	970
Change in storage (table 15)	140
Total	7,800

D. Ground-water gain from or loss to the Snake River and tributaries:

Snake River reach	Calculated loss (−) or gain	Measured loss (−) or gain
1. Hagerman to King Hill	−1,530	−1,410
2. Buhl to Hagerman	−3,830	−3,610
3. Kimberly to Buhl	1,430	−1,220
4. Milner to Kimberly	−100	−300
5. Minidoka to Milner	0	−130
6. Neeley to Minidoka	−20	−180
7. Near Blackfoot to Neeley	−2,640	−2,620
8. At Blackfoot to near Blackfoot	200	270
9. Shelley to at Blackfoot	160	150
10. Lewisville to Shelley	370	380
11. Lorenzo to Lewisville	−40	−290
12. Heise to Lorenzo	180	150
Tributaries		
13. Lower Henrys Fork	−30	−120
14. Lower Salmon Falls Creek	−40	0
Totals	−8,750	−8,930

[1]Losses and gains totaled on a block-by-block basis.
[2]Losses and gains totaled for each river reach.

TWO-DIMENSIONAL STEADY-STATE SIMULATIONS AND PREVIOUS MODELING STUDIES

A nonlinear, least-squares regression technique for estimating aquifer parameters was initially applied to the regional aquifer system in the eastern Snake River Plain. The application and results are explained in an earlier report (Garabedian, 1986); only major points are repeated here.

The parameter-estimation computer program is based on a technique outlined by Cooley (1977, 1979, 1982) for two-dimensional steady-state ground-water flow. Hydrologic data for the 1980 water year (steady-state conditions assumed) were used to calculate recharge rates, boundary fluxes, and spring discharges. Ground-water use was estimated from irrigated-land maps and values of crop consumptive use. These mass-flux estimates and riverbed or spring-outlet conductances (hydraulic conductivity of confining bed and streambed divided by bed thickness) were used as fixed values during each model simulation for the calibration of transmissivity. Because the parameter-estimation model can calibrate some parameters automatically, but not all parameters at once (without prior information), some parameters were held fixed during the model run but were adjusted for better model fit for the next run. Riverbed or spring-outlet conductance values were adjusted between simulations by comparing simulated spring discharges with measured discharges.

Simulation results indicate a wide range in average transmissivity from about 0.05 to 44 ft^2/s (fig. 17) and, in average riverbed or spring-outlet conductance, from about 2×10^{-9} to 6×10^{-8} (ft/s)/ft. Along with parameter values, model statistics were calculated, including correlation coefficient between simulated and measured heads (0.996), standard error of head estimates (40 ft), and parameter coefficients of variation (about 10–40 percent). The high correlation coefficient indicates a good statistical fit between simulated heads and measured heads in the regional aquifer. About 95 percent of simulated head values were within 80 ft of measured head values. The coefficient of variation for simulated model parameters can be used to form confidence limits for these estimated parameters.

Estimated transmissivity values were lowest along the margins of the plain, where model errors were highest. Model errors, particularly along plain margins, were likely due to violation of the assumption that ground-water flow is two dimensional and steady state. Model fit improved slightly when y-direction (northwest-southeast) transmissivity values were larger relative to x-direction (northeast-southwest) transmissivity values. This result may be due to the impedance of flow in the vicinity of the northwest-trending rift zone between Arco and Lake Walcott (pl. 2). The difference between x and y transmissivity in modeling results is slight (about 20 percent), which indicates little regional anisotropy across the modeled area. The significant decrease in transmissivity immediately upgradient from the rift zone may be related to fracturing and, possibly, subsequent healing of fractures by later movements of magma.

Simulated heads were most sensitive to changes in recharge and, in some areas, transmissivity (particularly near springs along the Snake River from Milner to King Hill). Modeling results also were sensitive to the distribution and number of the discretized zones (fig. 17). As the number of zones (and model parameters) was increased, model fit generally improved; however, the tendency for nonconvergence also increased. Therefore, a sufficient number of zones were used to achieve a good model fit and still maintain model stability and convergence.

Transmissivity values of the regional aquifer system obtained by the preliminary two-dimensional steady-state simulation (fig. 17) were compared with values obtained by Mundorff and others (1964, pl. 6) (fig. 18), Norvitch and others (1969, p. 37) (fig. 19), and Newton (1978, p. 67–71) and are presented in table 17. In the central part of the plain, the transmissivity values are similar in all four studies, as indicated by high and low values. Major differences were noted along the margins of the plain where the model results were consistently lower.

Along the margins of the plain, hydraulic head generally decreases with depth and recharge is predominant. Where heads increase with depth, such as between Blackfoot and Neeley and between Milner and King Hill, discharge predominates, though recharge from surface-water irrigation also may take place. Three-dimensional simulation is needed to properly simulate the vertical variations in head. In the central part of the plain, heads generally do not change with depth, and flow is largely horizontal and two dimensional.

THREE-DIMENSIONAL GROUND-WATER-FLOW MODEL

As described in the preceding parts of this report, the regional aquifer system in the eastern Snake River Plain is three dimensional. Simulation in two dimensions is an oversimplification because heads change with depth in areas of recharge and discharge. Therefore, in the RASA study, a three-dimensional finite-difference model was the final stage of ground-water-flow modeling.

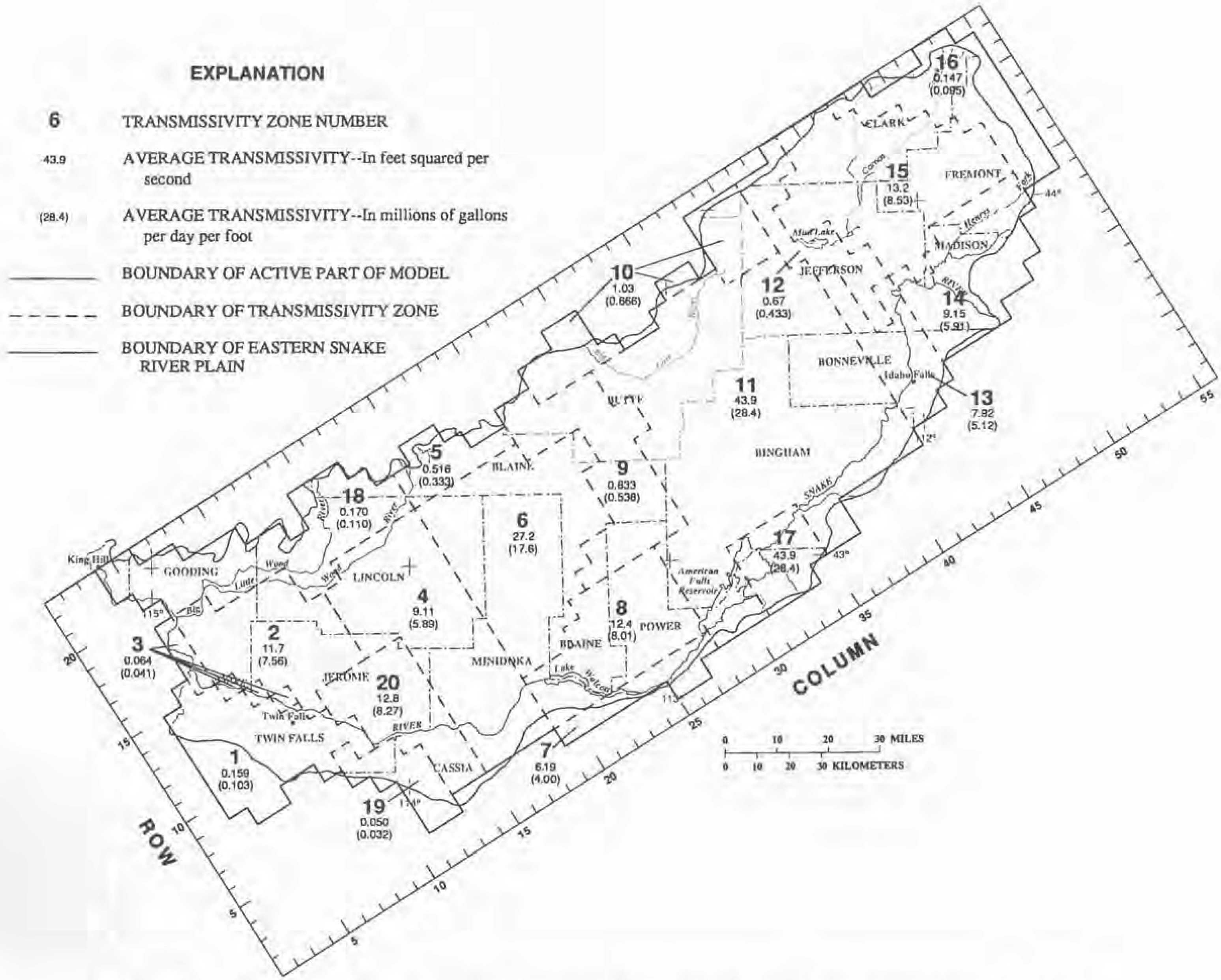


FIGURE 17.—Finite-difference grid and average transmissivity values estimated using a two-dimensional steady-state model.

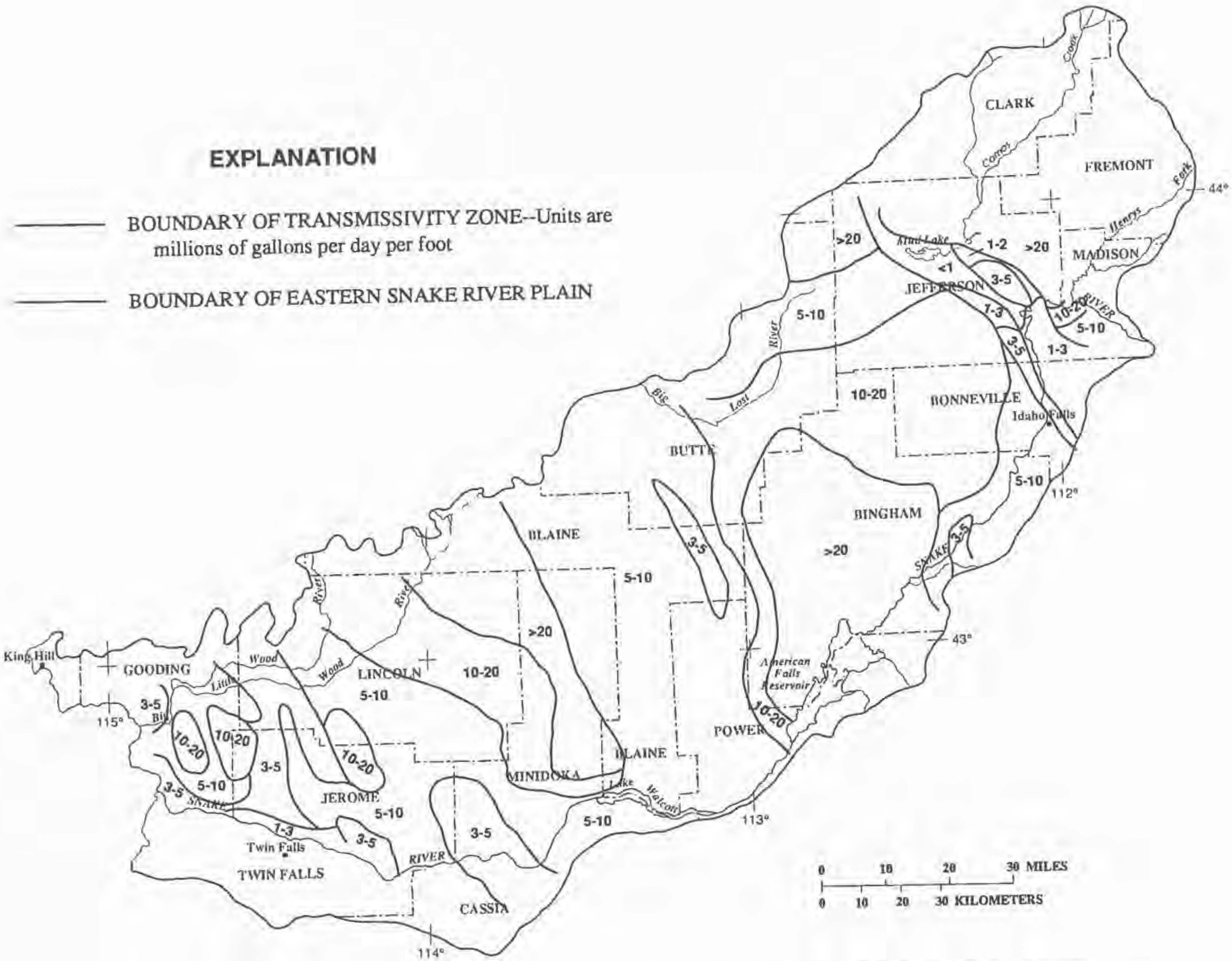


FIGURE 18.—Transmissivity distribution of the eastern Snake River Plain aquifer (Mundorff and others, 1964).

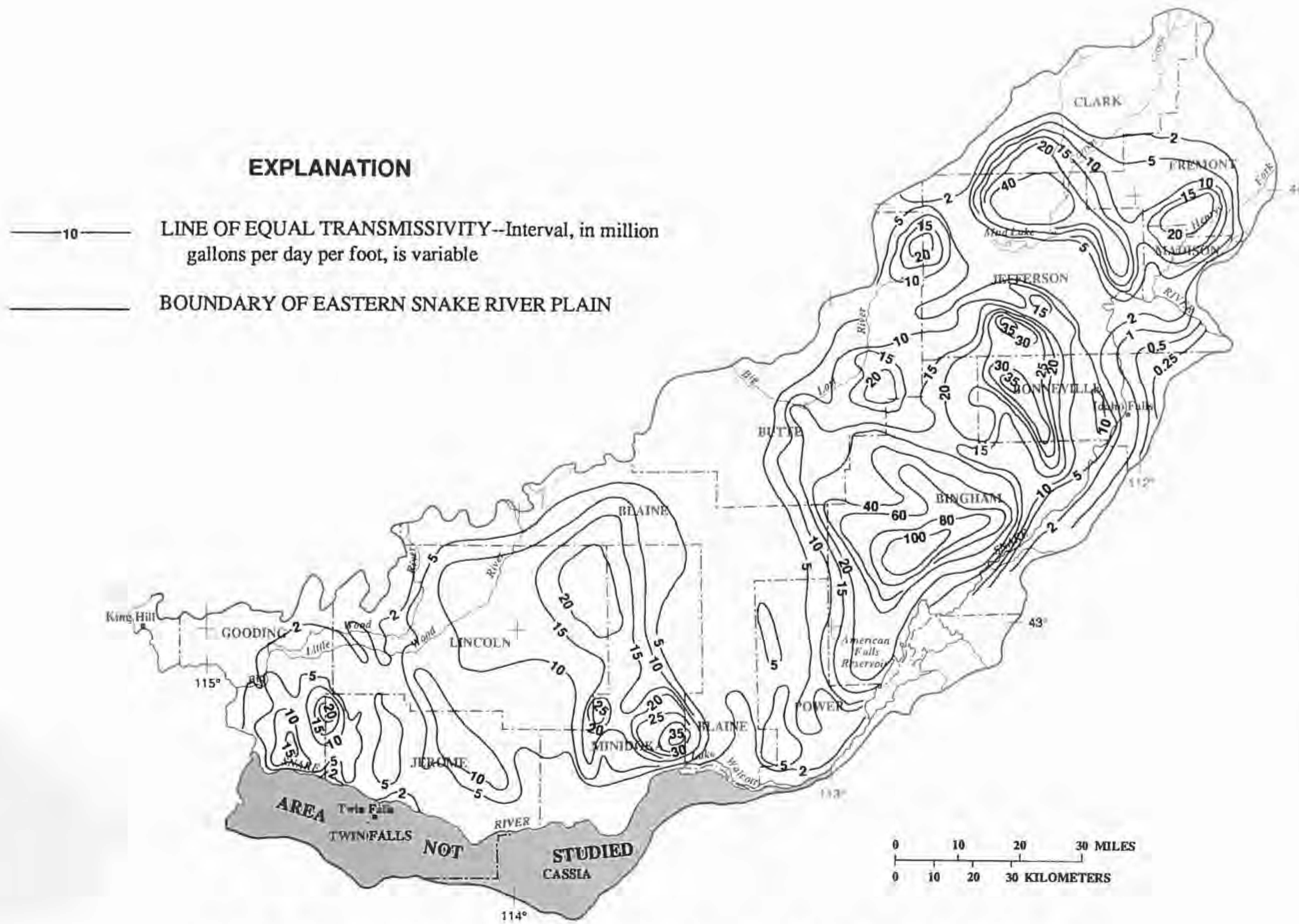


FIGURE 19.—Transmissivity distribution of the eastern Snake River Plain aquifer (Norvitch and others, 1969).

TABLE 17.— *Comparison of published transmissivity values with two-dimensional steady-state regression results*

[Values in feet squared per second; >, greater than; <, less than; —, no data available]

Model zone (fig. 17)	Mundorff and others (1964)	Norvitch and others (1969)	Newton (1978)	Garabedian regression results (1986)
1	—	—	—	0.16
2	8	11	8	12
3	5	3	8	.064
4	15	15	10	9.1
5	15	11	30	.52
6	30	20	35	27
7	—	<3	.3	6.2
8	11	8	2	12
9	8	8	3	.83
10	>30	15	6	1.0
11	30	50	35	44
12	1	8	4	.67
13	5	2	25	7.9
14	15	<8	3	9.2
15	>30	30	9	13
16	—	—	—	.15
17	>30	15	10	44
18	8	3	10	.17
19	—	—	—	.050
20	8	8	8	13

An equation describing three-dimensional flow of ground water is

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)-q(x,y,z,t)=S_s\frac{\partial h}{\partial t}, \quad (2)$$

where

x,y,z = Cartesian coordinate direction (length);
K_{xx}, K_{yy}, K_{zz} = hydraulic conductivity in the specified coordinate direction (length/time);
h = aquifer head (length);
q = flow from or into the aquifer from outside sources or sinks (1/time);
S_s = specific storage (1/length); and
t = time.

Equation (2) describes ground-water flow in a heterogeneous and anisotropic aquifer, and coordinate axes must be aligned with the major axes of hydraulic conductivity. For most problems, an analytical solution to equation (2) cannot be obtained, and approximate methods of solution are used. A finite-difference approximation to equation (2) using the notation around aquifer block i,j,k as shown in figure 20 is (McDonald and Harbaugh, 1988)

$$\begin{aligned}
&CR_{i,j-\frac{1}{2},k}\left(h^m{}_{i,j-1,k}-h^m{}_{i,j,k}\right)+CR_{i,j+\frac{1}{2},k}\left(h^m{}_{i,j+1,k}-h^m{}_{i,j,k}\right)\\
&+CC_{i-\frac{1}{2},j,k}\left(h^m{}_{i-1,j,k}-h^m{}_{i,j,k}\right)+CC_{i+\frac{1}{2},j,k}\left(h^m{}_{i+1,j,k}-h^m{}_{i,j,k}\right)\\
&+CV_{i,j,k-\frac{1}{2}}\left(h^m{}_{i,j,k-1}-h^m{}_{i,j,k}\right)+CV_{i,j,k+\frac{1}{2}}\left(h^m{}_{i,j,k+1}-h^m{}_{i,j,k}\right)\\
&+CRIV_{i,j,k}\left(R_{i,j,k}-h^m{}_{i,j,k}\right)+Q_{i,j,k}\\
&=S_{s\,i,j,k}\left(\Delta R_j\Delta C_i\Delta V_k\right)\frac{\left(h^m{}_{i,j,k}-h^{m-1}{}_{i,j,k}\right)}{t_m-t_{m-1}},
\end{aligned} \quad (3)$$

where an example of the CR, CC, and CV coefficients is

$$CR_{i,j+\frac{1}{2},k}=\frac{2\Delta C_i TR_{i,j,k}TR_{i,j+1,k}}{\left(TR_{i,j,k}\Delta R_{j+1}\right)+\left(TR_{i,j+1,k}\Delta R_j\right)}.$$

Here, CR is the harmonic mean of conductance at block faces along rows, where

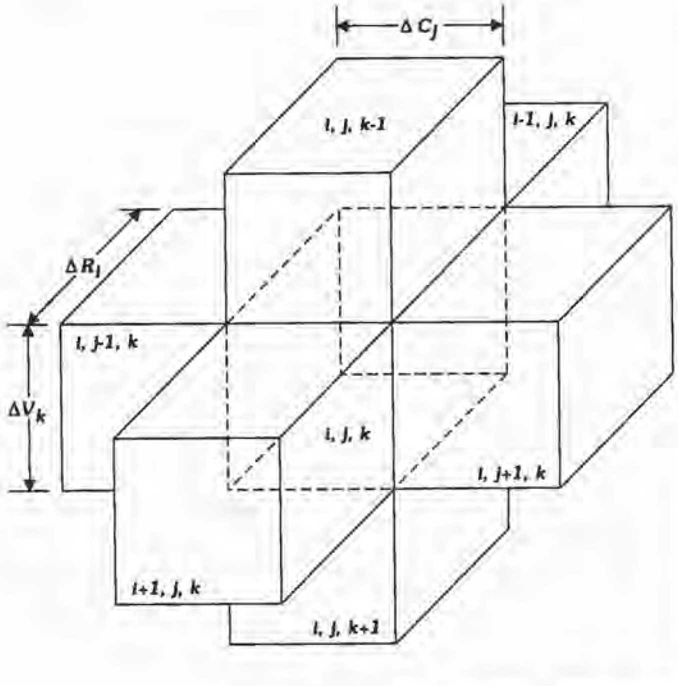
$TR_{i,j,k}$ = transmissivity of block i,j,k along the row;
ΔC_i = block length along columns;
ΔR_j = block length along rows;
ΔV_k = block length along layers—similar expressions for conductances along columns (CC) and layers (CV) can be made;
$CRIV_{i,j,k}$ = conductance of a riverbed or spring outlet, where hydraulic conductivity times river width times river length divided by riverbed thickness equals conductance;
$R_{i,j,k}$ = river-stage or spring-outlet altitude;
$Q_{i,j,k}$ = recharge or discharge (wells); and
$h^m{}_{i,j,k}$ = head at time step m.

In the modeling process, an aquifer is discretized into a number of blocks, and a set of algebraic equations similar to equation (3) is used to represent flow into and out of each block. These equations are solved simultaneously, usually using an iterating solution technique to solve the flow equation (equation 2). The large number of calculations requires the use of a computer.

GRID AND BOUNDARY CONDITIONS

The eastern Snake River Plain was subdivided areally as shown in figure 21. Blocks within the model boundary (active blocks) were assigned values of transmissivity, storage coefficient, and recharge. Blocks outside the model boundary (inactive blocks) were assigned values of zero. The grid was aligned in a southwest to northeast direction to minimize the number of inactive blocks and to align the x-axis in the principal direction of ground-water flow (pl. 4). Point of origin of the model grid (southwest corner) is at lat 41°55'10.00", long 114°28'55.00". The Transverse Mercator projection system was used with a central meridian of 113°30'; the model grid was rotated 31°24' counterclockwise from the central meridian. The grid used in the three-dimensional model is parallel to that used in the two-dimensional model, but the grid spacing is slightly greater—4 mi in the three-dimensional model as compared with 3.95 mi in the two-dimensional model. An even-mile grid facilitated use of Landsat-derived land-use data.

Horizontal and vertical boundaries of the active part of the model were treated as specified flux and along the Snake River as head-dependent flux. Recharge wells were used to simulate underflow from the tributary drainage basins (table 11).



Modified from McDonald and Harbaugh (1988, fig. 3)

FIGURE 20.—Finite-difference notation around aquifer block i,j,k.

The Henrys Fork, Snake, and Portneuf Rivers and Salmon Falls Creek (pl. 1) were represented by river blocks (head-dependent flux) within the modeled area, as shown in figure 21. Model rows and columns, along with river-stage or spring-outlet altitudes, riverbed or spring-outlet conductances, and leakage-cutoff altitudes are listed in table 18. River-stage or spring-outlet altitudes were estimated from topographic maps; estimates of riverbed or spring-outlet conductances were from two-dimensional modeling results. The leakage-cutoff altitude is the level below which leakage from rivers reaches a maximum value. This level was arbitrarily set at 30 ft below river stage along all river reaches and at the same altitude as spring vents to make these blocks discharge areas. The relation of river stage to aquifer head and how that relation controls water movement between the river and the aquifer are shown in figure 22. The rate of river leakage is proportional to the difference between river stage and head in the aquifer until aquifer head drops below a leakage-cutoff altitude. Once the aquifer head drops below the leakage-cutoff altitude, the river leakage is constant and is no longer head dependent.

The regional aquifer system was subdivided vertically into model layers as shown in figure 23. Assigned layers were of equal thickness because differentiation of the predominantly basalt aquifer system into distinct geohydrologic units was not possible. Layer 1 represents the upper 200 ft of the aquifer system; layer 2 is the next 300 ft below. Layers 1 and 2 contain both Quaternary basalt of the Snake River Group and Tertiary basalt. Layers 3 and 4, however, are of lesser areal extent and are present only where basalt of the Snake River Group and interlayered sedimentary rocks are greater than 500 ft thick. Layer 3 is 500 ft or less in thickness and is present only across the central part of the plain. Layer 4 ranges in thickness from 0 to about 3,000 ft in the central part of the plain. The base of the modeled system in the central part of the plain was estimated largely from electrical-resistivity soundings and a few deep drill holes. Underlying layer 4 and forming the assumed base of the regional aquifer system are Quaternary and Tertiary silicic volcanic rocks and Tertiary basalt.

Basalt thickness and generalized distribution of rock types in layers 1 through 4 are shown on plate 5. Basalt is the dominant rock type in layer 1 in the central part of the plain; minor occurrences of rhyolite form isolated buttes in the central part and along the northeastern margin of the plain (pl. 5). Most sedimentary rocks along the boundary of the plain are fine grained, except in the Henrys Fork–Rigby Fan area, the Fort Hall–Portneuf area, the Camas

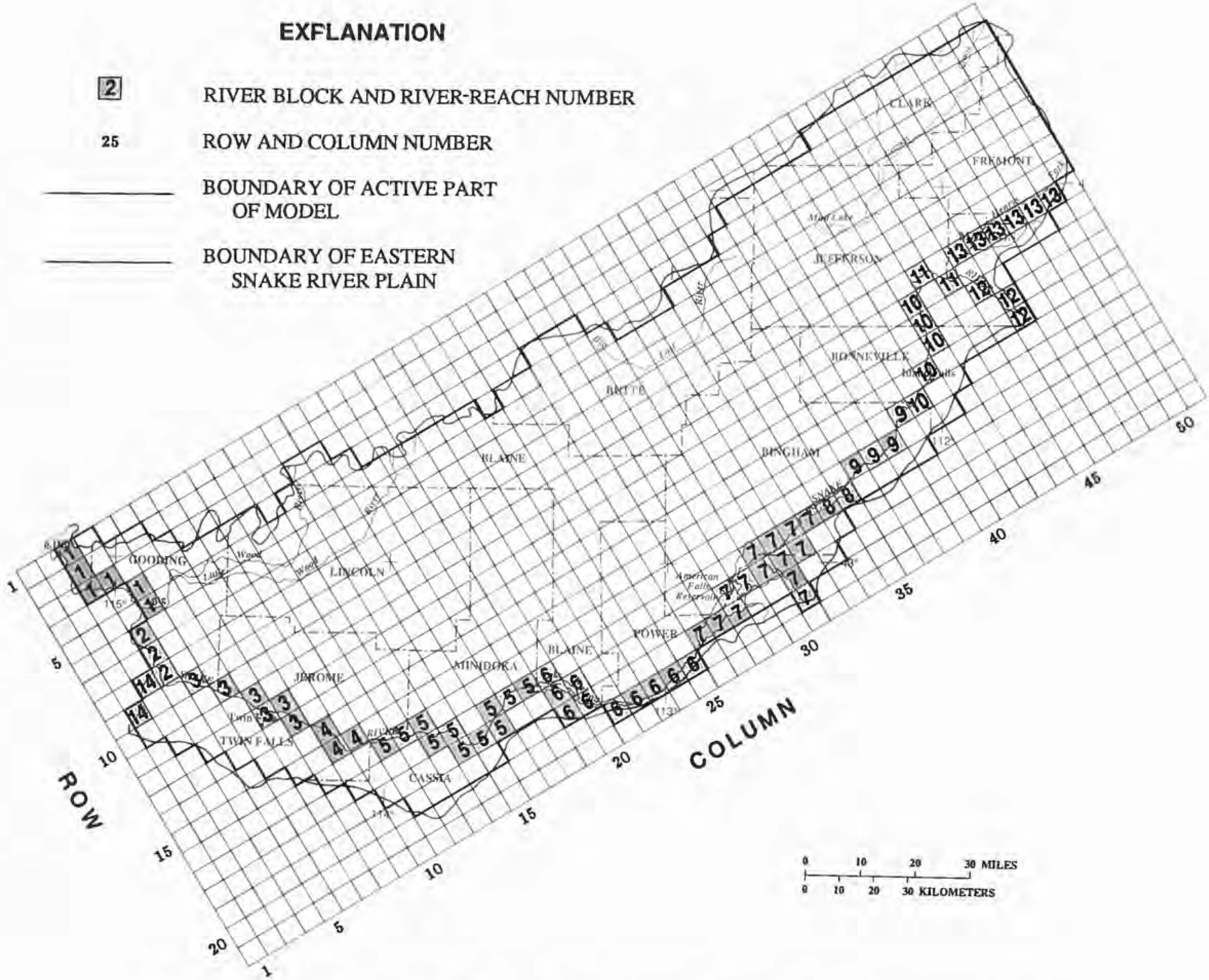


FIGURE 21.—Finite-difference grid, river blocks, and river-reach numbers used for three-dimensional model.

TABLE 18.—*River-block locations, stages, riverbed or spring-outlet conductances, leakage-cutoff altitudes, and reach numbers*

[Altitude, in feet, refers to distance above sea level; row and column numbers shown in fig. 21]

Row	Column	River altitude	Conductance (leakance × river-block area) (feet squared per second)	Leakage-cutoff altitude	Reach Number	Reach Name
1	3	2,600	0.134	2,600	1	Snake, Hagerman to
2	3	2,650	.134	2,650	1	King Hill
3	3	2,725	.134	2,725	1	
3	4	2,800	.134	2,800	1	
4	5	2,900	.134	2,900	1	
5	5	3,050	40.0	3,050	1	
6	4	3,050	40.0	3,050	2	Snake, Buhl to Hagerman
7	4	3,000	40.0	3,000	2	
8	4	3,050	40.0	3,050	2	
9	5	3,100	1.3	3,100	3	Snake, Kimberly to Buhl
10	6	3,150	1.3	3,150	3	
11	7	3,200	1.3	3,200	3	
12	7	3,300	1.3	3,300	3	
12	8	3,500	1.3	3,500	3	
13	8	3,600	1.3	3,600	3	
14	9	3,700	60.0	3,700	4	Snake, Milner to Kimberly
15	9	3,850	60.0	3,850	4	
15	10	3,850	60.0	3,850	4	
16	11	4,130	.20	4,100	5	Snake, Minidoka to Milner
16	12	4,130	.20	4,100	5	
16	13	4,130	.20	4,100	5	
17	13	4,130	.20	4,100	5	
17	14	4,130	.20	4,100	5	
18	14	4,130	.20	4,100	5	
18	15	4,130	.20	4,100	5	
17	16	4,130	.20	4,100	5	
18	16	4,130	.20	4,100	5	
17	17	4,130	.20	4,100	5	
17	18	4,150	.20	4,120	5	
17	19	4,190	2.0	4,160	6	Snake, Neeley to Minidoka
18	19	4,190	2.0	4,160	6	
18	20	4,190	2.0	4,160	6	
19	19	4,190	2.0	4,160	6	
19	20	4,190	2.0	4,160	6	
20	21	4,190	2.0	4,160	6	
20	22	4,190	2.0	4,160	6	
20	23	4,190	2.0	4,160	6	
20	24	4,200	2.0	4,170	6	
20	25	4,240	2.0	4,210	6	
19	26	4,355	.0446	4,325	7	Snake, near Blackfoot to
19	27	4,355	.0446	4,325	7	Neeley; Portneuf,
18	28	4,355	.0446	4,355	7	Pocatello to mouth
19	28	4,355	.0446	4,355	7	

TABLE 18.—*River-block locations, stages, riverbed or spring-outlet conductances, leakage-cutoff altitudes, and reach numbers*—Continued

Row	Column	River altitude	Conductance (leakance × river-block area) (feet squared per second)	Leakage-cutoff altitude	Reach Number	Reach Name
18	29	4,355	0.0446	4,355	7	Snake, near Blackfoot to Neeley; Portneuf, Pocatello to mouth (continued)
17	30	4,380	.0446	4,380	7	
18	30	4,355	.0446	4,355	7	
17	31	4,380	11.0	4,380	7	
18	31	4,355	11.0	4,355	7	
19	31	4,360	11.0	4,360	7	
20	31	4,370	11.0	4,370	7	
17	32	4,380	11.0	4,380	7	
18	32	4,380	11.0	4,380	7	
17	33	4,400	11.0	4,400	7	
17	34	4,440	10.0	4,410	8	Snake, at Blackfoot to near Blackfoot
17	35	4,475	10.0	4,445	8	
16	36	4,500	1.3	4,470	9	Snake, Shelley to at Blackfoot
16	37	4,530	1.3	4,500	9	
16	38	4,560	1.3	4,530	9	
15	39	4,600	1.3	4,570	9	
15	40	4,625	2.5	4,595	10	Snake, Lewisville to Shelley
14	41	4,700	2.5	4,670	10	
11	42	4,760	2.5	4,730	10	
12	42	4,750	2.5	4,720	10	
13	42	4,740	2.5	4,710	10	
10	43	4,770	25.0	4,740	11	Snake, Lorenzo to Lewisville
11	44	4,800	25.0	4,770	11	
12	45	4,860	2.0	4,830	12	Snake, Heise to Lorenzo
13	46	4,950	2.0	4,920	12	
14	46	4,980	2.0	4,950	12	
10	46	4,815	.7	4,785	13	Henrys Fork, Ashton to mouth
10	45	4,810	.7	4,780	13	
10	47	4,830	.7	4,800	13	
10	48	4,860	.7	4,830	13	
10	49	4,910	.7	4,880	13	
10	50	5,000	.7	4,970	13	
8	3	3,200	1.34	3,200	14	Salmon Falls
9	2	3,400	1.34	3,400	14	

Creek area, and the Big Lost River alluvial fan. Sedimentary rocks are thick along the Snake River upstream from Milner.

Rhyolite in layer 2 extends beyond the northeastern boundary of the plain. Sedimentary rocks in layer 2 are similar to those in layer 1; coarse-grained rocks predominate in upper reaches of the Snake River and Henrys Fork, and fine-grained rocks predominate along the lower Snake River and at the mouths of tributary drainage basins.

Quaternary basalt with relatively minor, interlayered, fine-grained sedimentary rocks is confined to the central part of the plain in layer 3. Basalt as much as 3,000 ft thick dominates layer 4 in the central part of the plain.

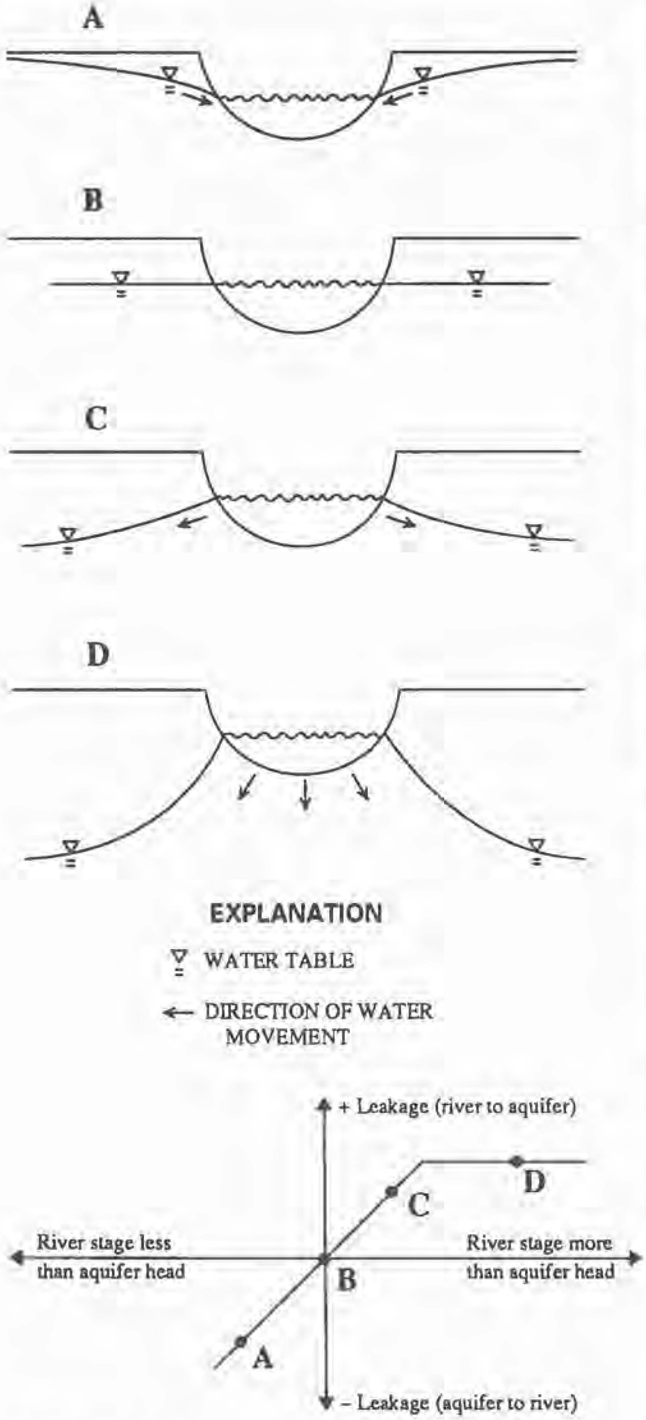


FIGURE 22.—Relations among aquifer head, river stage, and river leakage.

TRANSMISSIVITY, LEAKANCE, AND STORAGE COEFFICIENT

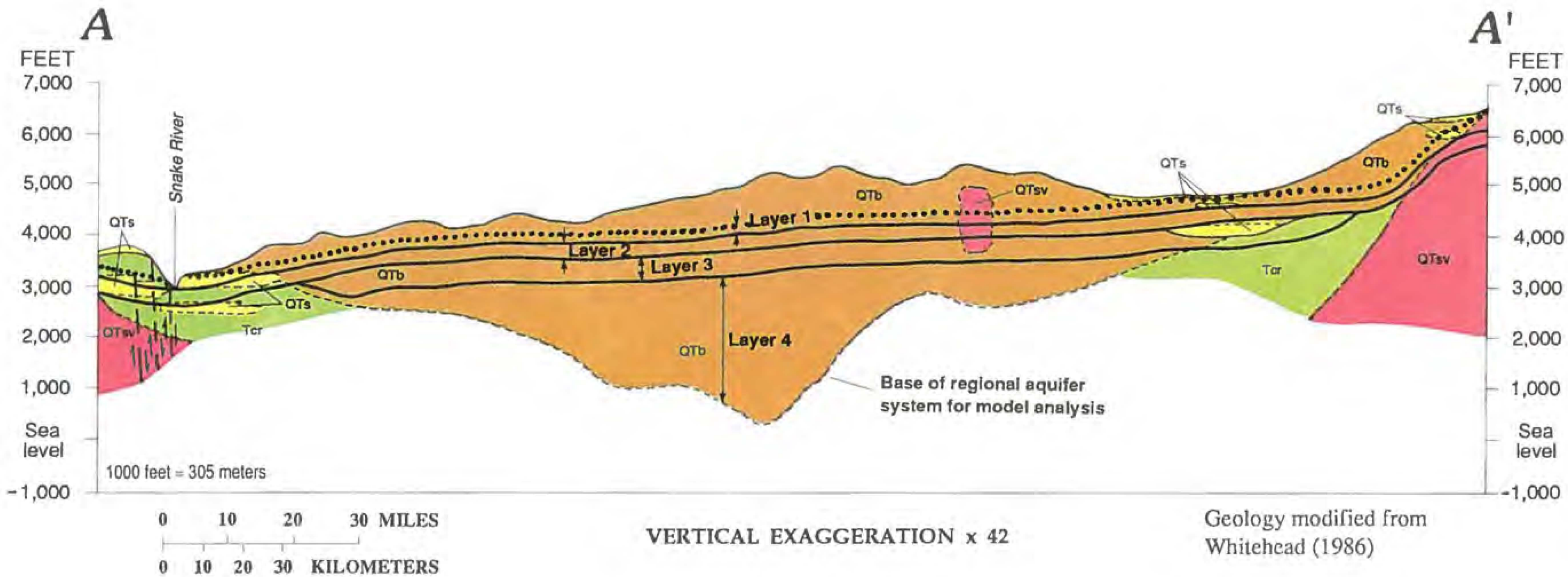
Model transmissivity was calculated for each active block using hydraulic conductivity for each rock type (table 19) distributed by zones across the plain (pl. 6). The thickness of each rock type in each layer is shown on plate 5. Hydraulic-conductivity values were calibrated (along with other model input parameters) to achieve an acceptable match between three-dimensional steady-state simulated heads and measured heads. Isotropic conditions were assumed for horizontal movement of water. Previous efforts to improve model fit using anisotropic transmissivities indicated little evidence for regional anisotropy in the ground-water system.

Hydraulic-conductivity values given in table 19 were used to calculate transmissivities of layers 1 and 2. Transmissivity values for each block in a layer were calculated by multiplying the thickness of each rock type for that block (pl. 5) by the rock hydraulic conductivity (table 19, pl. 6) and adding all the individual rock-type transmissivities to obtain the total layer transmissivity. During model calibration, hydraulic-conductivity values given in table 19 were reduced by one-third for layer 3 and two-thirds for layer 4 to account for decreasing hydraulic conductivity with depth. Average transmissivity values for layers 1 through 4 are shown on plate 6. The range in values of combined transmissivity for all layers in the three-dimensional model exceeded that for the two-dimensional model because of finer definition of aquifer properties and greater vertical resolution of head.

Vertical flow in the regional aquifer system was simulated as leakage between model layers. Vertical leakage was calculated using a leakance parameter, defined as the vertical hydraulic conductivity divided by the distance between vertically adjacent blocks. McDonald and Harbaugh (1988, p. 5–12) referred to this parameter by the Fortran variable name, Vcont. The leakance parameter was calculated for model input in the following manner:

1. Each model block was subdivided into identifiable rock-type subunits; for example, a block of layer 1 might consist of the following rock types and corresponding thicknesses:

Rock type	*Thickness (feet)*
Basalt	100
Sand and gravel	10
Sand	20
Clay and silt	20
Silicic volcanics	50
Total thickness	200



EXPLANATION

QTs	CLAY, SILT, SAND, AND GRAVEL	QUATERNARY AND TERTIARY
QTb	BASALT	
QTsv	RHYOLITE, LATITE, AND ANDESITE	
Tcr	BASALT	TERTIARY

CONTACT--Dashed where inferred

FAULT--Arrow shows direction of movement

WATER TABLE

FIGURE 23.—Section *A–A′* showing geology and model layers. (Line of section and map symbols shown on pl. 2)

TABLE 19.—*Hydraulic conductivities by rock type, model layers 1 and 2*

[Values in feet per second]

Zone	Rock type: Basalt ($\times10^{-4}$)	Sand and gravel ($\times10^{-4}$)	Sand ($\times10^{-4}$)	Clay and silt ($\times10^{-6}$)	Silicic volcanics (rhyolite) ($\times10^{-6}$)
1	0.052	11	0.11	2.3	7.5
2	5.5	90	.90	.75	7.5
3	550	73	.73	2.3	7.5
4	.9	17	.17	.75	7.5
5	803	110	1.1	2.3	7.5
6	2.4	47	.63	2.3	7.5
7	2.1	41	.41	2.3	7.5
8	56	140	1.4	.38	7.5
9	.75	7.5	.075	.75	7.5
10	5.7	110	1.1	.75	7.5
11	3.8	3.8	3.8	.38	7.5
12	23	75	.75	2.3	7.5
13	580	2,000	.10	.38	7.5
14	1,100	1,900	1.9	2.3	7.5
15	11	71	.71	.38	7.5
16	230	38	.38	2.3	7.5
17	61	330	.66	2.3	7.5
18	6	11	1.1	2.3	7.5
19	670	1,700	1.7	2.3	7.5
20	150	71	.71	2.3	7.5
21	590	83	.83	2.3	7.5
22	50	29	.29	.38	7.5
23	120	83	.83	2.3	7.5
24	440	83	.83	2.3	7.5
25	2.9	59	.59	2.3	7.5
26	200	48	.48	2.3	7.5
27	68	47	.62	2.3	7.5
28	3	58	.58	2.3	7.5
29	1.5	31	.31	.75	7.5
30	3.9	11	.11	.38	7.5
31	1.6	26	.26	.75	7.5
32	380	38	.38	2.3	7.5
33	420	210	2.1	2.3	7.5
34	250	300	.30	2.3	7.5
35	66	140	66	.38	7.5
36	600	1,500	600	7.5	7.5
37	15	15	.23	2.3	7.5
38	150	83	.83	3.8	7.5
39	120	18	.18	2.3	7.5
40	200	260	.26	2.3	7.5

2. Vertical hydraulic conductivity for each rock-type subunit was calculated by multiplying hydraulic-conductivity values in table 19 by a model-calibrated vertical-to-horizontal anisotropy factor (used across the entire modeled area):

Rock type	*Anisotropy factor* (K_v/K_h)
Basalt	0.01
Sand and gravel	.1
Sand	.05
Clay and silt	.05
Silicic volcanics	.01

3. Vertical hydraulic conductivity for the block was calculated using the rock-type subunit values in an equation for average hydraulic conductivity for a series of layers (Freeze and Cherry, 1979, p. 34):

$$K_z = \frac{D_t}{\sum_{i=1}^{n} D_i / K_i}, \tag{4}$$

where

K_z = block vertical hydraulic conductivity,
D_t = total block thickness,
D_i = rock-type subunit thickness,
K_i = rock-type subunit vertical hydraulic conductivity, and
n = number of rock-type subunits.

4. Leakance was calculated using a harmonic mean between vertically adjacent blocks:

$$L = \frac{2(K_{z1}K_{z2})}{(K_{z1}D_2) + (K_{z2}D_1)}, \tag{5}$$

where

L = leakance between blocks 1 and 2,
K_{z1} = block 1 vertical hydraulic conductivity,
K_{z2} = block 2 vertical hydraulic conductivity,
D_1 = block 1 thickness, and
D_2 = block 2 thickness.

The harmonic mean was used to calculate leakance between blocks because if either block were inactive (K_{z1}=0), the calculated leakance value would be zero and a no-flow boundary would exist. Average leakance values between model layers are shown on plate 7. Between layers 2 and 3 and layers 3 and 4, leakance was zero in some zones and a no-flow boundary was specified.

Storage-coefficient values were calculated for layer 1 using the distributions of rock types shown on plate 5 and the specific-yield values as follows: basalt, 0.05; sand and gravel, 0.20; sand, 0.20; silt and clay, 0.20;

FIGURE 24.—Hydraulic-conductivity zones and average storage coefficients, model layer 1.

and silicic volcanics, 0.05. These values of specific yield are consistent with the results from aquifer tests in unconfined sediments and basalts. The average storage coefficient for each zone in layer 1 is shown in figure 24. Below layer 1, all layers are considered to be confined aquifers and are assigned a storage coefficient of 0.0001.

STEADY-STATE SIMULATIONS

The primary objective of steady-state three-dimensional simulations was to calibrate aquifer transmissivity, leakance, and riverbed or spring-outlet conductance values such that the simulated heads reasonably matched heads measured in 1980. Three-dimensional modeling results generally were better than two-dimensional results because simulation of head changes with depth in recharge and discharge areas gave a more realistic representation of the regional ground-water flow. The approach used in steady-state three-dimensional simulation was to calculate recharge to the regional aquifer system and then to use this information (along with measured heads) as a basis for calibrating the aquifer parameters. Recharge for steady-state simulations was based on 1980 water-year data and was distributed to each block using the following expression:

$$RB_{(i,j)} = \frac{1}{AB_{(i,j)}}\left[\frac{1\text{ft}^3/\text{s}}{724\,\frac{\text{acre}-\text{ft}}{\text{yr}}}\left(\sum_{k=1}^{N} SW_k A_k\right)(i,j) + P_{(i,j)} + \Delta S_{(i,j)}\right], \quad (6)$$

where

$RB_{(i,j)}$ = recharge rate for block (i,j) in feet per second;

$AB_{(i,j)}$ = area of block (i,j), in square feet;

SW_k = recharge rate for irrigation area (k) (table 5), in feet per year;

A_k = total acreage for irrigation area (k) in block (i,j), in acres;

$P_{(i,j)}$ = recharge from precipitation in block (i,j) (table 12), in cubic feet per second; and

$\Delta S_{(i,j)}$ = change in storage per unit time in block (i,j), in cubic feet per second.

Snake River, Henrys Fork, and Salmon Falls Creek gains and losses were simulated using river blocks (fig. 21). Other stream and canal losses (table 10), tributary drainage-basin underflow (table 11), and average ground-water pumpage in water year 1980 (fig. 25) were simulated as recharge or discharge wells.

Mass-balance calculations for steady-state simulations are shown in table 16. Each category of model flux (wells, recharge, river leakage) includes both positive and negative values owing to the use of each flux category for various components of aquifer inflow and outflow. For example, wells were used to simulate underflow from tributary drainage basins and stream and canal losses, as well as outflow from irrigation pumpage.

A comparison of water-table contours based on simulated heads (layer 1) with contours based on measured water levels in 1980 is shown in figure 26. Generally, simulated and measured heads (and, therefore, direction of ground-water flow) are in close agreement in the central part of the plain. Differences between calculated and measured heads are significant along the margins of the plain; in the upper Camas Creek, Mud Lake, and Goose Creek areas; and near the Snake River from Milner to King Hill.

The difficulty in obtaining a good match between simulated and measured heads is due in part to major changes in aquifer properties over short distances. The large block size of the regional model (16 mi^2) also precludes simulation of small-scale, local variations in head, especially where head gradient is steep. In the Camas Creek area, gradients are steep and the basalt aquifer is thin; consequently, simulation was difficult. The Mud Lake and Big Lost River areas include shallow (perched) aquifers, which were not simulated. Ground-water levels in the Goose Creek area are declining as a result of pumping, and the nonequilibrium (transient) condition cannot be simulated accurately with the steady-state model. Near the Snake River from Milner to King Hill, changes in hydraulic conductivity over short distances cause local changes in ground-water levels that could not be simulated with the regional model.

Simulated head differences between layers 1 and 2 are shown in figure 27. Largest differences are in major recharge and discharge areas along the margins of the plain. In these areas, only layers 1 and 2 are active in the model (pl. 5). Reasonable matches of simulated and measured head changes with depth were achieved in the Rigby Fan area (near Idaho Falls), the Burley area, and the major discharge area in Jerome and Gooding Counties. Although increasing head with depth is indicated in the area northeast of American Falls Reservoir, most measured head differences between layers 1 and 2 are 10 to 20 ft rather than 5 ft, as simulated. Simulated and measured head changes with depth also were difficult to match in the Mud Lake and Big Lost River areas, probably because of the assumption of saturated flow in areas of perched water.

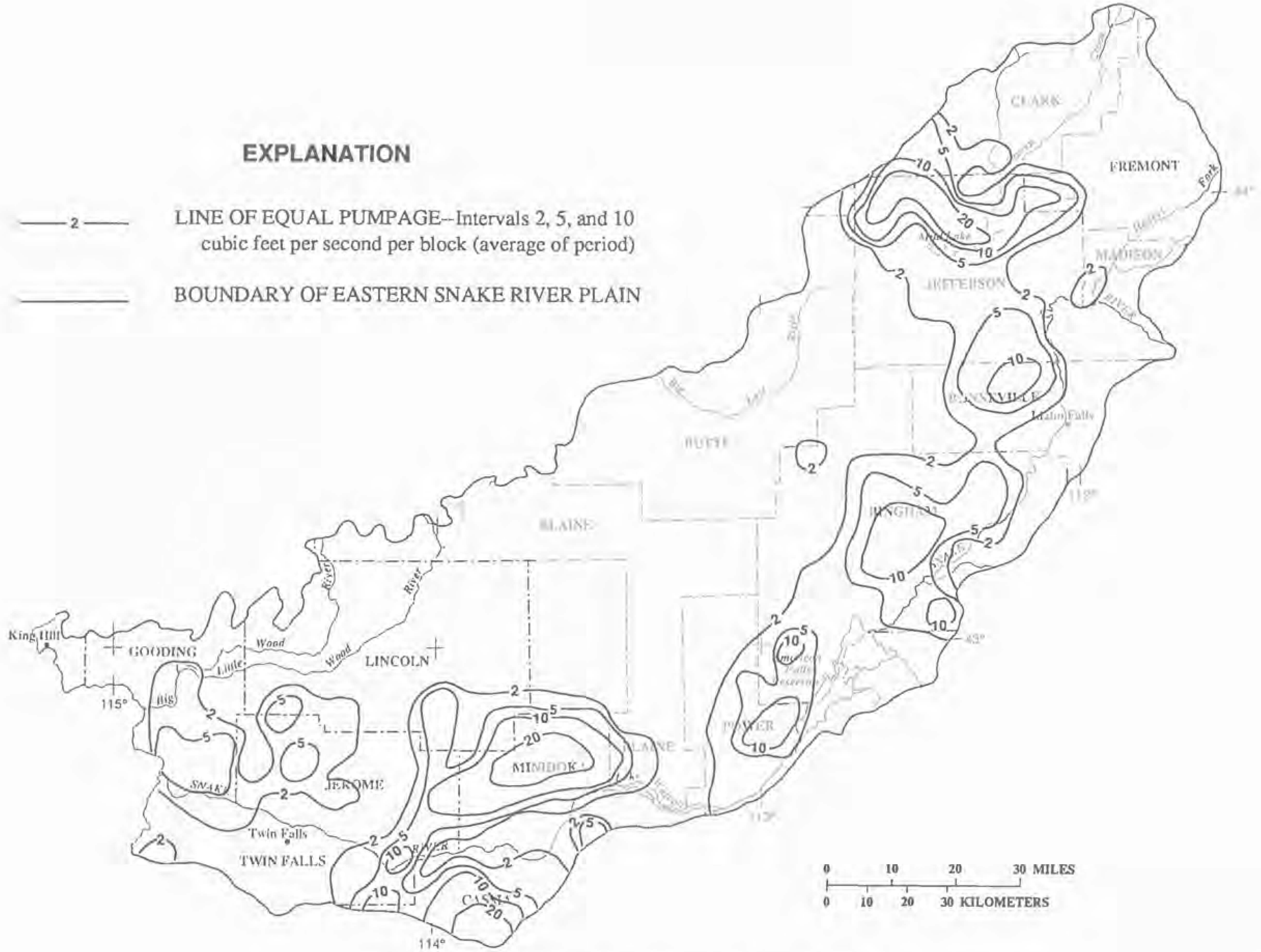


FIGURE 25.—Average ground-water pumpage, water year 1980.

EXPLANATION

— 4000 — WATER-TABLE CONTOUR--Shows altitude of the regional water table based on measured water levels, March 1980 (Lindholm and others, 1983). Intervals 100 and 200 feet. Datum is sea level

— — 3800 — — WATER-TABLE CONTOUR--Shows altitude of the regional water table based on simulated water levels in layer 1. Interval 100 feet. Datum is sea level

BOUNDARY OF EASTERN SNAKE RIVER PLAIN

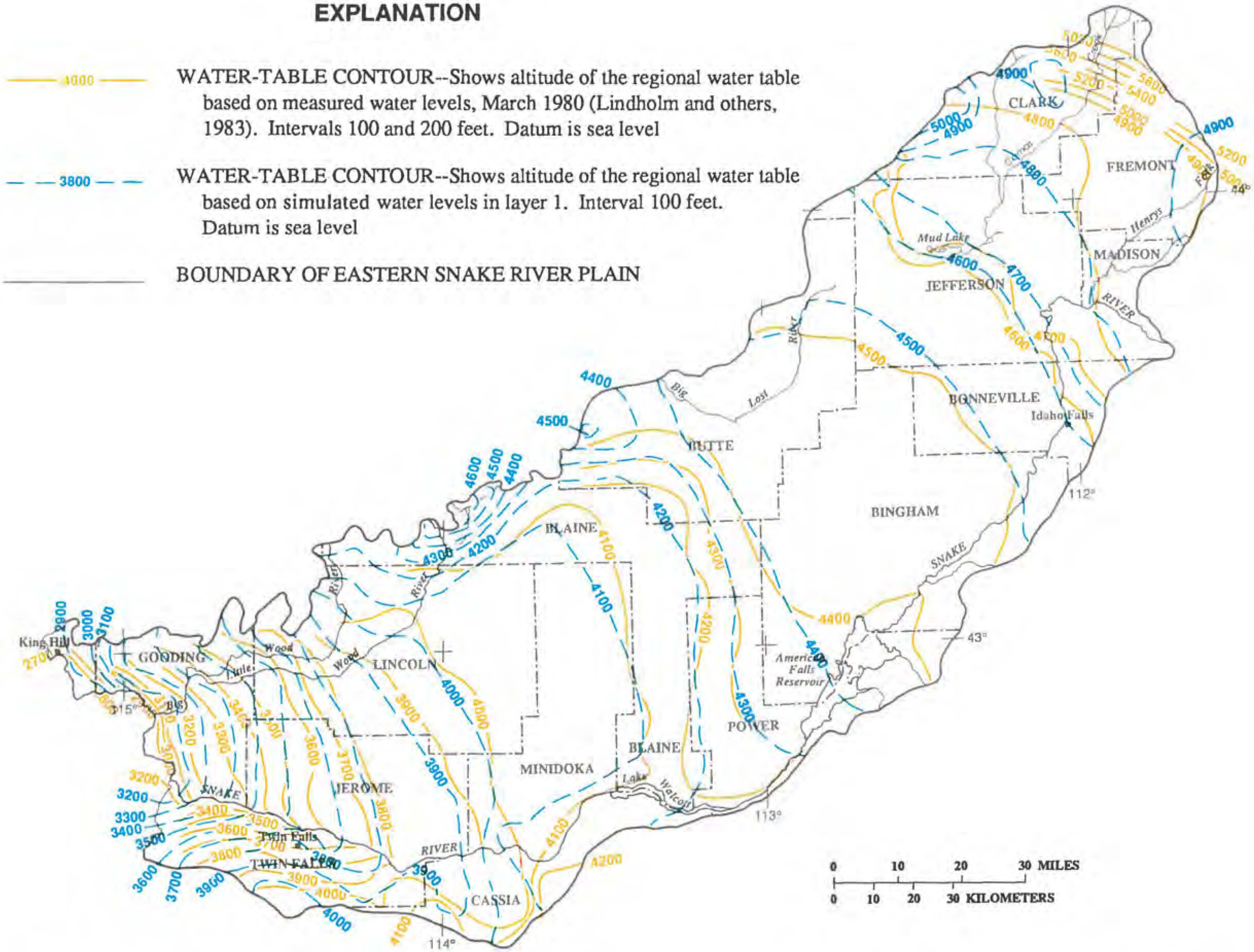


FIGURE 26.—Configuration of the water table, March 1980, based on measured water levels, and configuration based on simulated heads in layer 1.

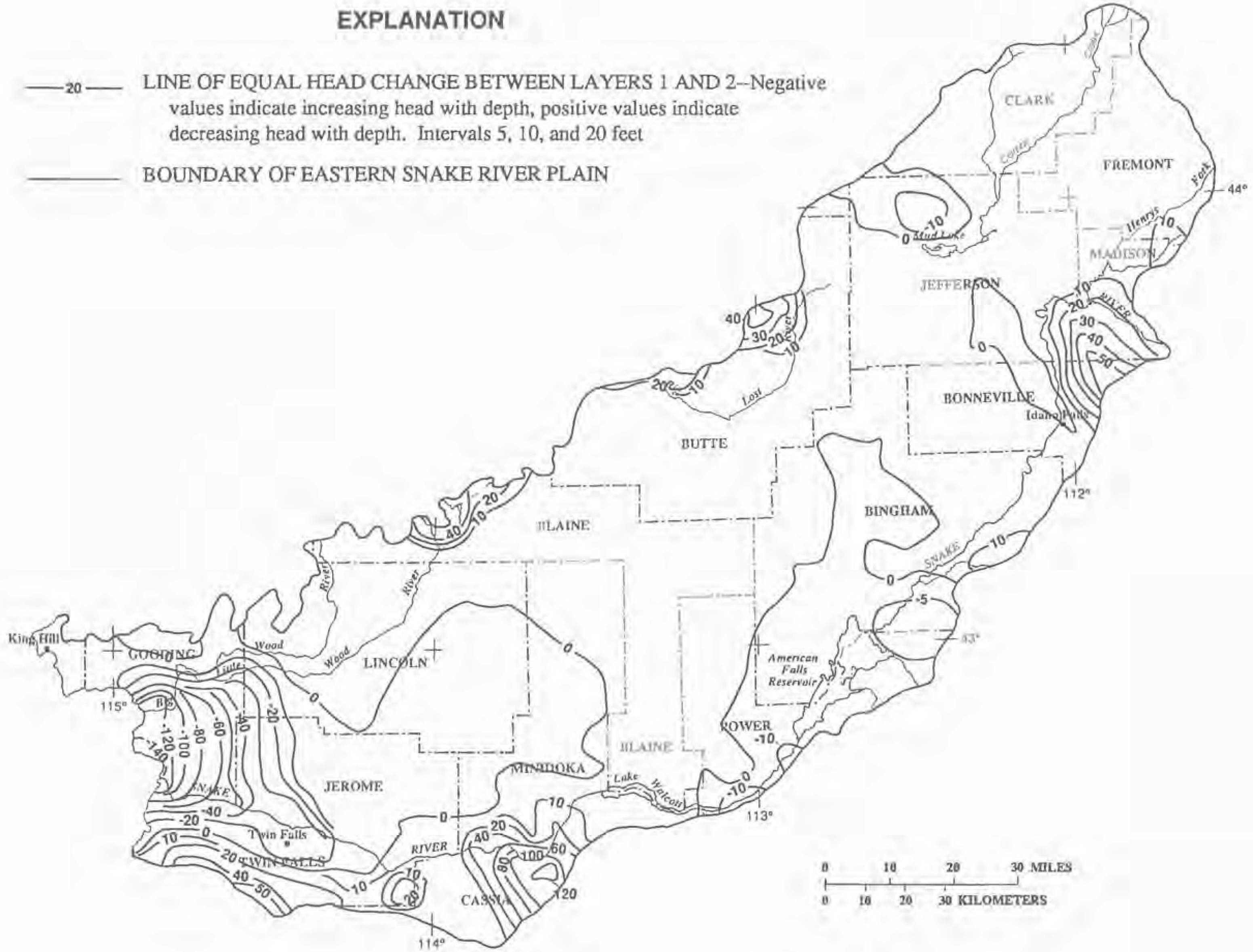


FIGURE 27.—Simulated head differences between model layers 1 and 2.

TABLE 20.—*Comparison of published hydraulic-conductivity values with those used in this study*

[Values in feet per second]

Rock type	Hydraulic conductivity (Freeze and Cherry, 1979, p. 29)	Table 19 (this report)
Basalt	$0.07–5\times10^{-7}$	$0.11–5.2\times10^{-6}$
Sand and gravel	$3–1\times10^{-5}$	$0.2–7.5\times10^{-4}$
Sand	$0.03–3\times10^{-7}$	$0.06–7.5\times10^{-6}$
Clay and silt	$5\times10^{-5}–3\times10^{-9}$	$2.3\times10^{-6}–3.8\times10^{-7}$
Silicic volcanics	$6\times10^{-4}–2\times10^{-8}$	7.5×10^{-6}

The steady-state model was calibrated by adjusting zonal hydraulic-conductivity values (table 19) and river-block conductances (table 18) within reasonable ranges. Rock-type hydraulic-conductivity values were adjusted to achieve an acceptable match between steady-state simulated and measured water levels and spring discharges, and estimated river-leakage values. Although several previous investigators (Mundorff and others, 1964; Norvitch and others, 1969; Newton, 1978) reported transmissivity distributions for the regional aquifer, areal hydraulic-conductivity values shown in table 19 are the first to be reported for this aquifer. To demonstrate the reasonableness of model values, a comparison between published hydraulic-conductivity ranges (Freeze and Cherry, 1979, p. 29) and ranges used in this study for transmissivity and leakance calculations (table 19) is presented in table 20. Lowest values of hydraulic conductivity are along the margins of the plain and in the Mud Lake area, where basalt is interlayered with sedimentary rocks. Highest values are in the central part of the plain, where volcanic activity is most recent and sediment interbeds in the basalt are few.

Riverbed or spring-outlet conductance values were adjusted to provide a reasonable match between simulated and measured river leakage or spring discharges. Conductance values along the Snake River are low along the Hagerman-to-King Hill and Minidoka-to-Milner reaches, and in the vicinity of American Falls Reservoir, where low-hydraulic-conductivity lacustrine deposits predominate. Conductance values are high along the Milner-to-Kimberly and Buhl-to-Hagerman reaches, where highly transmissive pillow lavas fill ancestral Snake River canyons and control the locations of large springs.

TRANSIENT SIMULATIONS

The objective of three-dimensional transient simulations was to evaluate the ability of the model to simulate long-term changes in the regional aquifer system. Evaluation consisted of comparing model results with 1890–1980 measured changes in water levels and ground-water discharges. Initial head conditions for transient simulations (fig. 28) were derived from a steady-state simulation of estimated preirrigation hydrologic conditions. Input for the preirrigation simulation included recharge from precipitation (fig. 9), stream losses (table 10), and riverbed or spring-outlet conductances (table 18). Preirrigation underflow was estimated by adding flow to estimated underflow values in table 11 to compensate for upstream consumptive use.

The general configurations of the simulated preirrigation water table (fig. 28) and the March 1980 water table (fig. 26) are similar. As might be expected, the general direction of ground-water flow, inferred to be perpendicular to equipotential lines, is the same on both maps. In nearly all places, the preirrigation water table is lower than the water table in 1980. However, near the mouth of the Big Lost River valley, the preirrigation water table is higher than the March 1980 water table owing to greater tributary drainage-basin underflow before irrigation in the valley began.

In places, the simulated preirrigation water table was more than 200 ft below the altitude of the March 1980 water table. Preirrigation water levels were below the bottom of layer 1 in the steady-state three-dimensional model. Therefore, layers 1 and 2 in the steady-state model were combined to form a three-layer model for transient simulations. In the northeastern part of the modeled area (upper Camas Creek area), the simulated preirrigation water table was more than 500 ft below the altitude of the March 1980 water table. In that area, initial heads in the transient model were modified so that they were 100 ft above the bottom of the layer. The three-layer model with the modified initial conditions was numerically stable during all transient simulations.

Eighteen 5-year stress periods (time intervals during which all external stresses are assumed to be constant) were used to simulate transient hydrologic conditions from 1891 to 1980. Recharge from surface-water irrigation for each block in the top layer of the model was calculated for each stress period using the recharge rates in table 6 and the irrigated acreage maps on plate 3. Total ground-water recharge from surface-water irrigation for each of the 5-year intervals is shown in table 7. Recharge from surface-water irrigation and precipitation for the periods 1896–1900, 1926–30, and 1976–80 is shown on plate 8. As indicated in table 7, recharge from surface-water irrigation from 1891 to 1925 increased significantly but has remained relatively stable from 1926 to 1980. Average annual underflow from tributary drainage

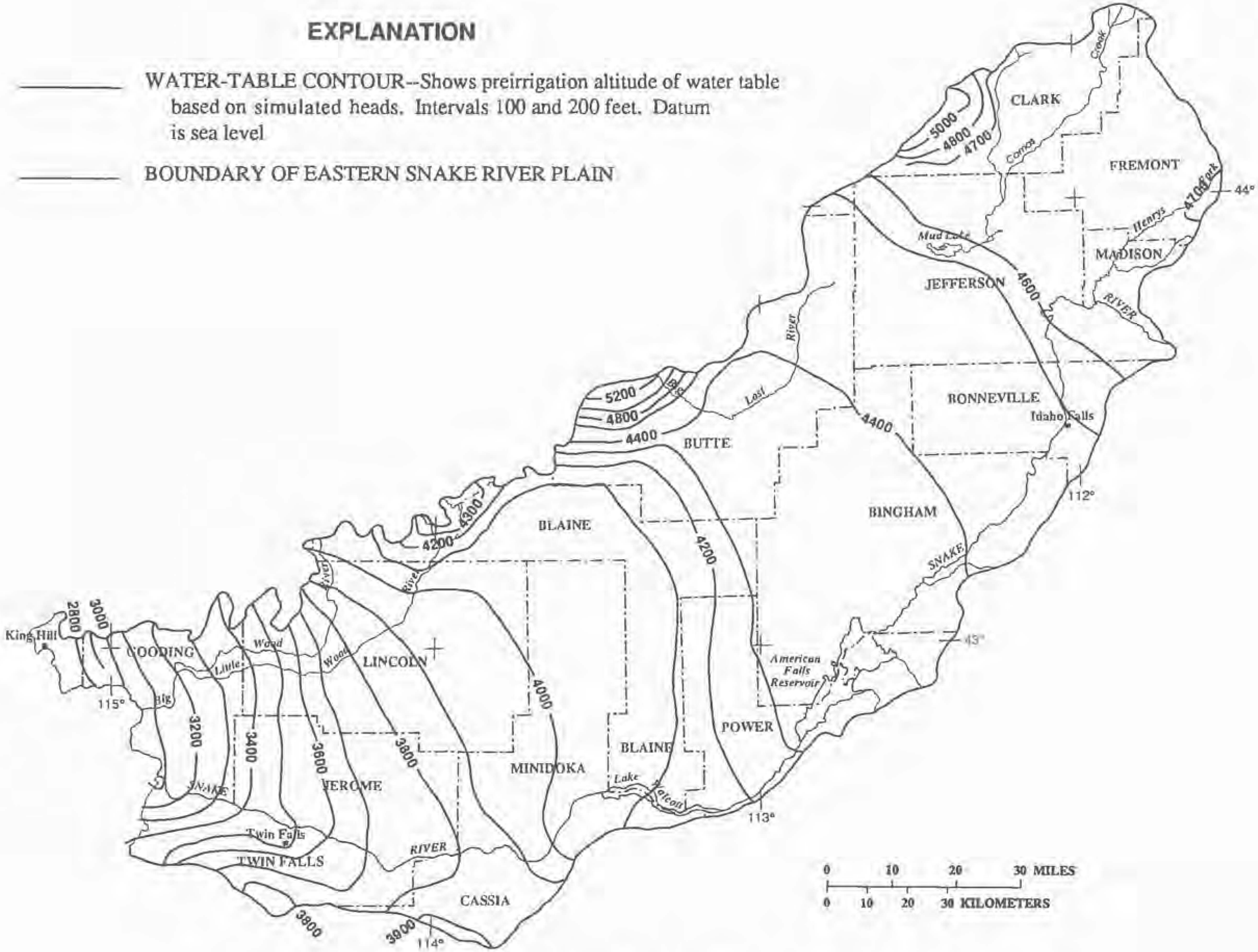


FIGURE 28.—Configuration of the preirrigation steady-state water table based on simulated heads in layer 1.

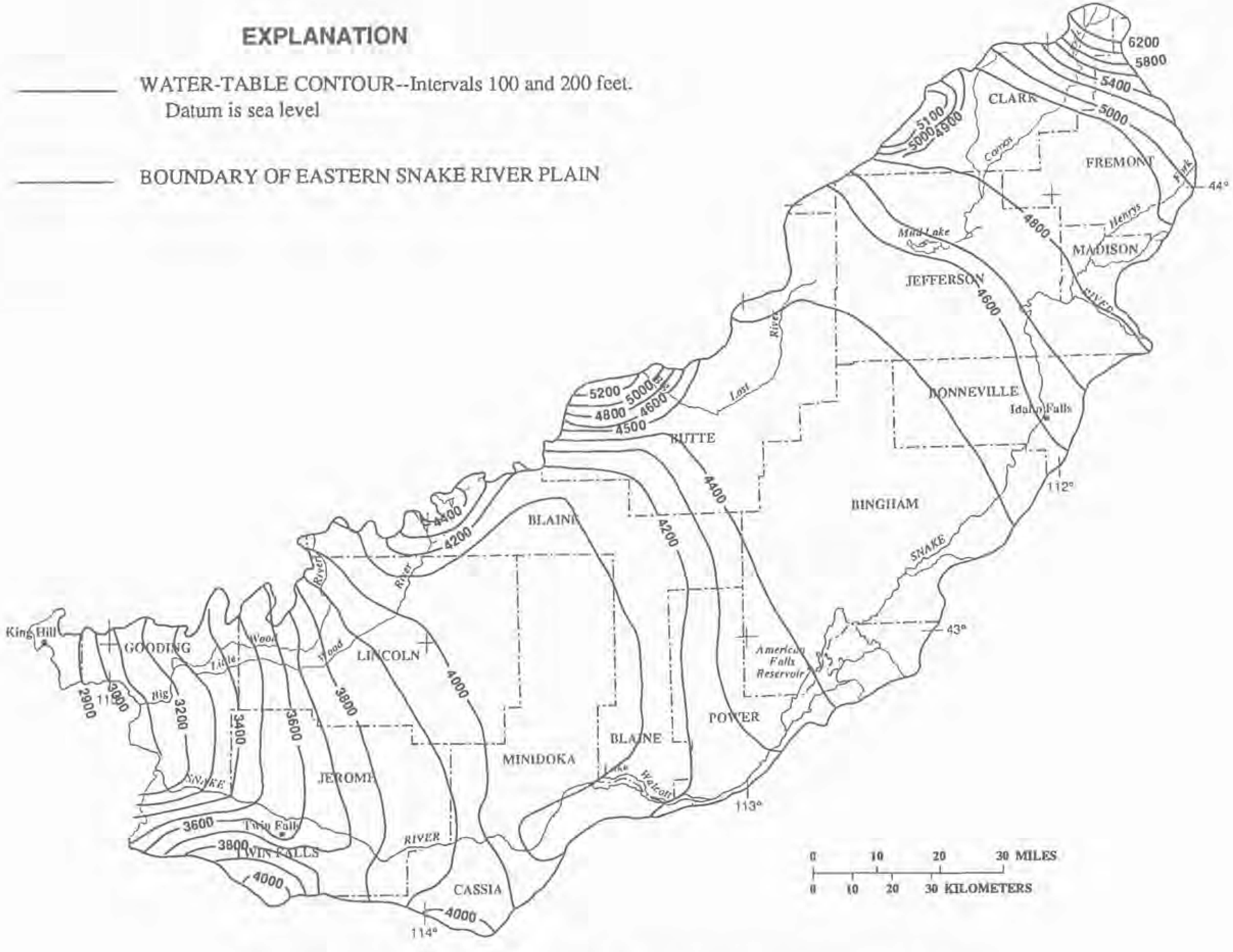


FIGURE 29.—Configuration of the water table in 1930 based on simulated heads in layer 1.

basins for the period 1891–1910 was calculated using basin-yield equations. Average annual underflow, stream losses, and precipitation from 1911 to 1980 (table 13) were estimated from measurements for different periods of record. Average ground-water pumpage was calculated using irrigated acreage maps (pl. 3) and estimated crop consumptive irrigation requirements (table 2) and is shown on plate 9 for the periods 1951–60, 1961–70, and 1971–80.

The configuration of the water table in 1930, based on simulated heads in layer 1, is shown in figure 29. Comparison of figure 29 with plate 4 indicates that water-level changes from 1930 to 1980 were small relative to changes that took place from preirrigation (fig. 28) to 1930. Changes in head from simulated preirrigation conditions (fig. 28) to simulated conditions in 1950 are shown in figure 30. Heads in the central part of the plain increased about 50 to 100 ft. Along the southern boundary of the plain near Twin Falls, simulated heads increased as much as 280 ft. Large increases in head also were simulated in the northeastern part of the plain (above Mud Lake) and may be due, in part, to the initial head conditions used in this part of the model. Although heads increased in most of the plain, declines were simulated in the region near the mouth of the Big Lost River. Declines likely were caused by decreases in underflow and river infiltration owing to upstream consumptive use of water for irrigation from 1890 to 1950 in the Big Lost River drainage area.

Changes in the water table from 1950 to 1980 (fig. 31) were generally smaller than those from preirrigation to 1950 (fig. 30). The model results indicate some increases in head along the boundary of the plain since 1950 and some declines in pumping areas, such as in Jefferson, Bonneville, Power, Minidoka, and Cassia Counties.

Comparisons of simulated and measured long-term head changes reported by Mundorff and others (1964) are shown in table 21. Although these data are for the southwestern end of the study area and cannot be used as an indicator of changes elsewhere, the agreement between measured and simulated head changes is generally good.

Changes in simulated head are due primarily to changes in input values of recharge, underflow, and pumpage that were varied with time to simulate changes in inflow and outflow. The simulated changes in inflow and outflow during the calibrated transient simulation are shown in table 22. The largest flux components are (1) recharge from surface-water irrigation, precipitation, and underflow; and (2) river losses and gains. Major changes in hydrologic conditions from preirrigation to 1950 were increased ground-water recharge and discharge as spring flow. After 1950, changes in recharge and discharge were smaller, and the net change was a decrease in ground-water storage, in part owing to a steady increase in ground-water withdrawals. Simulated changes in storage, river inflow, and river outflow approximated measured changes.

SENSITIVITY ANALYSIS

Modeling results discussed thus far represent the calibrated three-dimensional simulations using the described estimates of aquifer properties and fluxes. To determine model response to changes in various aquifer properties and fluxes, model runs were made for comparison with the calibrated run. The model thus was tested for sensitivity to changes in the input values of transmissivity, storage, leakance, recharge, riverbed or spring-outlet conductance, ground-water pumpage, and tributary drainage-basin underflow. Each model parameter was increased and decreased by 50 percent, with the exception of leakance, which was increased by a factor of 10 and decreased by a factor of 0.1. Parameter changes were applied uniformly across the entire modeled area.

Results of the sensitivity analysis are presented as a series of ground-water-level and river-gain/loss hydrographs in figures 32–46 (see following "References Cited" section). Measured, calibrated, and sensitivity-run hydrographs are included in each figure for comparison. Differences between measured and simulated heads for representative wells across the eastern Snake River Plain are presented in table 23, along with the square root of the average sum of squares difference (a measure of the absolute deviations from measured heads) for each sensitivity run, average sensitivity for the entire model, and average long-term head change.

Changes in model response owing to imposed changes in transmissivity are shown in figures 32 and 33. Hydrographs based on measured ground-water levels generally begin about 1950, after the major ground-water-level increases resulting from surface-water irrigation. Therefore, comparison of simulated head changes (table 21) with changes based on field observations (reported by Mundorff and others, 1964) is important in confirming approximate agreement of model results with actual field data. Simulated heads determined by increasing and decreasing estimates of transmissivity bracket most measured heads. When transmissivity was decreased by 50 percent, simulated heads averaged 34 ft higher than those in the calibrated run (table 23). When transmissivity was increased by 50 percent, simulated heads declined an average of

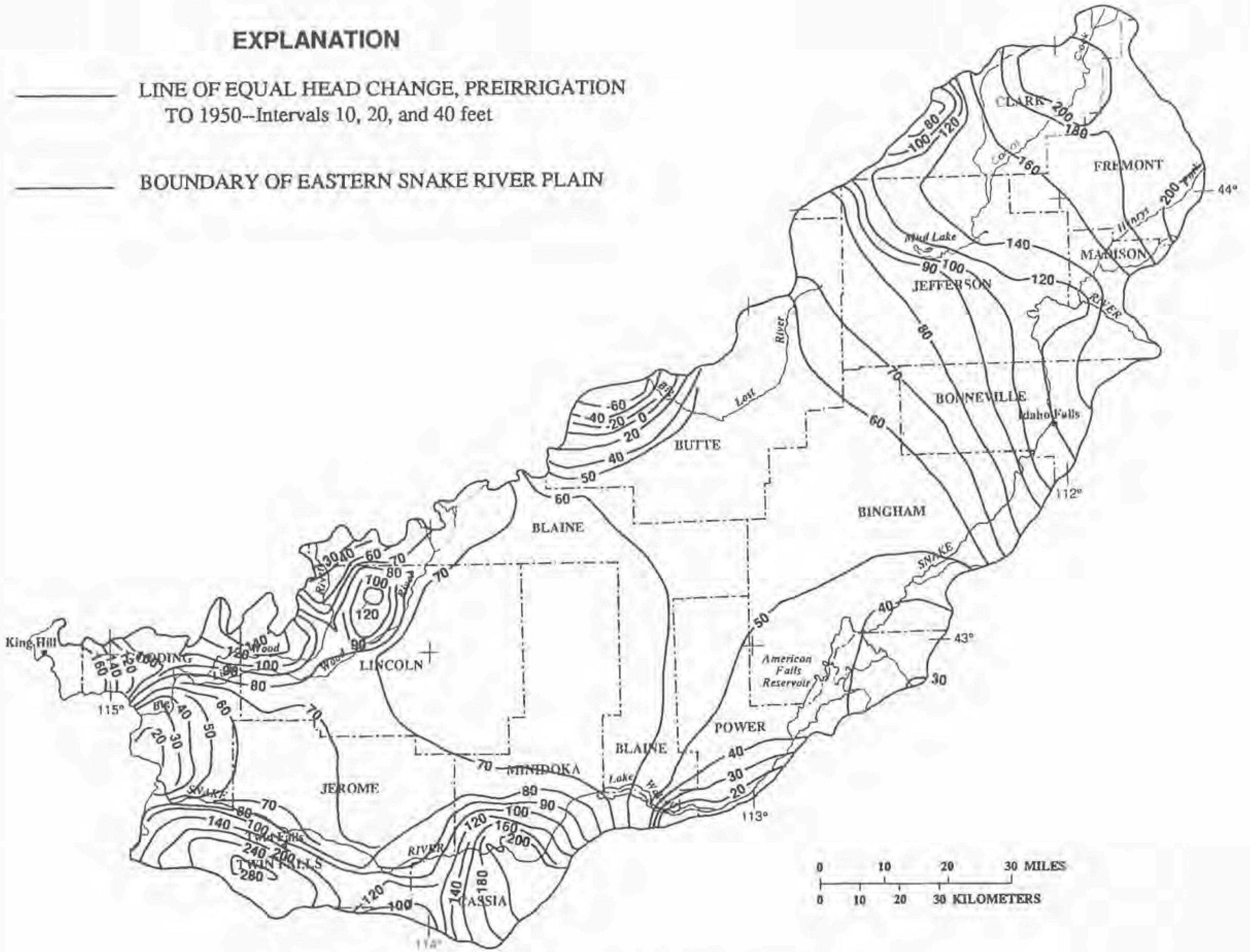


FIGURE 30.—Simulated changes in the water table, preirrigation to 1950.

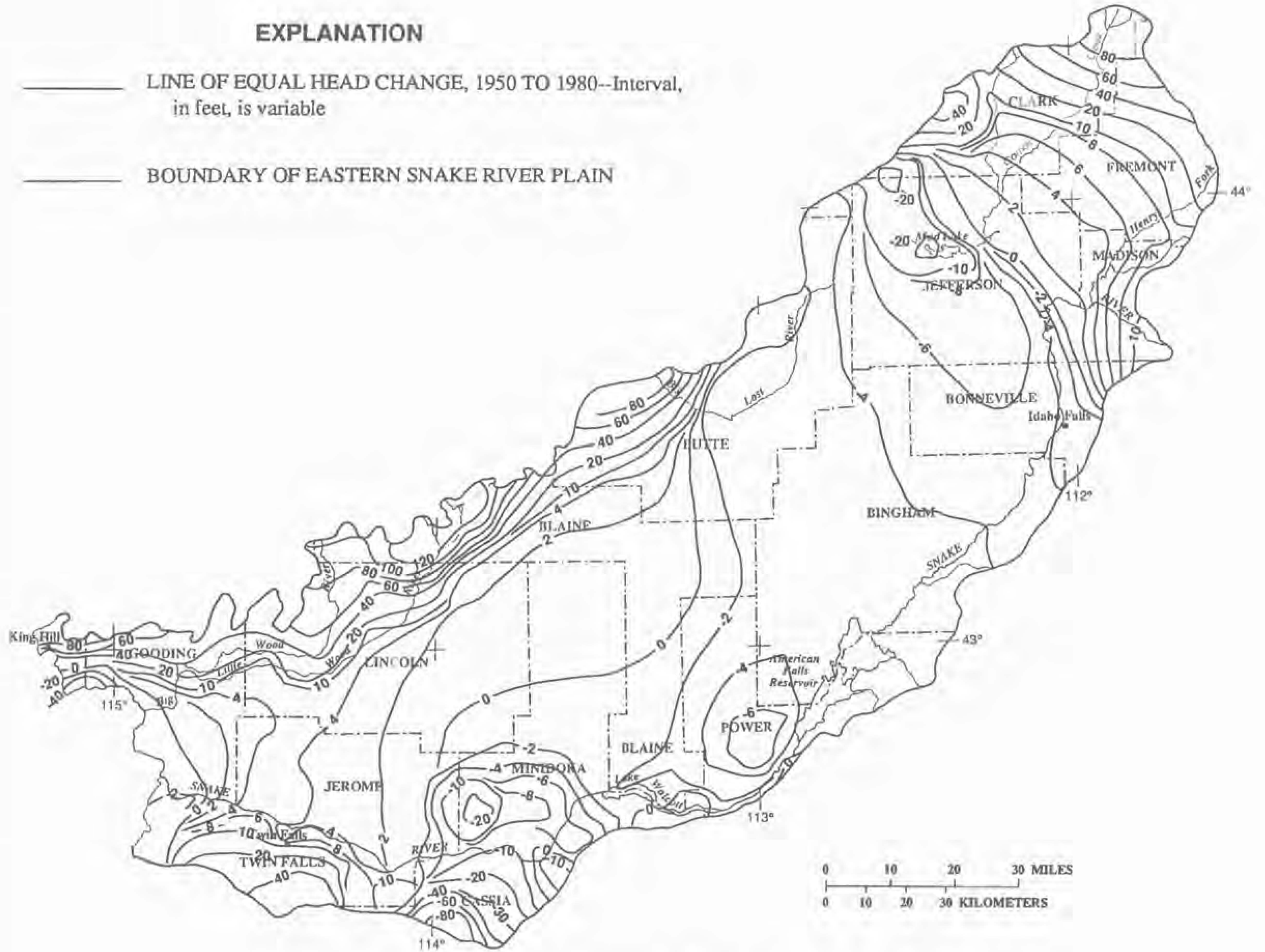


FIGURE 31.—Simulated changes in the water table, 1950–80.

TABLE 21.—*Reported and simulated head changes*

[Values in feet]

Well location	Measurement years and reported head change (Mundorff and others, 1964, p. 162)		Span of years for simulated head change	
6S-13E- 6DD	1901 and 1959	80	1900–60	118
8S-15E-28BA	1909 and 1959	32	1910–60	35
7S-15E-33	1907 and 1959	40	1905–60	42
7S-15E- 8	1907 and 1959	25	1905–60	35
5S-15E-31 or 32	1907 and 1959	35	1905–60	74
6S-17E- 2AB	1890 and 1952	70	1890–1950	87
8S-17E-19BB1	1907 and 1954	44	1905–55	66
8S-18E-15CC	1907 and 1959	118	1905–60	67
4S-19E-26DA1	1913 and 1957	19	1915–55	52
9S-19E-15AC	1907 and 1959	92	1905–60	70
9S-19E-26	1912 and 1959	62	1910–60	66
6S-20E-15DA	1901 and 1959	141	1900–60	65
7S-23E- 5	1901 and 1959	55	1900–60	61
8S-25E- 1CB1	1901 and 1959	190	1900–60	58
9S-24E-29AA1	1905 and 1951	42	1905–50	101
Average head change		70		66

about 23 ft (table 23). With higher transmissivities, lower average heads and smaller water-table gradients are needed to move the same amount of water to the major spring-discharge areas near American Falls Reservoir and to the Snake River from Milner to King Hill.

Hydrographs for well 8S-14E-16CBB1 (fig. 32) show the opposite relation between simulated heads and transmissivity. The well is in the extreme southwestern part of the study area, about 1 mi from the Snake River and major springs. When transmissivity upgradient from well 8S-14E-16CBB1 was increased, water levels in the well rose; when transmissivity was decreased, water levels declined. This relation is due to the increased volume of flow toward the southwest when transmissivities were increased. Although increasing transmissivity resulted in lower heads regionally, the model indicated that heads near the southwestern spring-discharge area would rise. Decreasing transmissivity caused heads near the springs to decline.

Model sensitivity to changes in transmissivity with respect to ground-water flux to and from the Snake River is shown in figure 33. Opposing effects are observed in the response curves for the Milner-to-King Hill and Blackfoot-to-Neeley reaches. When transmissivity was increased, ground-water discharge to the Milner-to-King Hill reach increased and discharge to the Blackfoot-to-Neeley reach decreased. When transmissivity was decreased, discharge to the Milner-to-King Hill reach decreased and discharge to the Blackfoot-to-Neeley reach remained essentially the same. The sensitivity of simulated aquifer heads to transmissivity changes is nonsymmetric. Head increases were generally greater when transmissivity was decreased by 50 percent than when they were increased by 50 percent.

Results of model-sensitivity analysis indicate that decreasing transmissivity produced larger head deviations than increasing transmissivity. Both decreasing and increasing transmissivity resulted in larger absolute deviations than those in the calibrated run.

Changes in simulated ground-water levels and flux through the ground-water system in response to imposed changes in storage coefficient and leakance were relatively small (figs. 34–37). Differences in simulated head in response to imposed changes in storage coefficients were most evident when head changes were rapid owing to rapidly changing fluxes (1890–1930). Differences became smaller as hydrologic conditions approached equilibrium (1930–1980). This same result was observed on ground-water-discharge hydrographs, where differences were most pronounced during periods of changing flux. Heads were higher by an average of 2 ft (table 23) when storage coefficient was decreased by 50 percent; heads were lower by an average of 2 ft when storage coefficient was increased by 50 percent.

TABLE 22.—*Mass balance for the calibrated three-dimensional transient simulation*

[Values in cubic feet per second]

Simulation period	Inflow				Outflow		
	Change in ground-water storage	Underflow	Recharge from irrigation and precipitation	River losses	Change in ground-water storage	Pumpage[1]	River gains
1891–95	50	2,300	1,980	3,180	1,400	0	6,100
1896–1900	30	2,300	2,980	3,180	1,960	0	6,530
1901–05	20	2,300	4,040	2,970	2,070	0	7,250
1906–10	10	2,300	4,640	2,480	1,660	0	7,770
1911–15	0	2,530	6,400	2,010	2,180	0	8,760
1916–20	0	2,530	6,600	1,680	1,550	0	9,250
1921–25	50	2,240	7,980	1,480	1,640	0	10,090
1926–30	40	2,240	7,270	1,420	800	0	10,150
1931–35	500	1,710	6,620	1,550	370	0	10,020
1936–40	50	2,010	7,550	1,430	710	0	10,330
1941–45	10	2,660	7,310	1,390	790	0	10,560
1946–50	10	2,480	7,850	1,310	730	0	10,900
1951–55	160	3,000	8,310	1,260	680	870	11,180
1956–60	50	2,900	8,550	1,190	500	870	11,300
1961–65	500	3,050	7,670	1,320	290	1,370	10,910
1966–70	30	3,430	8,280	1,290	580	1,370	11,070
1971–75	70	4,060	9,280	1,110	1,170	1,980	11,360
1976–80	1,010	3,110	7,320	1,310	60	1,980	10,710

[1]Use of ground water for irrigation increased rapidly after 1945, although the irrigation maps shown on plate 3 indicate no ground-water irrigation in 1945. Therefore, the author estimated pumpage during the 1951–55 stress period on the basis of the 1959 irrigation map.

When storage coefficient was small, ground-water levels and discharge responded more rapidly to changes in flux, as shown in hydrographs for the period 1890–1930 (figs. 34, 35). When storage coefficient was large, the aquifer was less responsive to changes in flux in discharge areas (Milner to King Hill, fig. 35) than in recharge areas, such as the losing Snake River reach from Heise to Blackfoot. These results are due to the increased lag time for water-level changes when storage coefficient was increased.

Simulated ground-water-level changes in response to imposed changes in leakance were small, averaging 1 ft or less (fig. 36). Generally, heads increased about 1 ft when leakance was multiplied by 0.1, and heads declined about 0.5 ft when leakance was multiplied by 10. Ground-water flux remained essentially unchanged when leakance was decreased or increased (fig. 37). Model insensitivity to changes in leakance is due to the thickness of the upper model layer (500 ft) and does not imply that there is no vertical movement of water within a single model layer or from one model layer to another. The square root of the average sum of squares difference was the same or nearly the same for the calibrated model run as it was for tested changes in storage coefficient and leakance; the model was relatively insensitive to changes in these parameters.

Changes in simulated ground-water levels and flux in response to imposed changes in model recharge are shown in figures 38 and 39. Generally, water levels were higher and ground-water flux was greater (except for the Heise-to-Blackfoot reach) when recharge was increased. When recharge was increased 50 percent, water levels rose an average of 26 ft; when recharge was decreased the same amount, heads declined an average of 32 ft. As was true for transmissivity, a 50-percent change in recharge brackets most of the measured water-level hydrographs and also brackets most of the measured ground-water-flux hydrographs. The losing reach of the Snake River from Heise to Blackfoot (fig. 39) lost more water when recharge was decreased. When aquifer heads declined in response to reduced recharge, river leakage to the aquifer increased; the opposite was true in gaining reaches. The fact that absolute deviations for increased and decreased

TABLE 23.—*Differences between measured and simulated heads*

[Values in feet; NA, not applicable]

Model run	Observation well number 3N-29E-14ADD1	8S-14E-16CBB1	5N-34E-9BDA1	8S-19E-5DAB1	7N-38E-23DBA1	4S-24E-6BBC1	Average head difference for observation wells	Average head difference for calibrated and model runs	$\sqrt{\frac{(\text{Difference})^2}{\text{number of comparisons}}}$	Average sensitivity for all simulated heads	Average head changes for wells shown in table 21
Calibrated model	-19	-55	-2	-37	18	1	-17	NA	32	NA	66
Transmissivity × 0.5	7	-72	37	-2	94	44	17	33.58	56	68	96
Transmissivity × 1.5	-39	-38	-30	-60	-31	-28	-40	-22.52	41	-39	51
Storage coefficient × 0.5	-18	-49	-1	-34	20	4	-15	1.70	31	2	67
Storage coefficient × 1.5	-20	-50	-3	-40	15	-3	-19	-2.19	33	-4	63
Aquifer leakance × 0.1	-18	-52	-2	-36	19	4	-16	1.01	32	3	67
Aquifer leakance × 10	-20	-52	-1	-37	16	0	-18	-.55	32	-1	66
Recharge × 0.5	-43	-60	-33	-80	-28	-40	-49	-32.14	53	-44	33
Recharge × 1.5	1	-44	24	-4	58	32	9	26.10	37	36	93
River conductance × 0.5	-4	-12	8	-13	19	17	1	18.31	14	15	84
River conductance × 1.5	-24	-70	-4	-48	19	-5	-24	-6.54	40	-5	58
Ground-water pumpage × 0	-12	-54	7	-23	24	19	-8	9.31	30	17	72
Ground-water pumpage × 0.5	-16	-54	3	-30	21	10	-12	4.76	31	9	69
Ground-water pumpage × 1.5	-23	-54	-6	-44	15	-9	-22	-4.92	34	-9	64
Boundary flux × 0.5	-30	-54	-14	-47	0	-15	-29	-11.47	36	-28	61
Boundary flux × 1.5	-8	-54	10	-27	35	16	-6	10.96	32	25	71

recharge were larger than the calibrated deviations (table 23) indicates that a closer comparison of simulated and measured aquifer heads can be achieved by further refinement of input data.

Changes in model response owing to imposed changes in riverbed or spring-outlet conductances are shown in figures 40 and 41. Ground-water levels averaged 18 ft higher when conductances were decreased by 50 percent and averaged 7 ft lower when conductances were increased by 50 percent. Effects of changes in conductance on water levels are dependent on proximity of a well to the major ground-water discharge area between Milner and King Hill. Simulated head in well 8S-14E-16CBB1 (closest to the discharge area) increased more than 40 ft when riverbed or spring-outlet conductance was decreased, whereas head in well 7N-38E-23DBA1 (farthest from the spring area) increased only about 1 ft after 1960.

Total flux to and from the aquifer in all Snake River reaches except Neeley to Milner increased when riverbed or spring-outlet conductance was increased. In the Neeley-to-Milner reach (fig. 41), head changes resulting from downstream changes in ground-water discharge were greater than changes resulting from local flux changes. As a result, fluxes in the Neeley-to-Milner reach were larger when conductances were reduced and smaller when conductances were increased.

Hydrographs for wells 3N-29E-14ADD1, 5N-34E-9BDA1, and 4S-24E-6BBC1 show a reversal of head relations between increased and decreased riverbed or spring-outlet conductances. This reversal was due to initial head conditions used at the beginning of the simulation. Initial heads for all transient simulations were calculated using parameters from the calibrated model run. Therefore, initial heads were not in equilibrium with the changed parameter (in this case, riverbed or spring-outlet conductance) and, at the beginning of a transient simulation, heads changed in response to changes in both flux and initial head conditions. In the northeastern part of the aquifer, heads declined when conductance was decreased, whereas closer to the major springs, heads rose. As simulation proceeded, heads rose in the entire modeled area and, eventually, the hydrographs crossed.

Differences between measured and simulated heads (table 23) were smaller when riverbed or spring-outlet conductance was decreased by 50 percent than when conductance was increased by 50 percent. The smallest difference between measured and simulated heads for all the model runs was achieved when conductance was decreased by 50 percent. This reduction in model error is reasonable because riverbed or spring-outlet conductance values were originally calibrated during steady-state simulation of the four-layer model. Therefore, conductance values (defined in equation 3) should be adjusted for the increase in thickness of the upper layer in transient simulations. Because thickness of the upper layer was increased from 200 to 500 ft, conductances should be reduced to 40 percent of the steady-state values.

The reduction in riverbed or spring-outlet conductance compensates for the averaged conditions in the thicker upper layer in the three-layer model. Overall, ground-water levels and ground-water discharge were closer to measured values when conductance was reduced.

Changes in model response owing to imposed changes in ground-water pumpage are shown in figures 42 and 43. Pumpage had no regional effect on ground-water levels until after 1950. The simulated effect of a 50-percent increase in 1980 pumpage was an average head decline of about 5 ft. When pumpage was decreased 50 percent, heads rose about 5 ft; when pumpage was removed, heads rose about 9 ft. Absolute deviations were similar to those of the calibrated model run (table 23). Effects of ground-water pumpage on ground-water discharge to the Snake River are shown in figure 43. The similarity of hydrographs based on model-calibrated and measured water levels indicates that pumpage estimates are reasonable.

Model response to imposed changes in tributary drainage-basin underflow (boundary flux) was similar to model response to changes in aquifer recharge, although the magnitude of change was smaller (figs. 44, 45). A 50-percent increase in tributary drainage-basin underflow raised aquifer heads about 11 ft; a 50-percent reduction resulted in about an equal head decline. Head changes at well 8S-14E-16CBB1 were smaller than average, owing to its proximity to major springs with constant head. Absolute deviations were larger than deviations in the calibrated model run (table 23).

Across the study area, the model was most sensitive to changes in transmissivity and recharge. In major spring areas, near American Falls Reservoir and along the Snake River from Milner to King Hill, the model was most sensitive to changes in riverbed or spring-outlet conductances. The importance of riverbed or spring-outlet conductances as controls on aquifer head decreased with increasing distance from spring-discharge areas. The model was relatively insensitive to changes in boundary flux and ground-water pumpage. However, if these parameters are considered in conjunction with recharge flux, their determination is critical to proper simulation of the aquifer system. Of relatively minor importance to model response were changes in storage coefficient and leakance.

HYPOTHETICAL DEVELOPMENT ALTERNATIVES

The transient model was used to simulate aquifer response to three hypothetical development alternatives that might take place by the year 2010: (1) continuation of 1980 hydrologic conditions and pumping rates, (2) increased pumpage, and (3) increased recharge. These alternatives are highly simplified, and a large number of plausible situations involving various combinations of existing conditions and hypothetical changes in pumpage and recharge could develop. The purpose of testing development alternatives was to evaluate general hydrologic trends that might be expected should some or all of the alternatives be realized. Although testing of specific management alternatives was not an objective of this study, it is possible, with the calibrated model, to evaluate the effects of water-management proposals on the regional aquifer system.

Average annual mass-balance calculations for each of the three hypothetical alternatives are shown in table 24. Simulation of a continuation of 1980 hydrologic conditions and ground-water pumpage was based on recharge and discharge calculations used in the calibrated steady-state model discussed earlier. The simulation indicates possible changes in the ground-water budget over the period 1980–2010 if recharge and discharge remain the same as in 1980. The result is a gradual decrease in the release of water from storage from about 8 percent of total aquifer discharge in 1980 to about 1 percent in 2010. Accompanying declines in aquifer head from 1980 to 1985 and from 1980 to 2010 are shown on plate 10. The results are consistent with head declines measured during the 1980 water year. Simulation of a continuation of 1980 conditions indicated that after 5 years, water levels in the central part of the eastern Snake River Plain would decline about 2 ft and, after 30 years, would decline 2 to 8 ft. The model indicated that declines would be much greater in several areas along the margins of the plain. However, these areas are less accurately simulated than the central part of the plain, owing to large changes in hydraulic conductivity along the margins. Consequently, confidence in the magnitude of change along the margins of the plain is lower. Changes in head would cause changes in ground-water discharge and river leakage (fig. 46). Generally, these simulated changes were small; ground-water discharge (river gains) decreased about 5 percent and river leakage (losses) increased about 9 percent across the modeled area (table 24).

Aquifer response to increased ground-water pumpage also was simulated. All potentially arable lands (1,070,000 acres) shown in figure 47 were assumed to be irrigated with ground water. It was further assumed that 1.6 acre-ft of water per acre (average consumptive irrigation requirement on the plain) were applied annually. The result was an average annual increase of 2,400 ft^3/s in ground-water pumpage. The model indicated that heads would decline 5 to 15 ft across the central plain within 5 years (pl. 10) and would decline 10 to 50 ft within 30 years. Simulated head declines along the margins of the plain were greater but, because of model uncertainties, probably are less reliable. In addition to the large head changes, river leakage to the aquifer was increased by about 50 percent and ground-water discharge was decreased by about 20 percent (table 24). Although an increase in pumpage of this magnitude is unlikely, this simulation illustrates the potential for large changes in aquifer conditions if pumpage were increased substantially.

Aquifer response to increased recharge was simulated by adding an average annual 800 ft^3/s to 1980 base conditions. Norvitch and others (1969) simulated an average increase in recharge of 500 ft^3/s during a study of potential effects of artificial recharge on the regional aquifer. Results of simulations demonstrate the usefulness of ground-water-flow models in evaluating possible effects of artificial recharge. Recharge was increased in four areas, model blocks (row-column) 9–11, 16–37, 14–39, and 8–45, that were used as artificial recharge sites in the study by Norvitch and others (1969). Of the three hypothetical development alternatives simulated, increasing recharge resulted in the least amount of change. After 5 years of increased recharge, water levels in the central part of the plain increased from 0 to 5 ft (pl. 10) and showed little additional change during the next 25 years. These increases in head are similar to those reported by Norvitch and others (1969), which ranged from less than 1 to more than 5 ft. Large declines simulated along the margins of the plain likely are due to poorly estimated underflow rates. Increased recharge decreased simulated leakage from rivers to the regional aquifer by about 6 percent, and spring flow (river gains) remained essentially the same (table 24). This simulation indicates that increasing recharge would have little regional effect on hydrologic conditions, although in the immediate area of application, ground-water levels would rise.

SUMMARY AND CONCLUSIONS

Flow in the regional aquifer system in the eastern Snake River Plain is controlled largely by the Snake River and its major tributaries. Most ground-water recharge is from infiltration of surface water diverted for irrigation and leakage from the Snake River and its major tributaries. A poorly defined but likely

TABLE 24.—*Average annual mass-balance calculations, 1980–2010*

[Values in cubic feet per second]

Simulation period	Inflow: Change in ground-water storage	Inflow: Underflow	Inflow: Recharge from irrigation and precipitation	Inflow: River losses	Outflow: Change in ground-water storage	Outflow: Pumpage	Outflow: River gains
			Continuation of 1980 hydrologic conditions and pumpage				
[1]1976–80	1,010	3,110	7,320	1,310	60	1,980	10,710
1981–85	690	2,740	7,670	1,360	80	1,790	10,580
1986–90	480	2,740	7,670	1,380	10	1,790	10,460
1991–95	360	2,740	7,670	1,390	0	1,790	10,370
1996–2000	280	2,740	7,670	1,410	0	1,790	10,300
2001–2005	220	2,740	7,670	1,420	0	1,790	10,250
2006–2010	170	2,740	7,670	1,430	0	1,790	10,210
			Pumpage increased by 2,400 ft^3/s				
[1]1976–80	1,010	3,110	7,320	1,310	60	1,980	10,710
1981–85	1,990	2,740	7,670	1,580	20	4,150	9,820
1986–90	1,320	2,740	7,670	1,750	0	4,150	9,330
1991–95	930	2,740	7,670	1,840	0	4,150	9,030
1996–2000	670	2,740	7,670	1,900	0	4,150	8,830
2001–2005	500	2,740	7,670	1,930	0	4,150	8,700
2006–2010	380	2,740	7,670	1,960	0	4,150	8,610
			Recharge increased by 800 ft^3/s				
[1]1976–80	1,010	3,110	7,320	1,310	60	1,980	10,710
1981–85	540	3,540	7,670	1,250	320	1,790	10,880
1986–90	340	3,540	7,670	1,230	80	1,790	10,890
1991–95	260	3,540	7,670	1,230	20	1,790	10,870
1996–2000	200	3,540	7,670	1,230	10	1,790	10,840
2001–2005	160	3,540	7,670	1,230	0	1,790	10,810
2006–2010	130	3,540	7,670	1,230	0	1,790	10,780

[1]Calibrated model values included for comparison purposes.

small amount of recharge is from precipitation on the plain; most precipitation on the plain is either evaporated or transpired. Aquifer discharge is largely spring flow to the Snake River and water pumped for irrigation. Largest well yields are obtained from Quaternary basalt; some sand and gravel aquifers also yield relatively large quantities of water. Older basalt and rhyolite generally yield less water but are important aquifers in places. In some areas, clay and silt lenses are confining layers that impede vertical flow.

Regional ground-water flow is generally southwestward, from major recharge areas in the northeast to the major discharge area along the Snake River from Milner to King Hill. Hydraulic-head changes with depth are defined in major recharge and discharge areas and where silt, clay, and unfractured crystalline basalt layers impede vertical flow. Ground-water levels rose and ground-water discharge (largely spring flow) increased soon after surface-water irrigation began on large tracts of land after 1890. Water levels and ground-water discharge peaked in about 1950 and have declined since, owing to a combination of factors, including increased ground-water pumpage. Ground-water levels fluctuate seasonally in response to recharge from precipitation and surface-water irrigation and pumping stress; they also fluctuate in response to long-term climatic trends.

Two-dimensional steady-state, three-dimensional steady-state, and three-dimensional transient simulations were used to analyze the regional aquifer system. The two-dimensional analysis incorporated a nonlinear, least-squares regression technique to estimate aquifer variables (or parameters). Major assumptions in the parameter-estimation analysis were that ground-water flow is two dimensional, and that the ground-water system in 1980 was at steady state.

Across much of the eastern plain, flow in the regional aquifer system is virtually two dimensional.

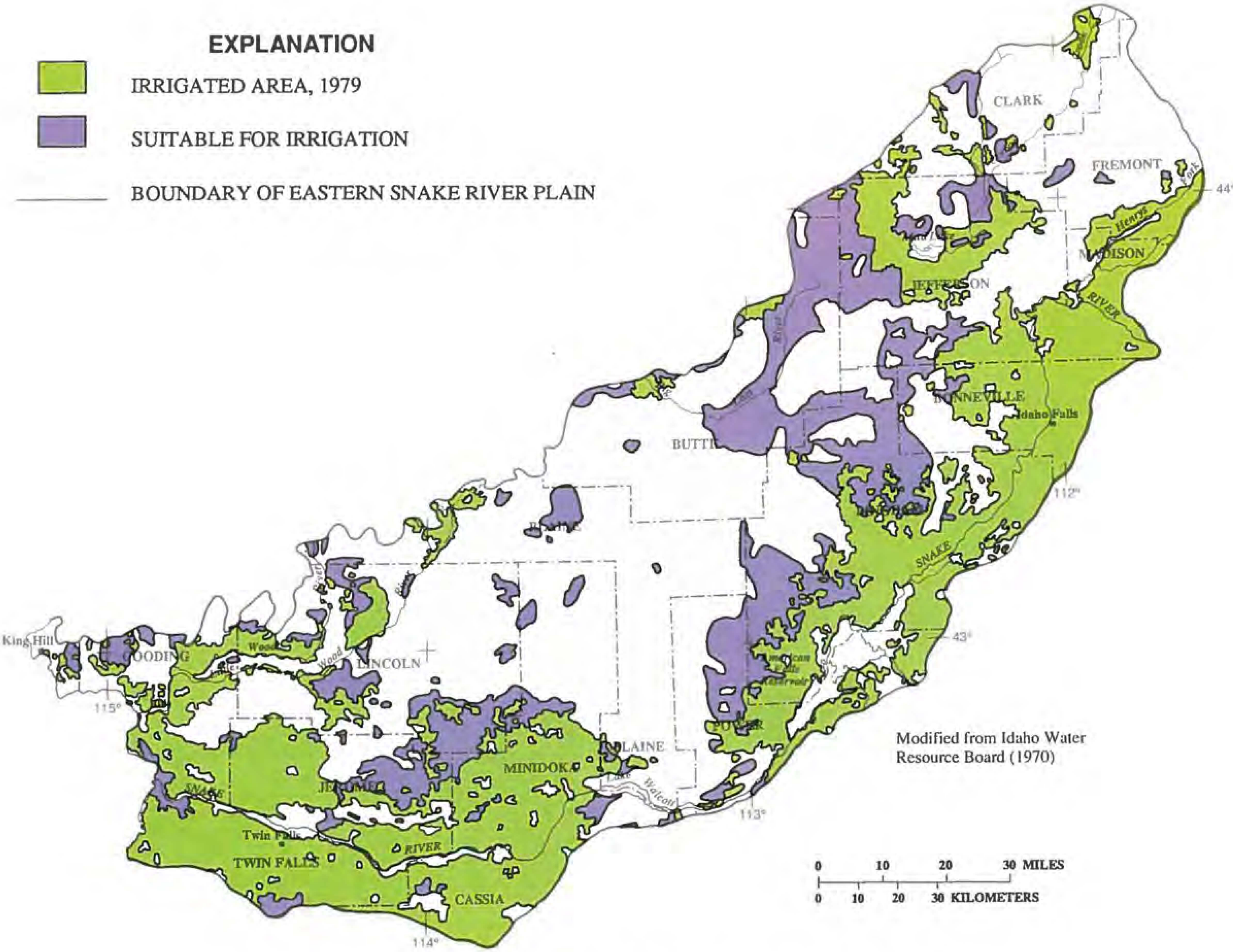


FIGURE 47.—Irrigated areas (1979) and areas suitable for irrigation.

However, large vertical head differences were measured in major recharge and discharge areas and along the margins of the plain. Simulations indicated that an average of 700,000 acre-ft of water per year were removed from ground-water storage from 1976 to 1980, whereas 100,000 acre-ft of water were removed from storage in 1980. The large average amount is undoubtedly influenced by the severe drought in 1977, when more water probably was removed from storage.

Sensitivity analysis indicated that simulated recharge, underflow, leakance, riverbed or spring-outlet conductance, ground-water pumpage, ground-water discharge, and transmissivity are reasonable and are the most important determinants of model response. Storage coefficients are less important because high transmissivities allow rapid head changes throughout the regional system.

Historical records and results from transient simulations indicate how changes in ground-water levels dramatically change ground-water discharge. Mundorff and others (1964, p. 162) estimated that an average water-level rise of 60 to 70 ft from about 1910 to 1959 increased ground-water discharge along the north side of the Snake River from Milner to King Hill to about 2,500 ft^3/s (1,800,000 acre-ft/yr). Historical records and results from transient simulations also indicate a decreased amount of leakage from the Snake River to the ground-water system from Heise to near Blackfoot. These data show the importance of understanding stream-aquifer relations and how they change with time in response to development stresses. Sensitivity analysis indicated that aquifer heads are responsive to changes in riverbed or spring-outlet conductance, particularly near the Snake River.

The regional aquifer system in the eastern Snake River Plain responds quickly and over broad areas to changes in inflow and outflow, which include recharge from irrigation, stream and canal leakage, tributary drainage-basin underflow, and ground-water pumpage. Transient simulations were made to evaluate long-term regional changes in aquifer heads and ground-water discharge. For example, simulated results indicate that the ground-water system responds rapidly to changes in pumpage. Historical records of rapid water-level rises and increased ground-water discharge are approximated by the three-dimensional transient model results, which indicates that the model can reasonably simulate the regional ground-water system.

The transient model was used to simulate aquifer changes from 1980 to 2010 in response to three hypothetical development alternatives: (1) continuation of 1980 hydrologic conditions, (2) increased pumpage, and (3) increased recharge. Simulation of continued 1980 hydrologic conditions for 30 years indicated that head declines of 2 to 8 ft might be expected in the central part of the plain. The magnitude of simulated head declines was consistent with head declines measured during the 1980 water year. Larger declines were simulated along the model boundaries, but these declines may have resulted from underestimation of tributary drainage-basin underflow and inadequate aquifer definition. Simulation of increased ground-water pumpage (by 2,400 ft^3/s) for 30 years indicated head declines of 10 to 50 ft in the central part of the plain. These relatively large head declines were accompanied by increased simulated river leakage of about 50 percent and decreased spring discharge of about 20 percent. The effect of 30 years of increased recharge (800 ft^3/s) was a rise in simulated heads of 0 to 5 ft in the central part of the plain.

More and better data and continued model development and testing are needed to further improve understanding of the hydrologic system in the eastern Snake River Plain. Better definition of aquifer hydraulic conductivity is needed, particularly along the margins of the plain. Mass-flux estimates can be improved by obtaining better estimates of surface-water diversions, irrigation-return flow, streamflow, and ground-water pumpage. To better define stream-aquifer relations, data are needed on streambed hydraulic conductivities.

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FIGURES 32–46

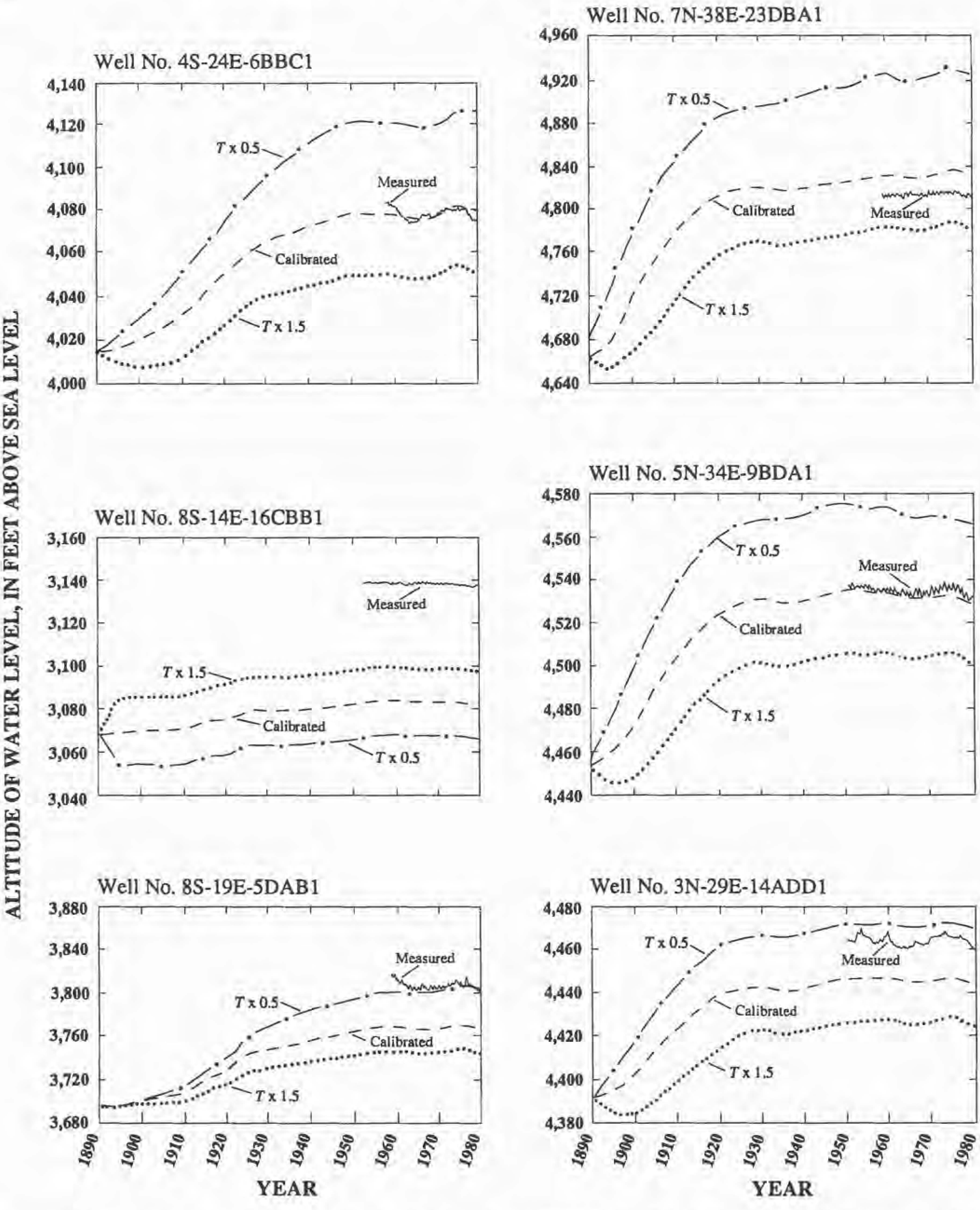

FIGURE 32.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in T (transmissivity).

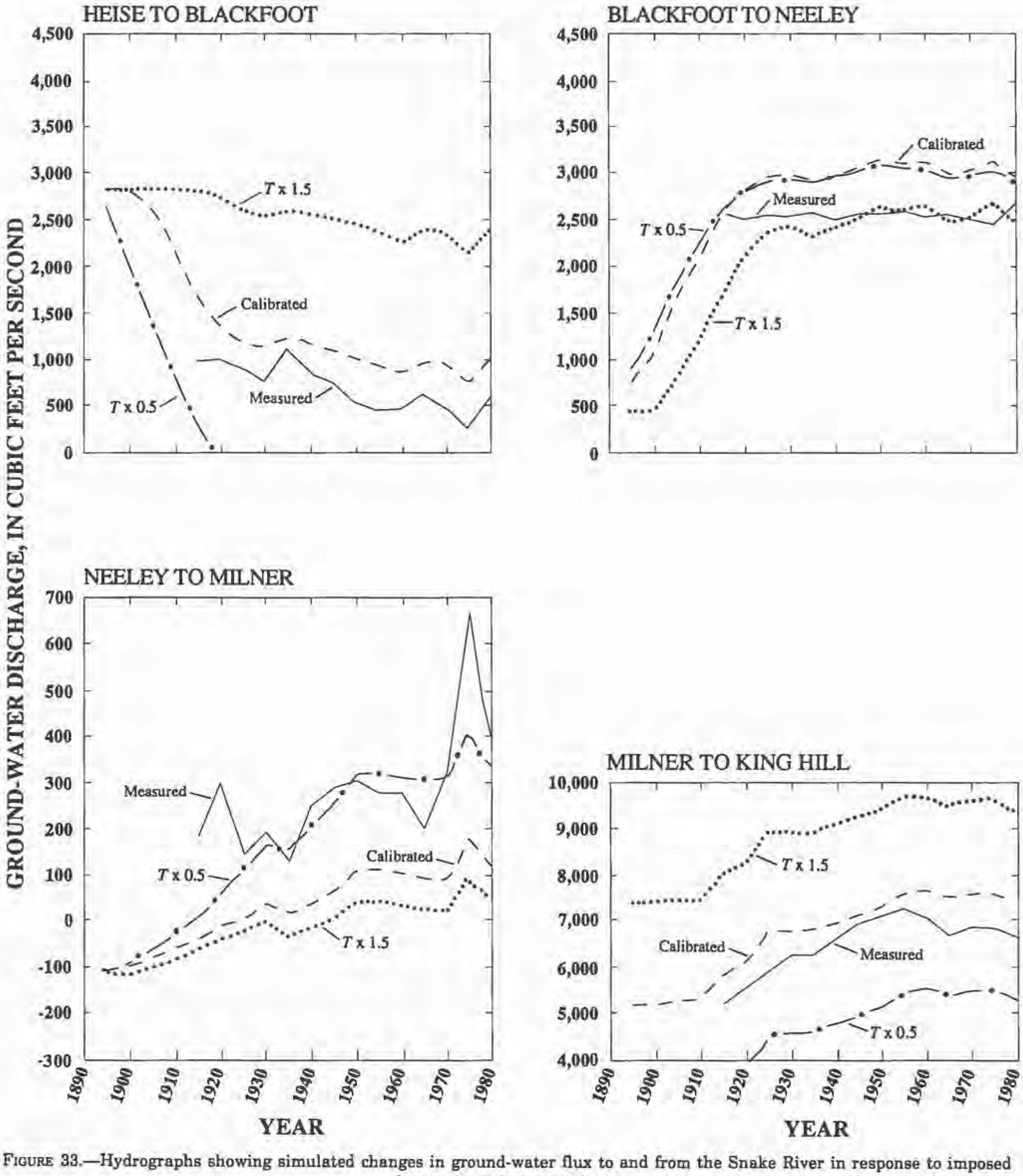

FIGURE 33.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in T (transmissivity).

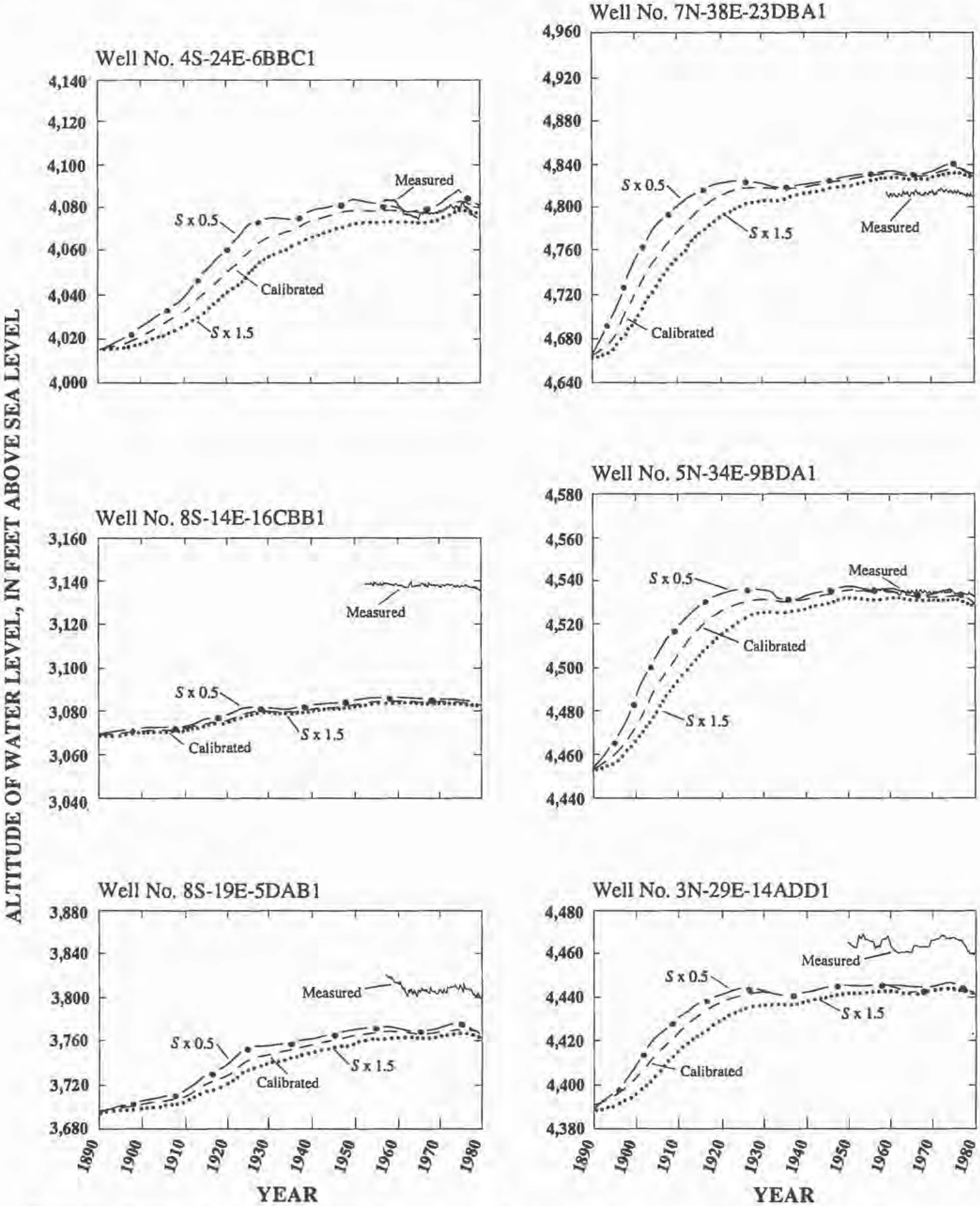

FIGURE 34.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in S (storage coefficient).

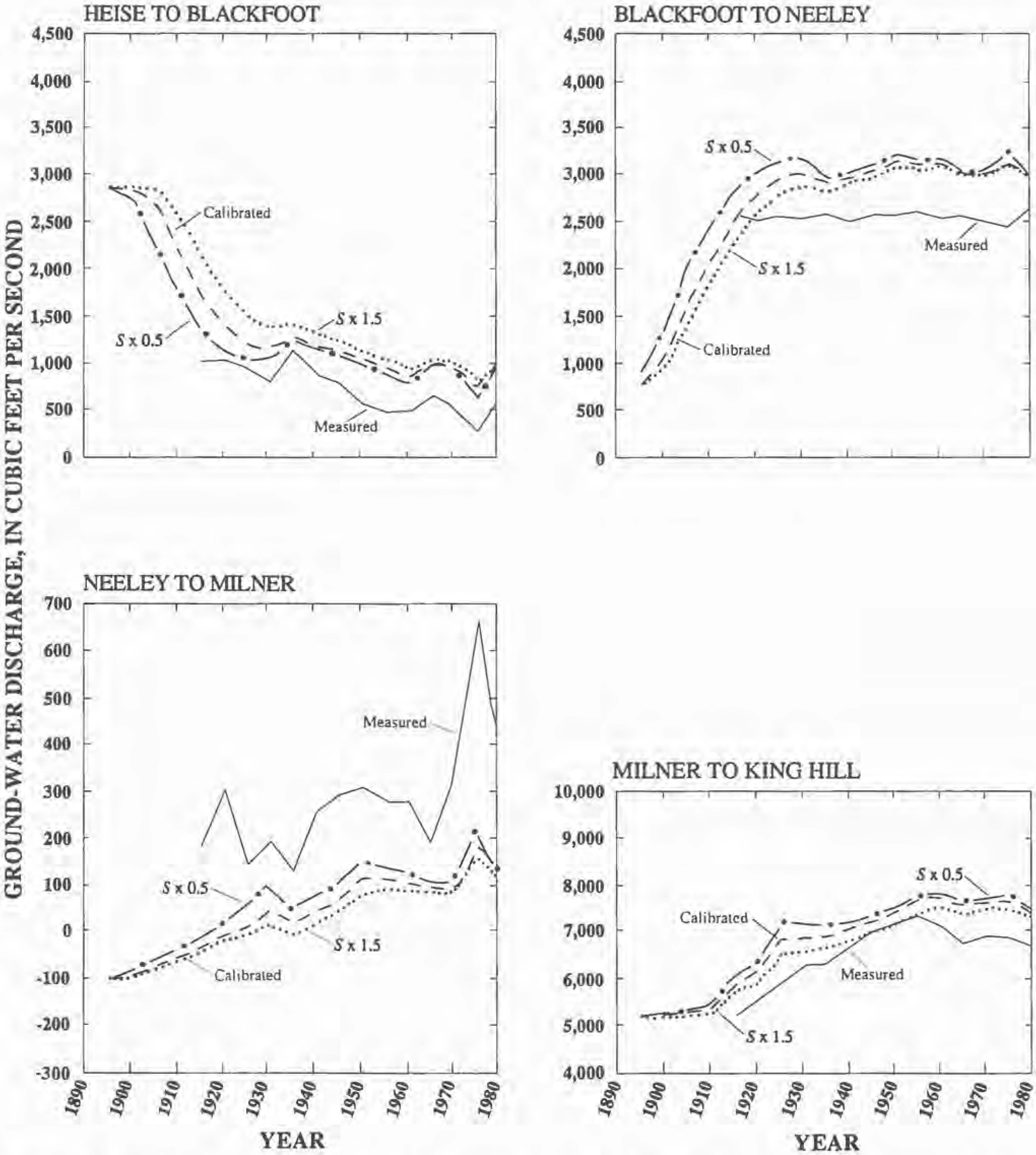

FIGURE 35.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in S (storage coefficient).

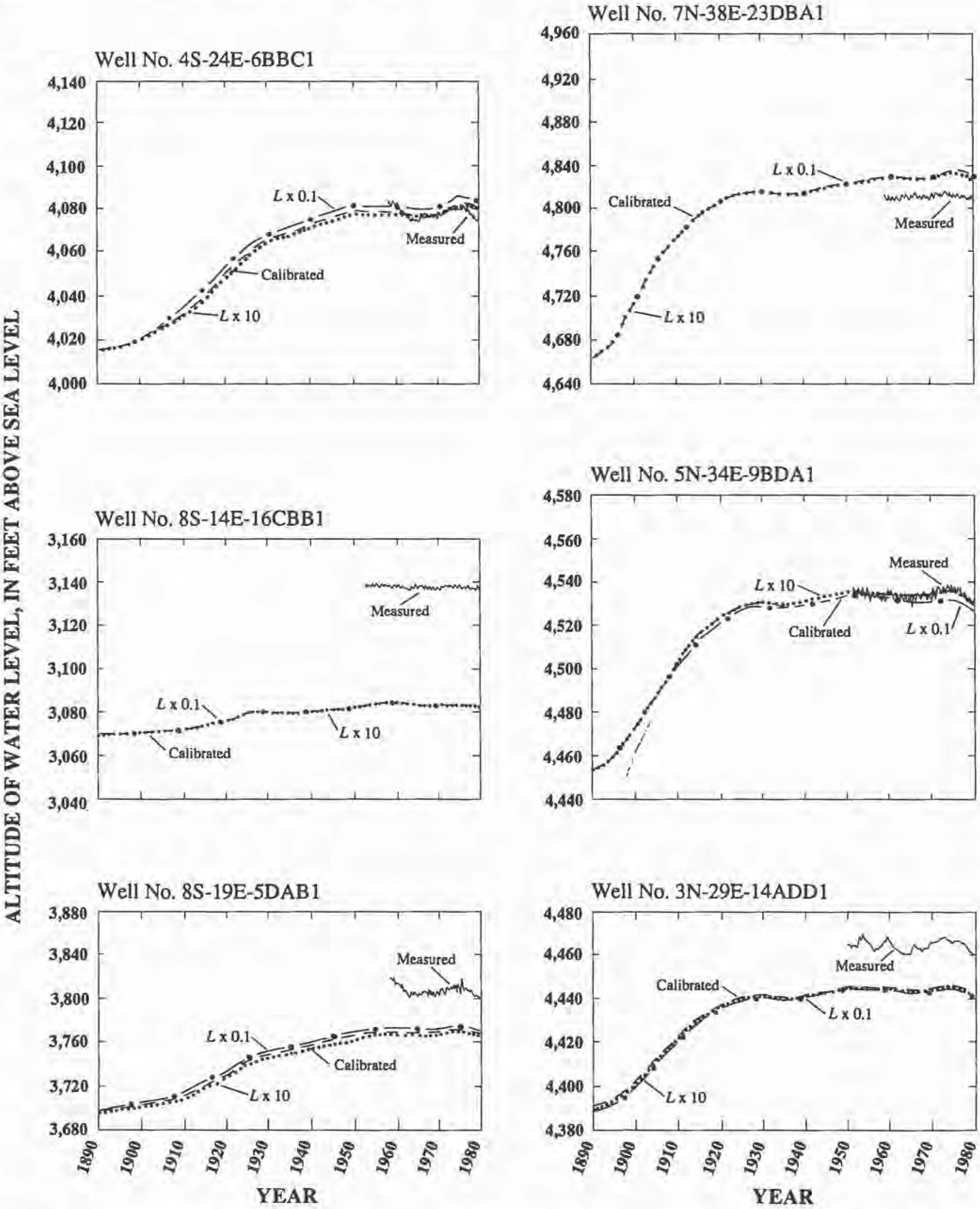

FIGURE 36.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in L (leakance).

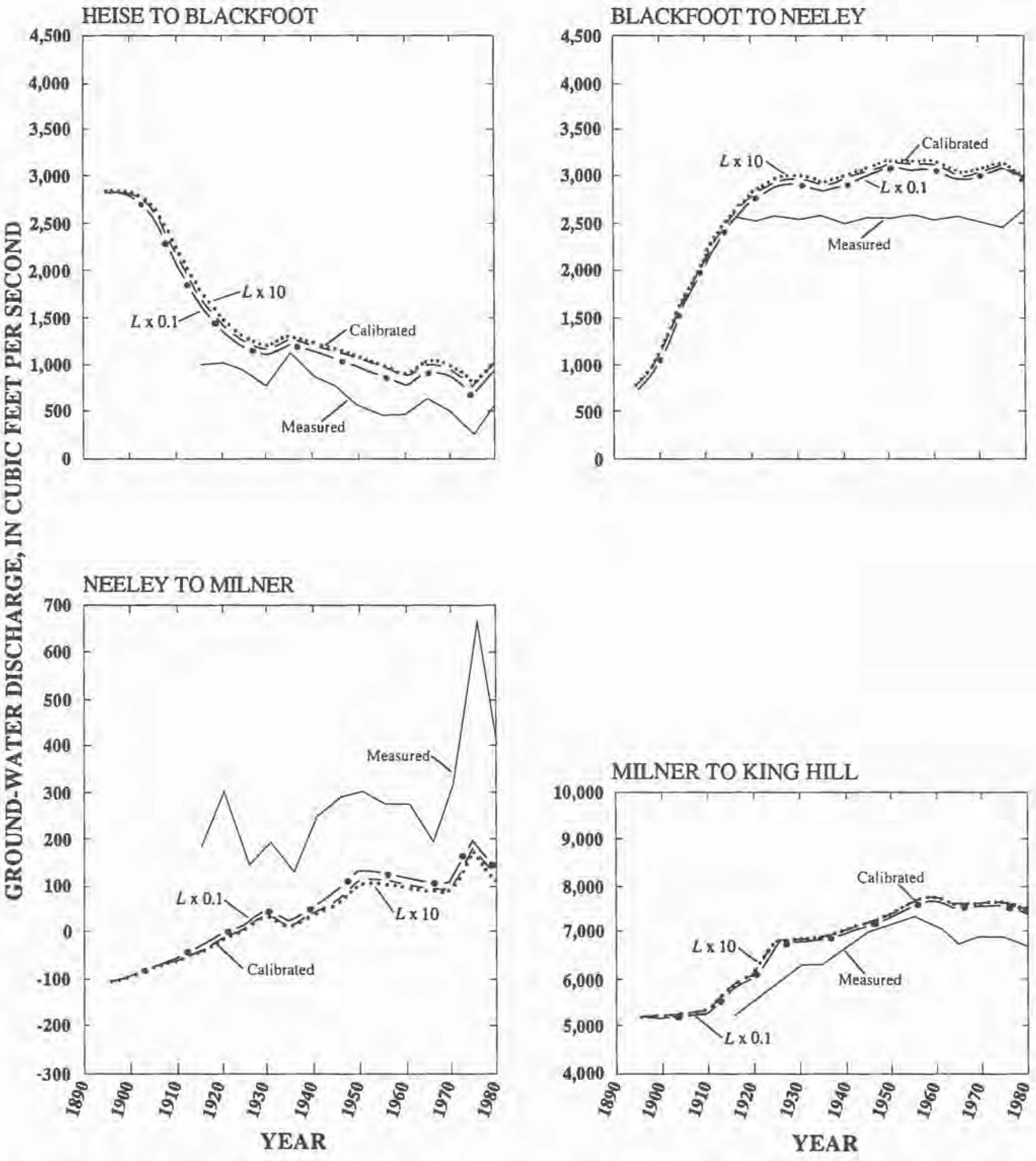

FIGURE 37.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in *L* (leakance).

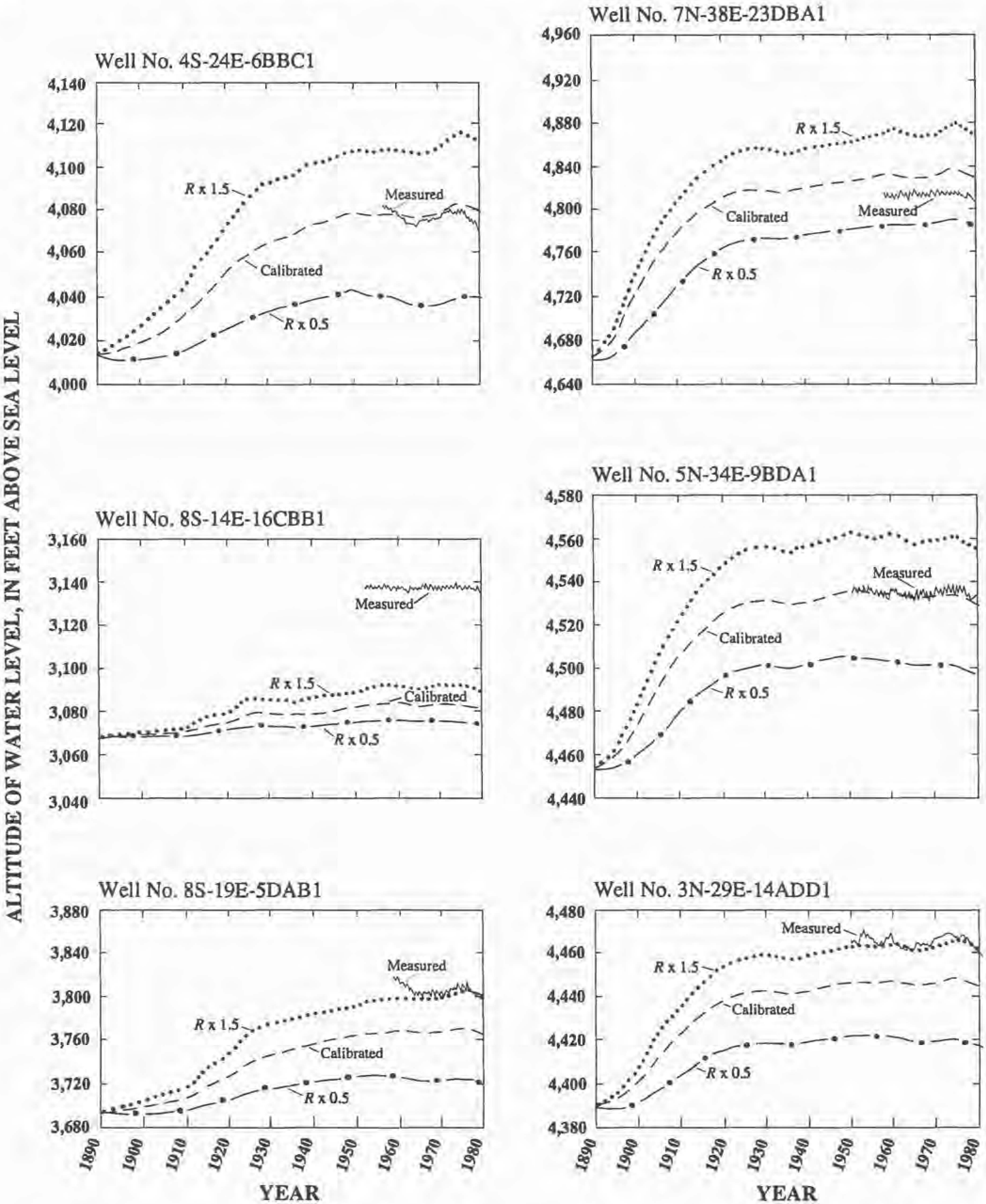

FIGURE 38.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in R (recharge).

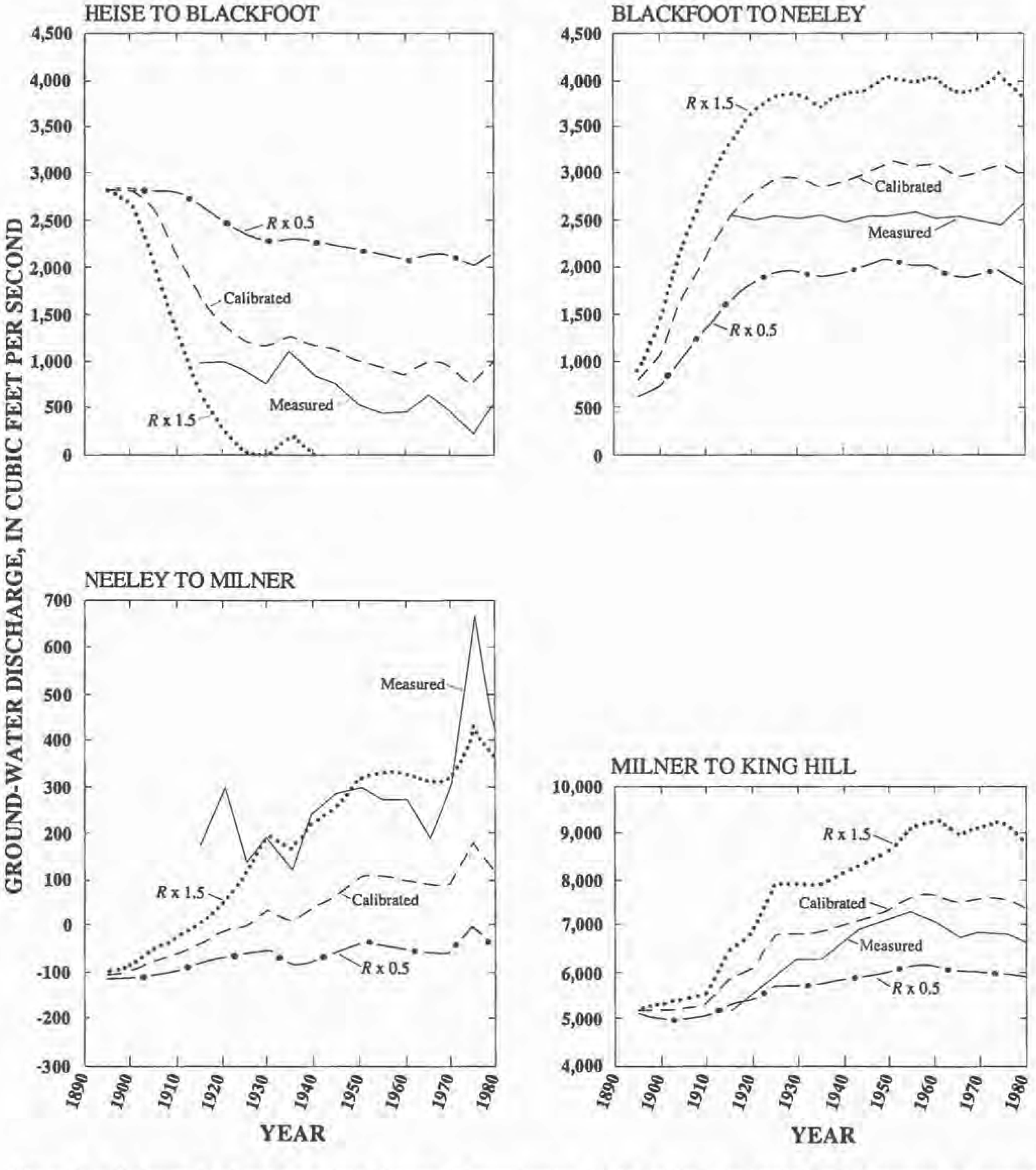

FIGURE 39.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in R (recharge).

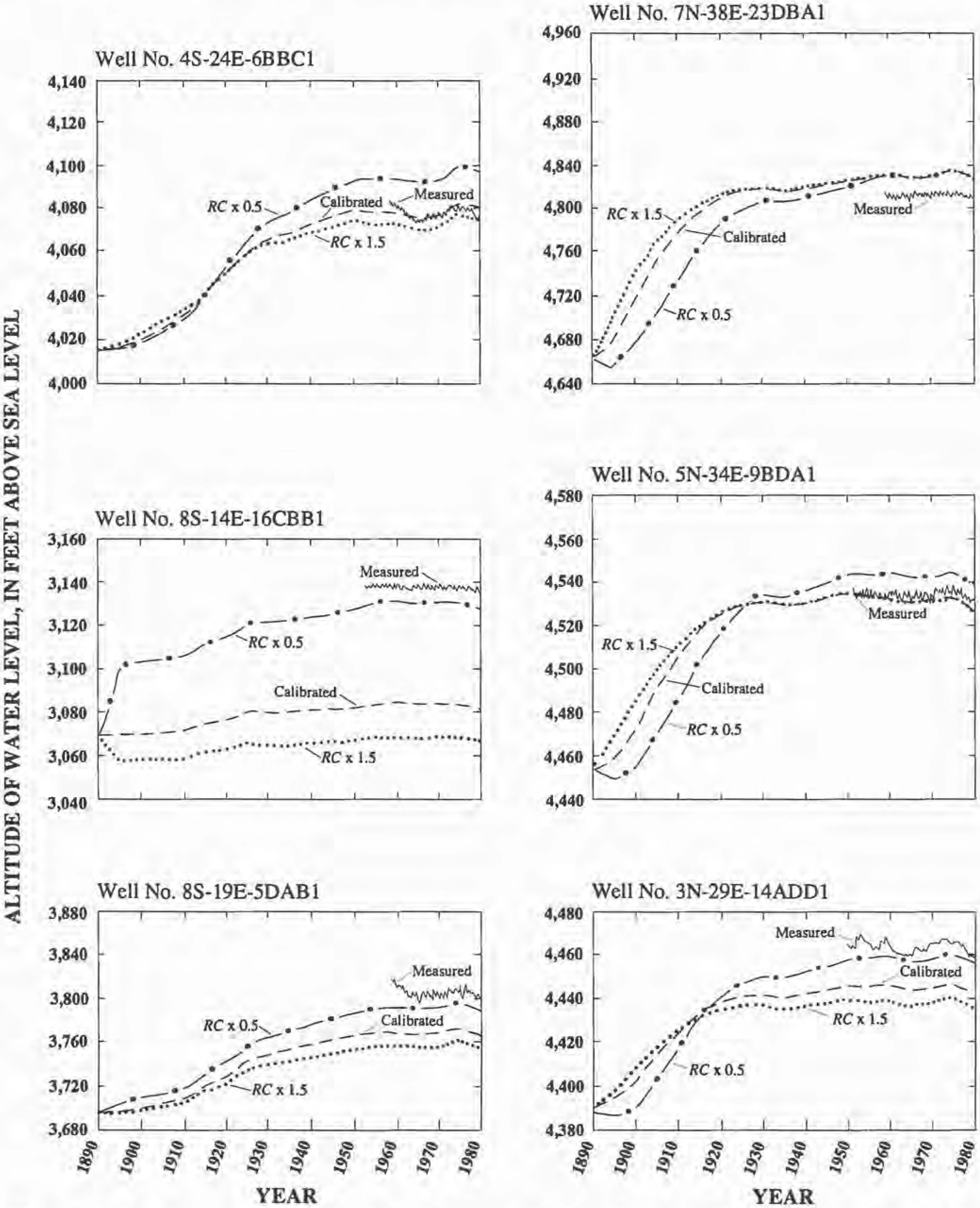

FIGURE 40.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in *RC* (riverbed or spring-outlet conductances).

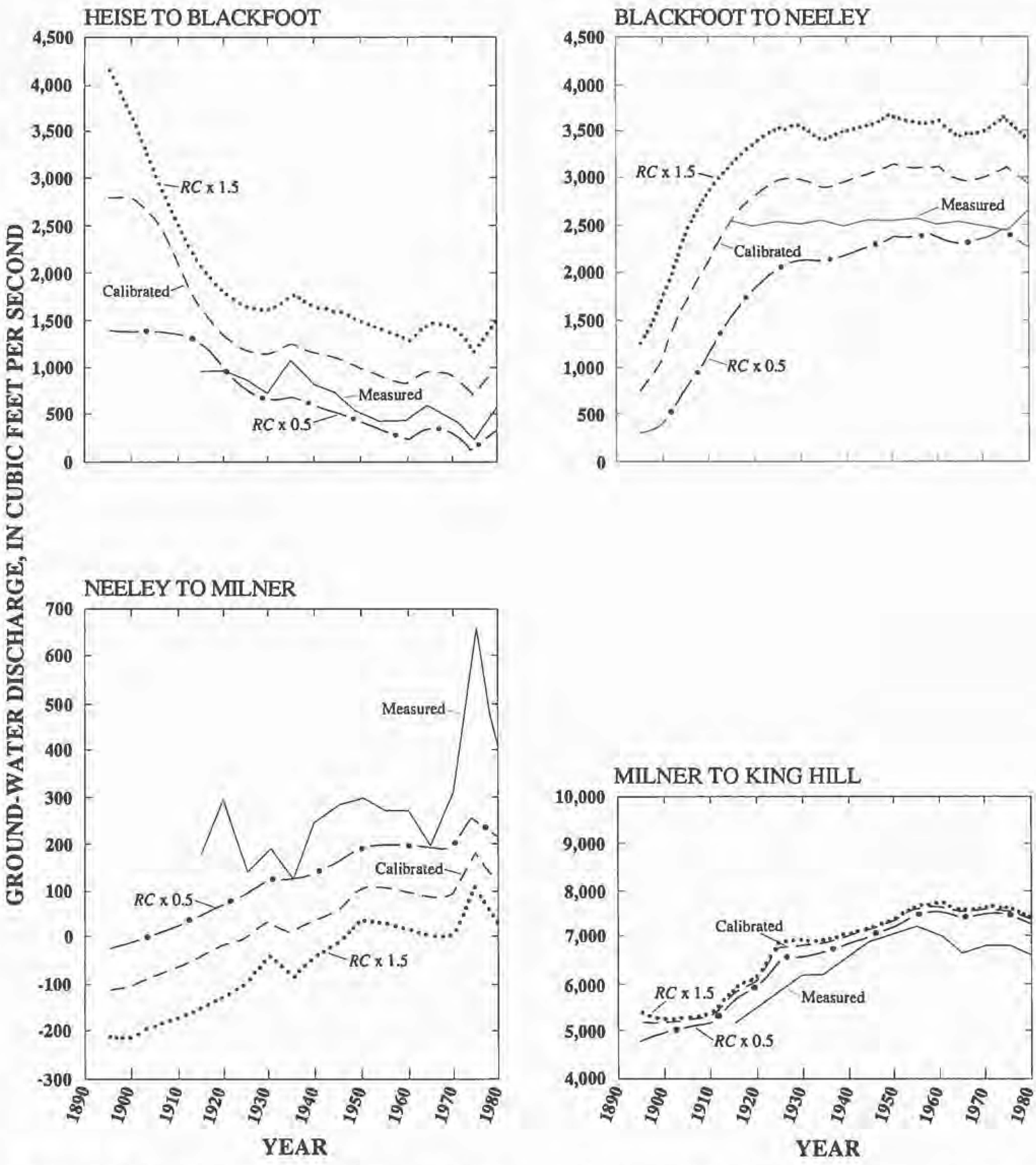

FIGURE 41.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in *RC* (riverbed or spring-outlet conductances).

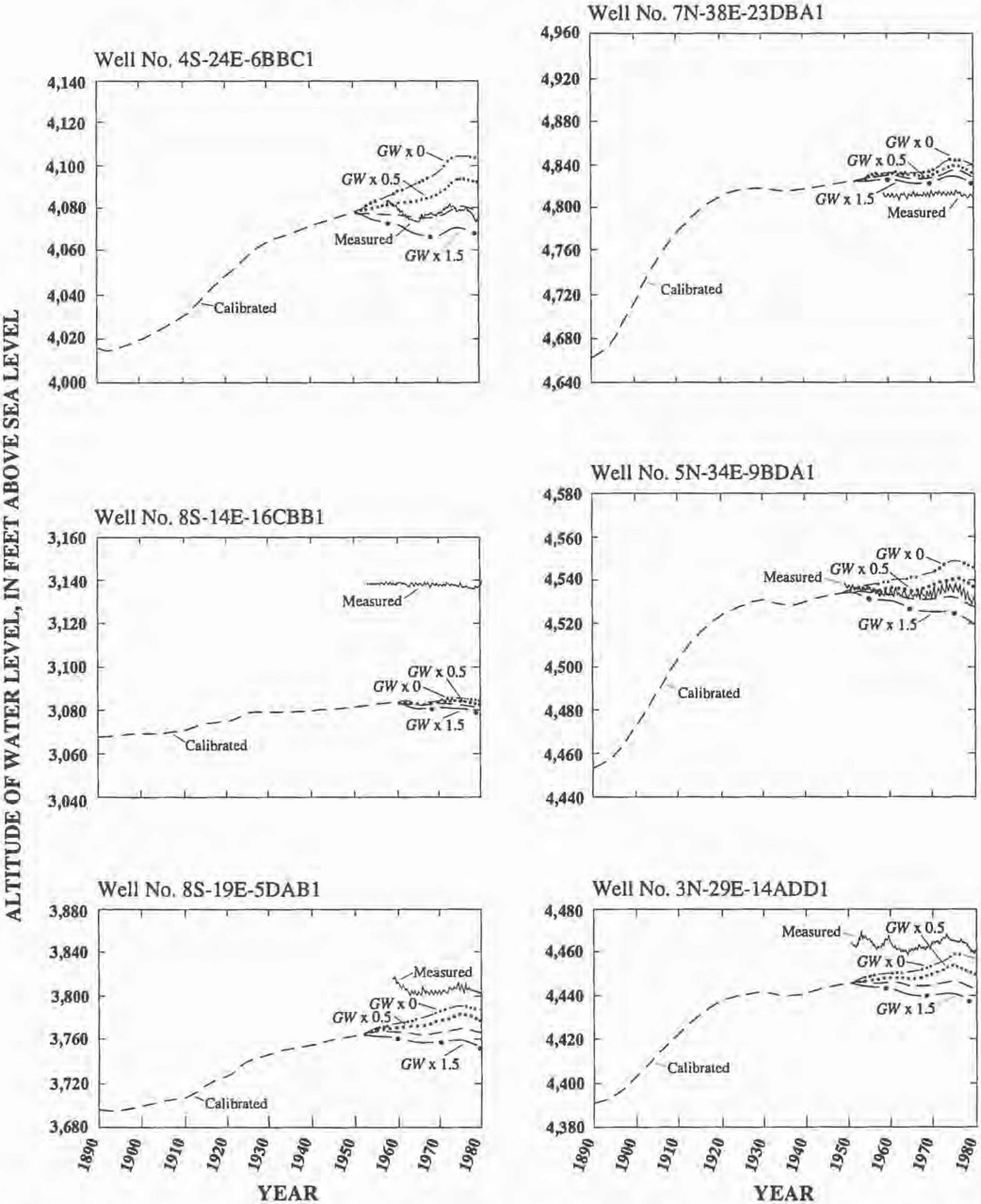

FIGURE 42.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in *GW* (ground-water pumpage).

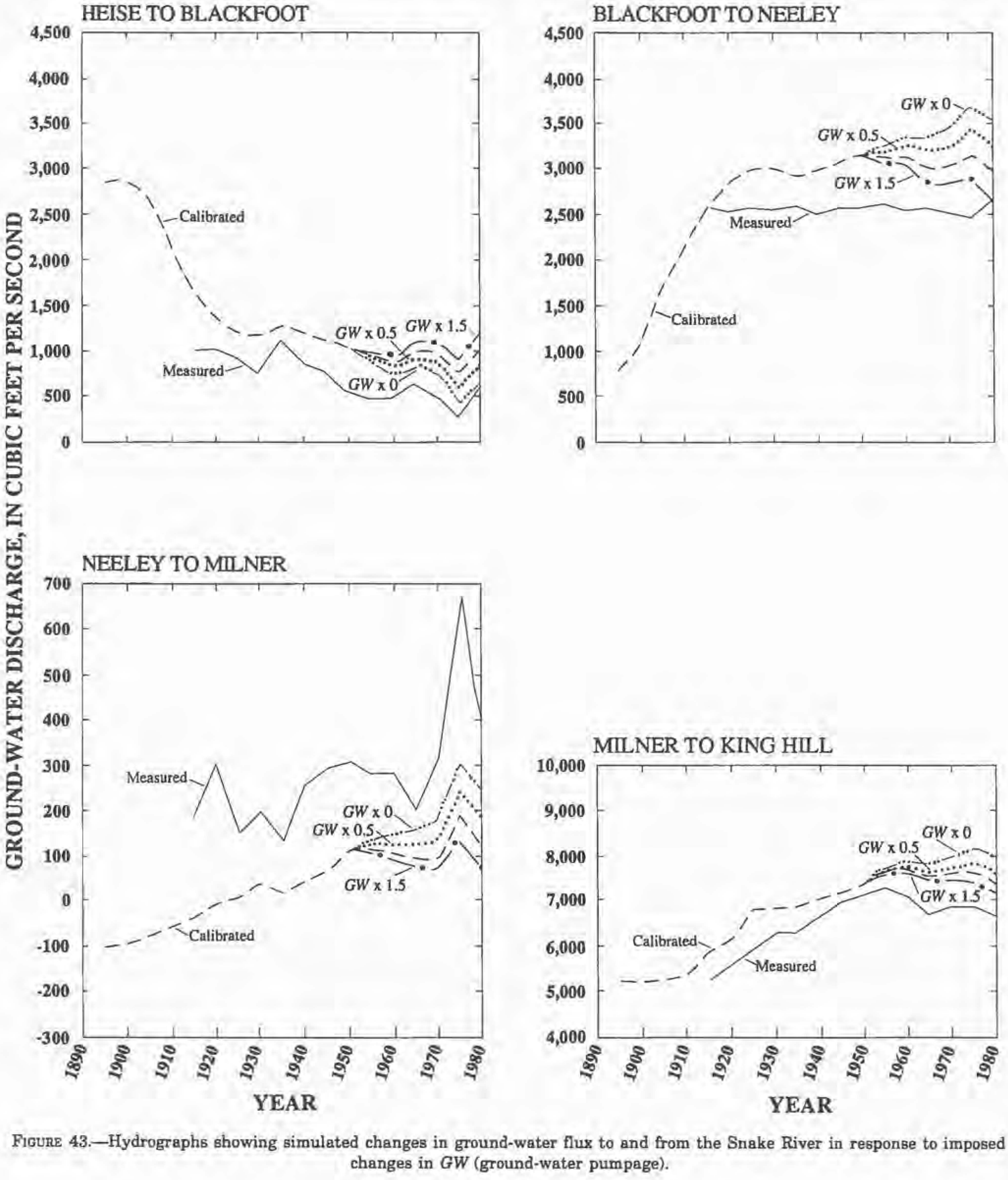

FIGURE 43.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in *GW* (ground-water pumpage).

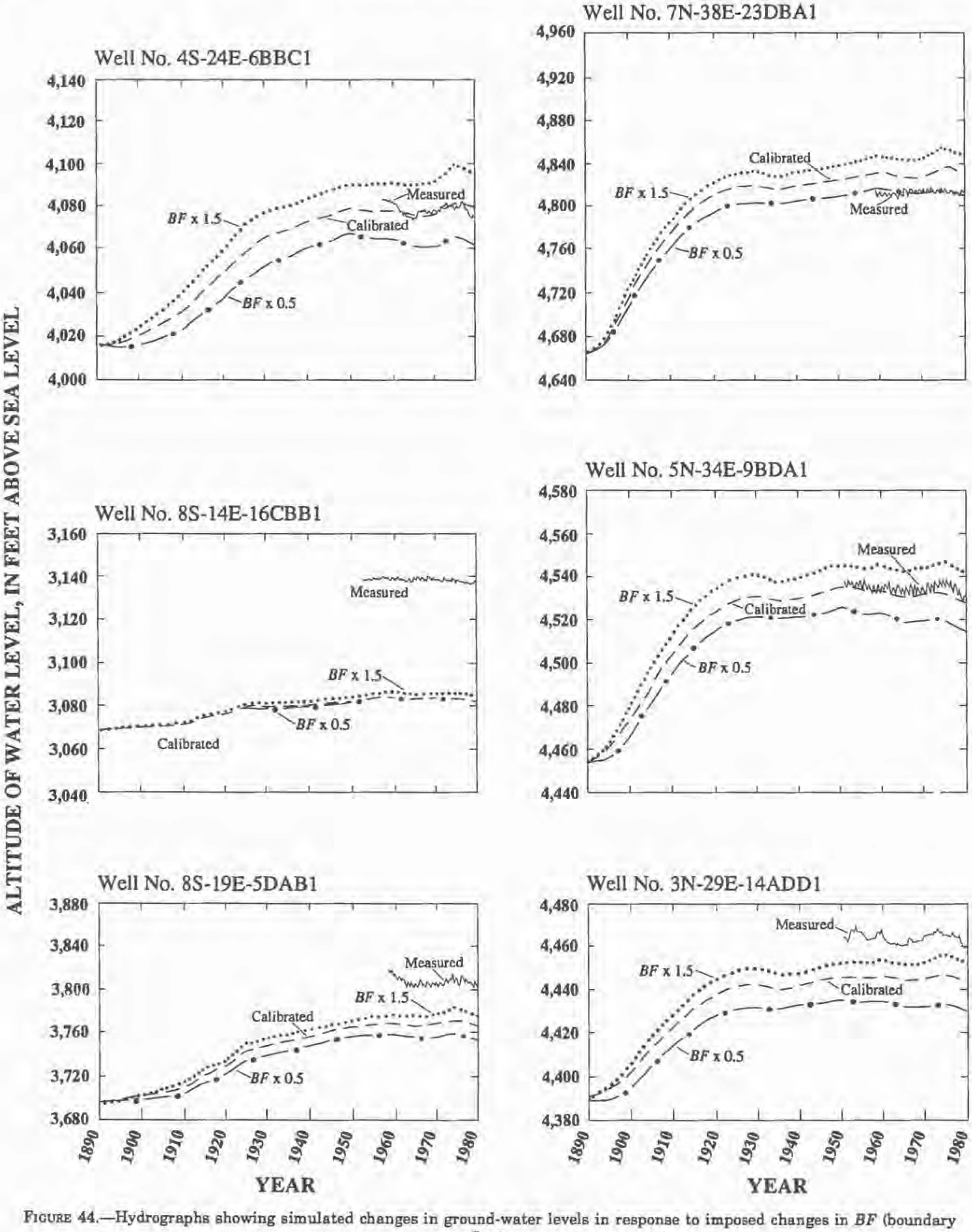

FIGURE 44.—Hydrographs showing simulated changes in ground-water levels in response to imposed changes in *BF* (boundary flux).

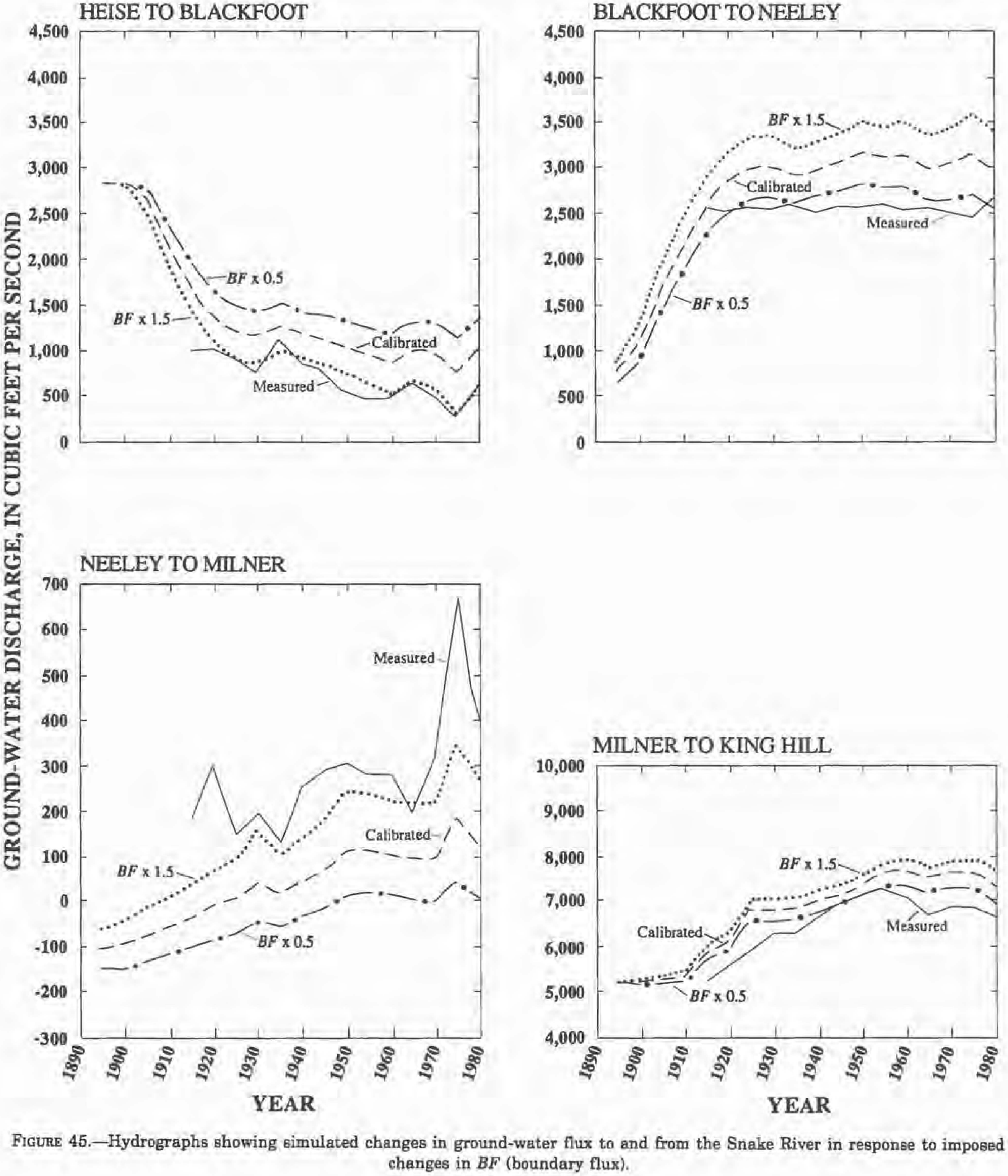

FIGURE 45.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to imposed changes in *BF* (boundary flux).

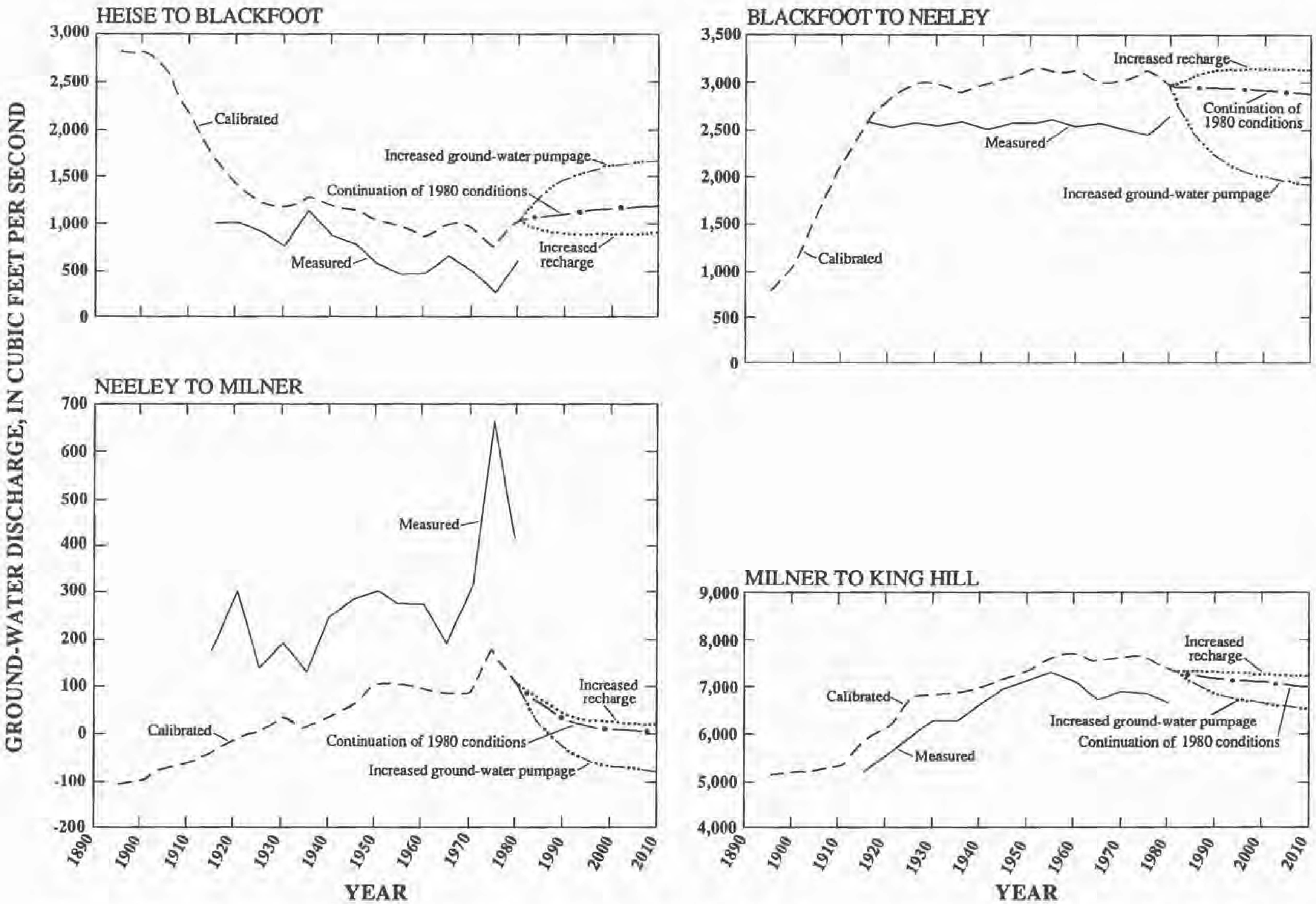

FIGURE 46.—Hydrographs showing simulated changes in ground-water flux to and from the Snake River in response to continuation of 1980 hydrologic conditions, increased recharge, and increased ground-water pumpage.

APPENDIXES A–C

APPENDIX A.—*Diversion and return-flow data for water year 1980*

This appendix lists 1980 water year diversion and return-flow data and data sources for surface-water-irrigated areas on the eastern Snake River Plain. Areas shown in figure 7 include surface-water-irrigated lands where diversion records are available. Sources of data are the following:

a. Idaho Department of Water Resources (1980)
b. U.S. Geological Survey (1980)
c. U.S. Bureau of Reclamation (written commun., 1981)
d. Water Districts 37, 37M (1980)
e. American Falls District No. 2 (written commun., 1981)
f. Wytzes (1980)
g. Kjelstrom (1986)
h. Idaho Department of Water Resources (written commun., 1981)

The data-source identifier (a–h) is used as a prefix in the following tables for the irrigation areas.

Irrigation Area 1.—*Diversions from Falls River*

Name		*Quantity (acre-feet)*
Marysville Canal	a	32,900
Farmers Own Canal	a	14,900
Yellowstone Canal	a	2,900
Orme Canal	a	800
Squirrel Creek	a	1,700
Boom Creek	a	800
Conant Creek	a	6,000
Total		60,000
Estimated surface-return flows		21,200
Diversions minus surface return		38,800

Irrigation Area 2.—*Diversions from Henrys Fork, Falls River, and Teton River*

Name		*Quantity (acre-feet)*
Silkey	a	5,000
McBee	a	500
Stewart	a	3,000
Pioneer	a	1,600
Wilford	a	52,200
Farmers Friend	a	33,500
Twin Groves	a	41,100
Pincock–Byington	a	4,200
Cross Cut	a	39,700
Fall River	a	55,300
Chester	a	19,000
Pumps	a	5,400
Total		260,500
Estimated surface-return flows		106,200
Diversions minus surface return		154,300

Irrigation Area 3.—*Diversions from Henrys Fork*

Name		*Quantity (acre-feet)*
St. Anthony Union	a	165,100
Last Chance	a	30,800
Dewey	a	5,100
Independent	a	90,700
St. Anthony Union Feeder	a	38,300
Egin	a	112,100
Total		442,100
Estimated surface-return flows		63,000
Diversions minus surface return		379,100

Irrigation Area 4.—*Diversions from Falls River, Henrys Fork, and Teton River*

Name		*Quantity (acre-feet)*
Salem Union	a	60,600
Roxana	a	4,400
North Salem	a	1,900
Consolidated Farmers	a	84,300
Curr	a	14,500
Enterprise	a	20,300
Teton Irrigation	a	24,500
Saurey–Somers	a	4,600
Island Ward	a	7,500
Teton Island Feeder	a	92,300
Pincock–Gardner	a	1,300
Rexburg City	a	5,000
Rexburg Irrigation	a	52,400
Woodmansee–Johnson	a	5,400
Siddoway	a	1,200
McCormick–Rowe	a	400
Bigler Slough	a	800
Pumps	a	400
Total		381,800
Estimated surface-return flows		135,200
Diversions minus surface return		246,600

Surface-return flows for irrigation areas 1–4 were estimated using data reported by Wytzes (1980) for the 1977 water year. Surface-return flows were adjusted for the 1980 water year by assuming that the total streamflow depletion for irrigation areas 1–4 was equal to the sum of the depletions within the areas, as expressed in the following equation:

basin inflow–basin outflow=(diversions–surface returns).

Therefore, if basin inflow, outflow, and diversions are known, the sum of all returns can be calculated. Knowing the total of all returns, one can adjust the returns reported by Wytzes (1980) by a common multiplier to equal the estimated total. Basin inflows (in acre-feet) for water year 1980 were

Henrys Fork at Ashton	g	1,102,400

Irrigation Area 4.—*Diversions from Falls River, Henrys Fork, and Teton River*—Continued

Falls River at Squirrel	g	550,400
Marysville Canal	g	32,900
Yellowstone Canal	g	2,900
Conant Creek	g	61,900
Teton River near St. Anthony	g	559,300
Moody Creek	g	10,800
Total		2,320,600

Basin outflows (in acre-feet) for water year 1980 were

Henrys Fork near Rexburg	g	1,491,900
Rexburg Canal drain	g	10,100
Total		1,502,000

Total diversions for areas 1–4 were 1,144,400 acre-ft. Total returns (in acre-feet) for areas 1–4 were

Surface inflow		*Outflow*		*Diversions*		*Returns*
2,320,600	–	1,502,000	–	1,144,400	=	–325,800

Surface-return flows (in acre-feet) estimated from data reported by Wytzes (1980) were

Area 1	6,000
Area 2	31,600
Area 3	17,800
Area 4	36,600
Total	92,000

The common multiplier is calculated as 325,800/92,000=3.54, and the estimated surface-return flows (in acre-feet) are

Area 1	21,200
Area 2	111,900
Area 3	63,000
Area 4	129,600

Irrigation Area 5.—*Right-bank diversions from the Snake River from Heise to Lorenzo*

Name		*Quantity (acre-feet)*
Hill–Pettinger	a	900
Nelson–Corey	a	1,700
Sunnydell	a	47,400
Lenroot	a	41,000
Reid	a	58,500
Texas, Liberty Park	a	79,100
Bannock Jim	a	5,200
Total		233,800

Irrigation Area 5.—*Right-bank diversions from the Snake River from Heise to Lorenzo*—Continued

Surface-return flows:

Texas Canal drain	g	19,100
Texas Slough	g	77,200
Bannock Jim Slough	g	8,800
Total		105,100
Diversions minus surface return		128,700

Irrigation Area 6.—*Left-bank diversions from the Snake River from Heise to Lorenzo*

Name		*Quantity (acre-feet)*
Riley	g	5,100
Anderson	g	93,400
Eagle Rock	g	135,400
Farmers Friend	g	112,900
Enterprise	g	56,500
Dry Bed	g	1,151,200
Nelson	g	700
Mattson–Craig	g	4,300
Pumps	g	700
Willow Creek near Ririe	g	73,500
Total		1,633,700

Surface-return flows:

Dry Bed	g	174,500
Spring Creek	g	21,700
Emigrant Creek	g	1,400
Drain	g	700
Anderson waste	g	6,300
Sand Creek	g	6,700
Little Sand Creek	g	3,500
Taylor	g	10,600
Henrys Creek	g	11,100
Willow Creek floodway	g	8,600
Total		245,100
Diversions minus surface return		1,388,600

Irrigation Area 7.—*Right-bank diversions from the Snake River downstream from Lorenzo to Shelley*

Name		*Quantity (acre-feet)*
Butte, Market Lake	a	71,600
Bear Trap	a	6,000
Osgood	a	9,300
Clements	a	700
Kennedy	a	3,500
Great Western	a	126,300
Porter	a	80,800
Woodville	a	21,500
McKay South	a	600
Total		320,300

Irrigation Area 7.—*Right-bank diversions from the Snake River downstream from Lorenzo to Shelley*—Continued

Surface-return flows:

Great Western waste	c	400
Great Western waste	c	30,700
Great Western waste	c	25,600
Butte, Market Lake return	c	7,200
Total		63,900
Diversions minus surface return		256,400

Irrigation Area 8.—*Left-bank diversions from the Snake River from Lewisville to Blackfoot*

Name		*Quantity (acre-feet)*
Idaho	a	295,200
Snake River valley	a	198,000
Blackfoot	a	111,500
Corbett	a	47,500
Nielson–Hansen	a	2,600
Sand Creek at Idaho Falls	c	6,700
Little Sand Creek at Ammon	c	3,500
Taylor	c	10,600
Henrys Creek	c	11,100
East Idaho Slough	c	13,800
Total		700,500
Surface-return flows:		
Cedar Point to Reservation Canal	c	2,700
Snake River valley waste to Reservation Canal (estimated)		20,000
Sand Creek to Reservation Canal	c	78,200
Idaho Canal to Blackfoot River	c	30,600
Shull Lateral waste	c	2,200
End of East Idaho Slough into Blackfoot River	c	25,500
Corbett Slough waste to Snake River	c	3,200
Blackfoot Canal waste to Snake River	c	10,200
Total		172,600
Diversions minus surface return		527,900

Irrigation Area 9.—*Diversions from the Snake and Blackfoot Rivers*

Name		*Quantity (acre-feet)*
Little Indian Creek	c	10,500
Fort Hall Main	c	178,900
Fort Hall North	c	70,200
Total		259,600
Surface-return flows:		
End of Fort Hall North	c	2,500
End of Gibson	c	2,100

Irrigation Area 9.—*Diversions from the Snake and Blackfoot Rivers*—Continued

Teak Lateral to Ross Fork	c	600
Indian Lateral to Ross Fork	c	700
Ross Fork downstream from Fort Hall Main	c	3,600
Tyhee waste to Ross Fork	c	13,000
Reider waste	c	2,000
Dubois Lateral waste	c	800
Tyhee Lateral waste	c	2,000
Church Lateral waste	c	2,700
End of Fort Hall Main	c	2,300
Total		32,300
Diversions minus surface return		227,300

Irrigation Area 10.—*Right-bank diversions from the Snake River downstream from Shelley to Blackfoot*

Name		*Quantity (acre-feet)*
New Lava Side	a	35,200
Peoples	a	109,000
Aberdeen–Springfield	a	312,000
Riverside	a	33,600
Danskin	a	58,800
Trego	a	17,700
Wearyrick	a	18,500
Watson	a	31,400
Parsons	a	14,500
Total		630,700
Surface-return flows:		
Riverside waste	c	15,500
Watson Slough waste	c	9,400
Peoples waste	c	8,700
Duncan waste	c	5,200
New Lava Side waste	c	4,500
Parsons waste	c	1,900
Crawford waste	c	2,400
Total		47,600
Diversions minus surface return minus canal loss (Aberdeen–Springfield) of 95,200 acre-ft		487,900

Irrigation Area 11.—*Left-bank diversions from Portneuf River*

Name		*Quantity (acre-feet)*
Fort Hall Michaud	c	30,600
Falls Irrigation	c	23,200
Bannock Creek	g	54,600
Total		108,400
Surface-return flows:		
Bannock Creek	g	34,500
Diversions minus surface return		73,900

Irrigation Area 12.—*Right-bank diversion from the Snake River at Lake Walcott*

		Quantity (acre-feet)
Diversion	a	385,900
Surface-return flows	g	47,400
Diversion minus surface return		338,500

Irrigation Area 13.—*Right-bank diversion from the Snake River at Lake Milner*

		Quantity (acre-feet)
Diversion	g	50,500
Surface-return flows	g	1,700
Diversion minus surface return		48,800

Irrigation Area 14.—*Right-bank diversions from the Snake River at Lake Milner*

Name		Quantity (acre-feet)
North Side Twin Falls	a	697,300
North Side Crosscut–Gooding	g	354,200
North Side "A" Lateral	a	18,100
PA Lateral	a	15,200
Total		1,084,800
Surface-return flows	g	62,700
Diversions minus surface return		1,022,100

Irrigation Area 15.—*Left-bank diversions from the Snake River at Lake Milner*

Name		Quantity (acre-feet)
South Side Twin Falls	a	1,090,200
Rock Creek	g	25,000
Dry Creek	g	9,000
Cedar Creek	g	8,300
Cottonwood, McMullen, Deep Creeks	g	15,000
Total		1,147,500
Surface-return flows	g	575,600
Diversions minus surface return		571,900

Irrigation Area 16.—*Left-bank diversion from the Snake River at Lake Milner*

		Quantity (acre-feet)
Diversion	g	61,100
Surface-return flows	g	500
Diversion minus surface return		60,600

Irrigation Area 17.—*Left-bank diversion from the Snake River at Lake Walcott*

		Quantity (acre-feet)
Diversion	a	312,300
Surface-return flows	g	66,500
Diversion minus surface return		245,800

Irrigation Area 18.—*Goose Creek diversion from Goose Creek Reservoir*

		Quantity (acre-feet)
Diversion	b	44,900
Surface-return flows		0
Diversion minus surface return		44,900

Irrigation Areas 19–26.—*Milner–Gooding Canal, Big Wood and Little Wood Rivers*

Records of measured flows in irrigation areas 19–26 are from Water Districts 37, 37M (1980), American Falls District No. 2 (written commun., 1981), and U.S. Geological Survey (1980). The approach was to sum the inflow and outflow for each irrigation area and determine the difference. This approach includes river and canal losses and field seepage. The total flow consumed in the basin was compared with the total consumed in six of the eight subbasin areas.

Name		Quantity (acre-feet)
Inflow:		
Big Wood River downstream from Magic Reservoir	b	314,100
Little Wood River near Carey	b	140,500
Silver Creek at Sportsman Access	b	114,100
Milner–Gooding Canal upstream from Little Wood River	b	335,400
X Canal	d	101,100
Total		1,005,200
Outflow:		
Big Wood River near Gooding	b	202,200
Y Canal	d	47,600
X Canal	d	22,200
Dietrich Canal	d	56,700
Total		328,700
Basin inflow minus basin outflow		676,500

Total of subbasin consumption:

Area	Inflow–Outflow (acre-feet)
19	67,100
20	62,600
21	226,100
22	106,800

Irrigation Areas 19–26.—*Milner–Gooding Canal, Big Wood and Little Wood Rivers*—Continued

25	94,700
26	129,200
Total	686,500

686,500−676,500/686,500×100=1.5 percent difference

Irrigation Area 19.—*South Gooding tract*

Name		Quantity (acre-feet)
Inflow:		
Little Wood River at Shoshone	d	168,700
X Canal	d	101,100
Big Wood River near Gooding No. 9	d	69,300
Total		339,100
Outflow:		
Big Wood River near Gooding No. 21	d	202,200
Y Canal	d	47,600
Z Canal	d	22,200
Total		272,000
Inflow minus outflow		67,100

Irrigation Area 20.—*North Gooding tract*

Name		Quantity (acre-feet)
Inflow:		
Head of North Gooding Main	d	62,600
Outflow		0
Inflow minus outflow		62,600

Irrigation Area 21.—*Shoshone tract*

Name		Quantity (acre-feet)
Inflow:		
Big Wood River downstream from Diversion No. 5	d	164,700
Milner–Gooding Canal downstream from Little Wood River	d	193,300
Total		358,000
Outflow:		
Head of North Gooding Main	d	62,600
Big Wood River near Gooding No. 9	d	69,300
Total		131,900
Inflow minus outflow		226,100

Irrigation Area 22.—*Lower Little Wood River*

Name		Quantity (acre-feet)
Inflow:		
Little Wood River near Richfield, non-irrigation season—estimated from historical records	g	60,000
Little Wood River near Richfield, irrigation season	d	65,400
JB Slough near Richfield	d	40,300
Marley Slough	d	20,300
Historic F–waste	h	4,100
Milner–Gooding Canal upstream from Little Wood River	d	335,400
Total		525,500
Outflow:		
Dietrich Canal No. 11	d	56,700
Milner–Gooding Canal downstream from Little Wood River	d	193,300
Little Wood River at Shoshone	d	168,700
Total		418,700
Inflow minus outflow		106,800

Irrigation Area 23.—*Dietrich tract*

Name		Quantity (acre-feet)
Inflow:		
Head of Dietrich Canal	d	56,700
Milner–Gooding diversion	e	16,600
Total		73,300
Outflow:		
Historic F–waste	h	4,100
Inflow minus outflow		69,200

Irrigation Area 24.—*Hunt tract*

		Quantity (acre-feet)
Inflow	e	36,000
Outflow		0
Inflow minus outflow		36,000

Irrigation Area 25.—*Richfield tract*

Name		Quantity (acre-feet)
Inflow:		
Head of Richfield Canal	d	159,300

Irrigation Area 25.—*Richfield tract*—Continued

Outflow:		
JB Slough near Richfield	d	40,300
Marley Slough	d	20,300
Sum of miscellaneous wastes	h	4,000
Total		64,600
Inflow minus outflow		94,700

Irrigation Area 26.—*Silver Creek, Upper Little Wood River diversions*

Name		*Quantity (acre-feet)*
Inflow:		
Silver Creek at Sportsman Access	b	114,100
Little Wood River near Carey	b	140,500
Total		254,600
Outflow:		
Little Wood River near Richfield, non-irrigation season—estimated from historical records		60,000
Irrigation season	d	65,400
Total		125,400
Inflow minus outflow		129,200

APPENDIX B.—*Ground-water recharge*

This appendix lists ground-water recharge-rate calculations for 5-year intervals from 1928 to 1980.

Area No.	1928–30 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1929 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	32.37	35	21.04	3.95	1.0	4.33
2	268.83	33	180.12	37.17	1.1	3.75
3	414.30	14	356.30	38.50	1.2	8.05
4	341.60	43	194.71	47.49	1.3	2.80
5	194.13	52	93.18	17.39	1.3	4.06
6	1,538.40	13	1,338.41	134.73	1.3	8.63
7	348.43	20	278.74	70.82	1.3	2.64
8	593.10	20	474.48	92.97	1.3	3.80
9	(1)	20		63.71	1.5	
10	586.07	20	468.86	143.69	1.5	1.76
11	(1)	20		.41	1.5	
12	515.60	12	453.73	79.45	1.6	4.11
13	(1)	3		3.61	1.6	
14	1,205.23	7	1,120.86	284.40	1.6	2.34
15	1,310.97	34	865.24	297.49	1.6	1.31
16	34.00	1	33.66	6.94	1.6	3.25
17	317.00	18	259.94	80.84	1.6	1.62
18	35.33	0	35.33	16.30	1.6	.57
19			(1)	41.47	1.6	
20			(1)	13.75	1.6	
21			(1)	6.01	1.6	
22			(1)	9.12	1.6	
23			(1)	17.16	1.6	
24			(1)	5.63	1.6	
25			(1)	19.65	1.6	
26			93.37	30.10	1.6	1.50

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1931–35 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1929–45 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	23.60	35	15.34	4.45	1.0	2.45
2	223.24	33	149.57	37.38	1.1	2.90
3	403.16	14	346.72	39.61	1.2	7.55
4	305.22	43	173.98	45.46	1.3	2.53
5	170.56	52	81.87	25.19	1.3	1.95
6	1,399.22	13	1,217.32	135.83	1.3	7.66
7	304.26	20	243.41	72.73	1.3	2.05
8	567.52	20	454.02	93.45	1.3	3.56
9	(1)	20		61.95	1.5	
10	538.68	20	466.94	159.21	1.5	1.43
11	(1)	20		.21	1.5	
12	418.50	12	368.28	84.11	1.6	2.78
13	(1)	3		5.27	1.6	
14	1,056.60	7	982.64	295.09	1.6	1.73
15	1,257.92	34	830.23	267.05	1.6	1.51
16	37.24	1	36.87	12.85	1.6	1.27
17	308.98	18	253.36	75.62	1.6	1.75
18	26.66	0	26.66	15.58	1.6	.11
19			(1)	39.49	1.6	
20			28.90	20.81	1.6	0
21			(1)	21.10	1.6	
22			(1)	6.74	1.6	
23			44.08	15.05	1.6	1.33
24			(1)	3.87	1.6	
25			63.05	21.12	1.6	1.39
26			97.08	26.37	1.6	2.08

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1936–40 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	Average 1929–45 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	36.22	35	23.54	4.45	1.0	4.29
2	263.00	33	176.21	37.38	1.1	3.61
3	432.56	14	372.00	39.61	1.2	8.19
4	351.84	43	200.55	45.46	1.3	3.11
5	194.84	52	93.52	25.19	1.3	2.41
6	1,485.64	13	1,292.51	135.83	1.3	8.22
7	352.32	20	281.86	72.73	1.3	2.58
8	650.28	20	520.22	93.45	1.3	4.27
9	([1])	20		61.95	1.5	
10	625.82	20	500.66	159.21	1.5	1.64
11	([1])	20		.21	1.5	
12	473.72	12	416.87	84.11	1.6	3.36
13	([1])	3		5.27	1.6	
14	1,176.66	7	1,094.29	295.09	1.6	2.11
15	1,269.80	34	838.07	267.05	1.6	1.54
16	41.46	1	41.05	12.85	1.6	1.59
17	345.08	18	282.97	75.62	1.6	2.14
18	33.14	0	33.14	15.58	1.6	.53
19			108.04	39.49	1.6	1.14
20			46.88	20.81	1.6	.65
21			160.06	21.10	1.6	5.99
22			94.52	6.74	1.6	12.42
23			55.60	15.04	1.6	2.10
24			([1])	3.87	1.6	
25			77.54	21.12	1.6	2.07
26			97.88	26.37	1.6	2.11

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1941–45 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1945 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	43.84	35	28.50	4.95	1.0	4.76
2	272.90	33	182.84	37.58	1.1	3.77
3	418.82	14	360.19	40.71	1.2	7.65
4	334.98	43	190.94	43.42	1.3	3.10
5	185.22	52	88.91	32.98	1.3	1.40
6	1,449.26	13	1,260.86	136.92	1.3	7.91
7	357.98	20	286.38	74.64	1.3	2.54
8	625.44	20	500.35	93.93	1.3	4.03
9	(1)	20		60.18	1.5	
10	656.64	20	525.31	174.73	1.5	1.51
11	(1)	20		0	1.5	
12	433.98	12	381.90	88.76	1.6	2.70
13	(1)	3		6.93	1.6	
14	1,218.04	7	1,132.78	305.78	1.6	2.10
15	1,233.90	34	814.37	236.61	1.6	1.84
16	45.24	1	44.79	18.75	1.6	.79
17	319.60	18	262.07	70.39	1.6	2.12
18	45.36	0	45.36	14.86	1.6	1.45
19			87.84	37.50	1.6	.74
20			53.96	27.86	1.6	.34
21			171.68	36.19	1.6	3.14
22			119.72	4.36	1.6	25.86
23			64.00	12.93	1.6	3.35
24			(1)	2.10	1.6	
25			87.20	22.59	1.6	2.26
26			80.66	22.63	1.6	1.96

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1946–50 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1945 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	56.60	35	36.79	4.95	1.0	6.43
2	286.16	33	191.73	37.58	1.1	4.00
3	431.88	14	371.42	40.71	1.2	7.92
4	342.10	43	195.00	43.42	1.3	3.19
5	201.32	52	96.63	32.98	1.3	1.63
6	1,496.36	13	1,301.83	136.92	1.3	8.21
7	412.40	20	329.92	74.64	1.3	3.12
8	669.48	20	535.58	93.93	1.3	4.40
9	(1)	20		60.18	1.5	
10	700.64	20	560.51	174.73	1.5	1.71
11	(1)	20		0	1.5	
12	440.24	12	387.41	88.76	1.6	2.76
13	(1)	3		6.93	1.6	
14	1,276.98	7	1,187.59	305.78	1.6	2.28
15	1,263.60	34	833.98	236.61	1.6	1.92
16	52.62	1	52.09	18.75	1.6	1.18
17	351.54	18	288.26	70.39	1.6	2.50
18	41.58	0	41.58	14.86	1.6	1.20
19			96.40	37.50	1.6	.97
20			58.46	27.86	1.6	.50
21			189.92	36.19	1.6	3.65
22			105.48	4.36	1.6	22.59
23			69.10	12.93	1.6	3.74
24			(1)	2.10	1.6	
25			98.36	22.59	1.6	2.75
26			102.26	22.63	1.6	2.92

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1951–55 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1959 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	65.82	35	42.78	29.67	1.0	0.44
2	297.16	33	199.10	39.04	1.1	4.00
3	447.40	14	384.76	28.31	1.2	12.39
4	362.06	43	206.37	36.81	1.3	4.31
5	218.66	52	104.96	22.80	1.3	3.30
6	1,584.54	13	1,378.55	126.38	1.3	9.61
7	429.44	20	343.55	78.73	1.3	3.06
8	686.70	20	549.36	94.91	1.3	4.49
9	([1])	20		64.00	1.5	
10	725.62	20	580.50	147.04	1.5	2.45
11	([1])	20		20.45	1.5	
12	450.08	12	396.07	80.52	1.6	3.32
13	([1])	3		29.60	1.6	
14	1,319.32	7	1,226.97	190.77	1.6	4.83
15	1,305.78	34	861.81	256.22	1.6	1.76
16	57.26	1	56.69	4.15	1.6	12.06
17	364.40	18	298.81	60.98	1.6	3.30
18	43.38	0	43.38	19.38	1.6	.64
19			86.40	28.53	1.6	1.43
20			61.72	18.06	1.6	1.82
21			236.12	29.37	1.6	6.44
22			100.92	11.97	1.6	6.83
23			74.84	23.66	1.6	1.56
24			([1])	24.31	1.6	
25			98.93	26.48	1.6	2.14
26			89.42	22.93	1.6	2.30

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1956–60 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1959 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	70.68	35	45.94	29.67	1.0	0.55
2	334.62	33	224.20	39.04	1.1	4.64
3	471.32	14	405.34	28.31	1.2	13.12
4	383.06	43	218.34	36.81	1.3	4.63
5	241.18	52	115.77	22.80	1.3	3.78
6	1,646.24	13	1,432.23	126.38	1.3	10.03
7	453.54	20	362.83	78.73	1.3	3.31
8	710.08	20	568.06	94.91	1.3	4.69
9	([1])	20		64.00	1.5	
10	755.94	20	604.75	147.04	1.5	2.61
11	63.26	20	50.61	20.45	1.5	.97
12	460.22	12	404.99	80.52	1.6	3.43
13	53.35	3	51.75	29.60	1.6	.15
14	1,294.20	7	1,203.61	190.77	1.6	4.71
15	1,254.88	34	828.22	256.22	1.6	1.63
16	63.54	1	62.90	4.15	1.6	13.56
17	367.44	18	301.30	60.98	1.6	3.34
18	38.88	0	38.88	19.38	1.6	.41
19			93.85	28.53	1.6	1.69
20			60.80	18.06	1.6	1.77
21			275.78	29.37	1.6	7.79
22			90.20	11.97	1.6	5.94
23			69.12	23.66	1.6	1.32
24			([1])	24.31	1.6	
25			96.74	26.48	1.6	2.05
26			67.42	22.93	1.6	1.34

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1961–65 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1966 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	56.70	35	36.86	21.56	1.0	0.71
2	324.26	33	217.25	39.61	1.1	4.38
3	459.90	14	395.51	42.75	1.2	8.05
4	365.52	43	208.35	62.45	1.3	2.04
5	258.78	52	124.21	57.87	1.3	.85
6	1,497.28	13	1,302.63	139.30	1.3	8.05
7	371.56	20	297.25	81.56	1.3	2.34
8	667.54	20	534.03	100.29	1.3	4.02
9	[1]	20		50.09	1.5	
10	676.82	20	541.46	153.67	1.5	2.02
11	82.64	20	66.11	35.13	1.5	.38
12	418.56	12	368.33	94.68	1.6	2.29
13	52.42	3	50.85	32.48	1.6	0
14	1,140.46	7	1,060.63	229.41	1.6	3.02
15	1,201.90	34	793.25	276.86	1.6	1.27
16	57.84	1	57.26	2.94	1.6	17.88
17	343.10	18	281.34	57.09	1.6	3.33
18	42.20	0	42.20	8.39	1.6	3.43
19			87.88	28.19	1.6	1.52
20			58.58	20.30	1.6	1.29
21			223.84	17.96	1.6	10.86
22			193.24	9.14	1.6	19.54
23			74.02	17.08	1.6	2.73
24			33.40	17.75	1.6	.28
25			86.60	20.27	1.6	2.67
26			71.74	18.82	1.6	2.21

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1966–70 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1966 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	59.86	35	38.91	21.56	1.0	0.80
2	341.90	33	229.07	39.61	1.1	4.68
3	453.82	14	390.29	42.75	1.2	7.93
4	376.64	43	214.68	62.45	1.3	2.14
5	300.84	52	144.40	57.87	1.3	1.20
6	1,611.22	13	1,401.76	139.30	1.3	8.76
7	379.08	20	303.26	81.56	1.3	2.42
8	702.48	20	561.98	100.29	1.3	4.30
9	([1])	20		50.09	1.5	
10	736.10	20	588.88	153.67	1.5	2.33
11	108.30	20	86.64	35.13	1.5	.97
12	470.48	12	414.02	94.68	1.6	2.77
13	52.42	3	50.85	32.48	1.6	0
14	1,290.72	7	1,200.37	229.41	1.6	3.63
15	1,274.34	34	841.06	276.86	1.6	1.44
16	65.74	1	65.08	2.94	1.6	20.54
17	367.26	18	301.15	57.09	1.6	3.68
18	34.02	0	34.02	8.39	1.6	2.45
19			83.10	28.19	1.6	1.35
20			61.58	20.30	1.6	1.43
21			248.64	17.96	1.6	12.24
22			132.30	9.14	1.6	12.87
23			66.32	17.08	1.6	2.28
24			32.38	17.75	1.6	.22
25			106.30	20.27	1.6	3.64
26			81.48	18.82	1.6	2.73

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1971–75 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1979 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	60.88	35	39.57	0	1.0	
2	337.12	33	225.87	35.82	1.1	5.19
3	427.34	14	367.51	31.15	1.2	10.60
4	382.44	43	217.99	29.02	1.3	6.21
5	261.34	52	125.48	24.64	1.3	3.79
6	1,774.38	13	1,543.71	123.42	1.3	11.21
7	398.10	20	318.48	75.31	1.3	2.93
8	751.16	20	600.93	105.84	1.3	4.38
9	(1)	20		38.77	1.5	
10	740.14	20	592.11	103.88	1.5	4.20
11	107.28	20	85.82	39.69	1.5	.66
12	443.92	12	390.65	85.78	1.6	2.95
13	51.44	3	49.90	23.41	1.6	.53
14	1,210.92	7	1,126.16	171.72	1.6	4.96
15	1,165.84	34	769.45	222.04	1.6	1.87
16	62.04	1	61.42	19.33	1.6	1.58
17	352.04	18	288.67	48.82	1.6	4.31
18	70.22	0	70.22	0	1.6	
19			68.47	27.81	1.6	.86
20			65.50	17.71	1.6	2.10
21			307.68	33.42	1.6	7.61
22			100.72	8.00	1.6	10.99
23			80.06	15.76	1.6	3.48
24			41.58	16.45	1.6	.93
25			121.60	31.47	1.6	2.26
26			111.86	24.85	1.6	2.90

APPENDIX B.—*Ground-water recharge*—Continued

Area No.	1976–80 average diversion (thousands of acre-feet)	Percent return	Diversion-return (thousands of acre-feet)	1979 acreage (thousands of acres)	Evapotranspiration (feet per year)	Recharge rate (feet per year)
1	58.34	35	37.92	0	1.0	
2	302.06	33	202.38	35.82	1.1	4.55
3	367.90	14	316.39	31.15	1.2	8.96
4	258.06	43	147.09	29.02	1.3	3.77
5	229.86	52	110.33	24.64	1.3	3.18
6	1,379.10	13	1,199.82	123.42	1.3	8.42
7	355.84	20	284.67	75.31	1.3	2.48
8	695.74	20	556.59	105.84	1.3	3.96
9	259.60	20	207.68	38.77	1.5	3.86
10	657.80	20	526.24	103.88	1.5	3.57
11	112.88	20	90.30	39.69	1.5	.78
12	391.12	12	344.19	85.78	1.6	2.41
13	52.06	3	50.50	23.41	1.6	.56
14	995.84	7	926.13	171.72	1.6	3.79
15	1,116.12	34	736.64	222.04	1.6	1.72
16	62.54	1	61.91	19.33	1.6	1.60
17	297.38	18	243.85	48.82	1.6	3.39
18	38.12	0	38.12	0	1.6	
19			62.88	27.81	1.6	.66
20			57.04	17.71	1.6	1.62
21			209.58	33.42	1.6	4.67
22			77.94	8.00	1.6	8.14
23			62.14	15.76	1.6	2.34
24			32.48	16.45	1.6	.37
25			90.04	31.47	1.6	1.26
26			108.68	24.85	1.6	2.77

[1]Records unavailable.

APPENDIX C.—*Soils*

This appendix lists selected information from the general soils map (U.S. Soil Conservation Service, 1976) used to estimate recharge to the ground-water system. Soils were generalized into three classes on the basis of their hydrologic group and thickness. Hydrologic groups are used to estimate rainfall runoff, which is influenced by depth to a water table or impermeable bedrock, infiltration rate, and depth to layers of low-permeability soils (U.S. Soil Conservation Service, 1976). Four hydrologic groups are defined: (A) low runoff potential; (B) moderately low runoff potential; (C) moderately high runoff potential; and (D) high runoff potential. Depth of soil is the depth to a limiting layer such as bedrock, fragipan, or gravel. Three groups of infiltration-rate potential were used in this report: (1) high infiltration-rate potential (hydrologic groups A and B), little or no soil cover, typically recent lava flows; (2) moderate infiltration-rate potential (hydrologic groups B and C), thin soil cover (less than 40 in.), typically alluvium and thin loess deposits; and (3) low infiltration-rate potential (hydrologic groups C and D), thick soil cover (greater than 40 in.), typically lacustrine and thick loess deposits. The following table lists textural, hydrologic, and thickness information for soil units in the eastern Snake River Plain.

[—, no data available]

Sequence No.	Textural description	Depth to limiting layer (inches)	Hydrologic group	Infiltration-rate group
12	Loamy, skeletal	20–40	B	2
13	Clayey	>40	C, D	3
14	Loamy	>60	D	3
16	Loamy, silty and loamy	20–40	B, D	3
17	Silty, loess	20–40	C, D	3
18	Silty or loamy	>60	B, C	3
21	Sandy, silty	40–60	B	2
22	Loamy, rock outcrops	20–40	C	2
26	Silty, loess	20–40	B, C	3
29	Loamy	>40	C, D	3
30	Silty	20–40	B, C	3
31	Silty, loamy	>40	B, C, D	3
32	Loamy	20–40	B, D	2
33	Loamy	—	B	2
34	Loamy, skeletal	20–40	B	2
35	Skeletal, loamy	20–40	B	2
43	Silty	>60	B, C	3
47	Silty, loamy	>60	B, C, D	3
48	Clayey, loamy, silty	40–60	B, C	3
62	Clayey, sandy, loamy	>40	C	3
64	Skeletal and calcic, clayey	<20	C, D	3
66	Sandy	>60	A	2
67	Sandy, loamy	>60	A, B	2
69	Silty, loess	>60	B	3
70	Loamy	40–60	B	2
72	Skeletal, loamy	<20	B	2
77	Silty	>40	C, B	3
82	Sandy, loamy	>60	A, D	2
91	Skeletal and stony	>60	B	2
112	Loamy	20–40	D	3

APPENDIX C.—*Soils*—Continued

Sequence No.	Textural description	Depth to limiting layer (inches)	Hydrologic group	Infiltration-rate group
118	Silty	>60	D	3
136	Loamy	20–40	B, C	2
160	Clayey	20–40	C	3
191	Loamy	20–40	A, B	2
199	Calcic	20–40	B	2
200	Loamy, silty	>60	B	3
201	Clayey and stony	40–60	C	3
206	Skeletal, clayey	<20	D	3
207	Loamy, skeletal	20	B, C, D	3
215	Loamy	20–40	C, D	2
220	Sandy	>60	A	2
221	Loamy	40	B	2
227	Loamy	40–60	B, C	3
229	Silty	>60	B	3
234	Skeletal	40	A	2
235	Loamy, skeletal	>60	B	3
244	Silty	>60	B	3
246	Sandy	>60	A, D	2
248	Loamy	40	B	2
250	Loamy	>60	B	3
251	Silty or loamy	>60	B, C	3
252	Silty	>60	B	3
257	Loamy	40	B	2
259	Loamy	40	B	2
265	Silty	>60	B, C	3
266	Silty	>60	B, C	3
267	Loamy	20	B, D	2
268	Loamy	40	B	2
271	Loamy	20–60	B, D	3
274	Canyon walls			3
275	Bare lava flows			1
276	Hillslopes, rocky			2
277	Active sand dunes			2
278	Mountains, rocky			2

U S DEPARTMENT OF THE INTERIOR

March 3, 1849