GEOHYDROLOGY AND SIMULATION OF FLOW AND WATER LEVELS IN THE AQUIFER SYSTEM IN THE MUD LAKE AREA OF THE EASTERN SNAKE RIVER PLAIN, EASTERN IDAHO

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 93–4227

Prepared in cooperation with the

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By JOSEPH M. SPINAZOLA

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
acre	4,047	square meter
acre-foot (acre-ft)	1,233	cubic meter
foot (ft)	0.3048	meter
foot per mile (ft/mi)	.1894	meter per kilometer
foot squared per day[1] (ft^2/d)	.0929	meter squared per day
cubic foot per day (ft^3/d)	.02832	cubic meter per day
cubic foot per foot squared per foot per day [(ft^3/ft^2)/ft/d]	1	cubic meter per meter squared per meter per day
gallon per minute (gal/min)	.003785	cubic meter per minute
gallon per minute per foot [(gal/min)/ft]	.01242	cubic meter per minute per meter
inch (in.)	25.4	millimeter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

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

[1] The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness. This mathematical expression reduces to foot squared per day, which is used in this report.

GEOHYDROLOGY AND SIMULATION OF FLOW AND WATER LEVELS IN THE AQUIFER SYSTEM IN THE MUD LAKE AREA OF THE EASTERN SNAKE RIVER PLAIN, EASTERN IDAHO

By Joseph M. Spinazola

ABSTRACT

Water users rely on surface water and ground water to irrigate crops and to maintain lakes on wildlife refuges in the 2,200-square-mile Mud Lake study area. Ground-water development between the late 1970's and 1989 increased withdrawals from about 240,000 acre-feet in 1983 to about 370,000 acre-feet in 1990. Concurrent with ground-water development, change from subirrigation to sprinkler irrigation was predicted to reduce recharge by 95,000 acre-feet, according to an independent study. Of the 660,000 acre-feet total estimated recharge from precipitation and irrigation in the study area in 1980, half was in the area in which irrigation methods were changed. Water managers need the ability to evaluate the effects of water-use changes on the future supply of surface water and ground water.

Basalt and rhyolite predominate on the surface and in the subsurface of the study area. Total basalt thickness is less than 4,000 feet; total sediment thickness (clay, silt, sand, and gravel) is less than 1,000 feet. Basalt and sediment interbeds contribute to confined ground-water conditions and affect movement and supply of water in parts of the aquifer system.

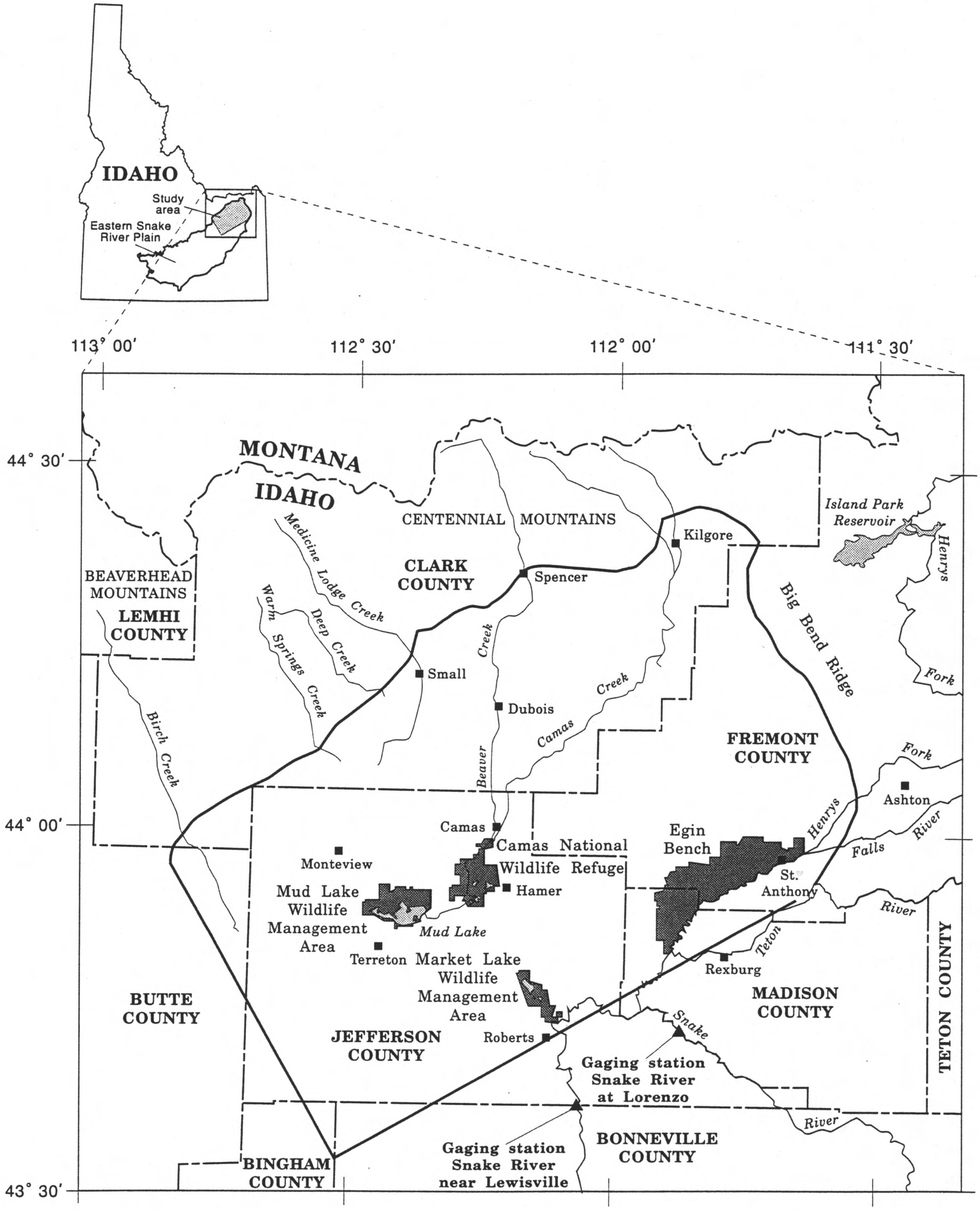
Estimated losses from and gains to perennial streams and lakes in 1980 were each about 110,000 acre-feet. Water-table altitudes ranged from about 4,500 to 6,200 feet above sea level, and water-table gradients were 3 to 120 feet per mile. Underflow from basins tributary to the study area was estimated to be about 450,000 acre-feet in 1980; measured discharge from flowing wells was about 10,000 acre-feet.

A five-layer, three-dimensional, finite-difference, numerical ground-water flow model was calibrated by trial-and-error to assumed 1980 steady-state hydrologic conditions to obtain a better understanding of the geohydrology and provide a tool to evaluate water-use alternatives. Water-level gradients simulated by the model were similar to gradients measured in 1980. Simulated underflow across model boundaries for 1980 was 932,000 acre-feet. Simulated losses from and gains to most streams and lakes were within 2 percent of estimated values. Simulated discharge from flowing wells matched measurements for 1980. An attempt to calibrate the numerical model to transient hydrologic conditions in monthly increments from 1981 to 1990 was discontinued because available data did not justify changes that were indicated by model simulations.

INTRODUCTION

Irrigators, wildlife managers, and others depend on an adequate supply of surface and ground water for agriculture, wildlife, and other uses in the Mud Lake area in the northernmost part of the eastern Snake River Plain (fig. 1). Most cultivated agricultural land in the area is irrigated with water pumped from wells completed in the eastern Snake River Plain aquifer system. Lakes within the Mud Lake Wildlife Management Area (WMA), Camas National Wildlife Refuge, and Market Lake WMA provide habitat for migratory waterfowl and native flora and fauna. Mud Lake WMA and Camas National Wildlife Refuge rely on streamflow from Beaver and Camas Creeks, natural ground-water inflow, and ground-water withdrawals to fill and maintain area lakes. Market Lake WMA is maintained solely by natural ground-water inflow.

Changes in water use have contributed to concern by many water users about an adequate future supply of surface and ground water in the 2,200-mi^2 study area. Many tracts of land were converted to agricultural use between the late 1970's and 1989. These tracts were developed with irrigation systems that relied on ground water for supply. Concurrently, decreased reliance on subirrigation and the systematic conversion to sprinkler irrigation on Egin Bench (fig. 1) were predicted to result in about 95,000 acre-ft less recharge



Base from U.S. Geological Survey, 1:100,000 quadrangles
Ashton, 1976; Circular Butte, 1979;
Dubois, 1978; and Rexburg, 1978
Projection, Universal Transverse Mercator, Zone 12

0 5 10 15 20 MILES
0 5 10 15 20 KILOMETERS

Figure 1. Location of study area.

to the Snake River Plain aquifer (King, 1987, p. 21). The need to evaluate the consequences of increased development and reduced recharge on future water levels and water supply led to a cooperative agreement among the U.S. Geological Survey (USGS), the Idaho Department of Water Resources (IDWR), and the U.S. Department of Energy. That agreement resulted in a 3-year study that began in the spring of 1989.

Purpose and Scope

The purpose of this report is to describe the geohydrology and to document the calibration of a three-dimensional, finite-difference, numerical ground-water flow model of the aquifer system in the Mud Lake area. Geohydrologic descriptions include those of surficial and subsurface geology; surface-water supply and use; ground-water occurrence, recharge, and discharge; and aquifer properties. Geohydrologic data were documented for conditions from 1980 to 1990 and were processed to develop the data sets used to calibrate the numerical model and to simulate the response of the aquifer system to several water-use alternatives. A companion report (Spinazola, 1994) describes the use of the model to simulate that response.

Two basic assumptions were made in this study to simplify the complexity of the natural geohydrologic processes that occur in the aquifer system and facilitate simulation with the numerical model. Geohydrologic data were assumed to be adequate to describe critical elements in the hydrologic regimen, and the numerical model was assumed to provide an adequate representation of the hydrologic regimen. The accuracy of the simulations made with the model is related to the validity of these assumptions.

Geohydrologic data were compiled to different levels of detail for this study. However, special mention is made to differentiate Egin Bench, the Henrys Fork of the Snake River, and the area south of Henrys Fork from the remainder of the study area. These areas were not included in the original study plan. Late in the final year of this study, the IDWR and USGS agreed that study results would be more useful if these areas were included. The hydrologic regimen in these areas was recognized to be complex, and resources were not available to analyze them with the same level of effort expended in the remainder of the study area. Therefore, only readily available data were used to describe the hydrologic regimen in these areas.

Site-Numbering System

Surface-water gaging stations are referred to by station name in this report. Station names are included in annual reports of water resources data published by the USGS.

The well-numbering system used by the USGS in Idaho indicates the locations of wells within the official rectangular subdivision of public lands, with reference to the Boise base line and Meridian. The first segment (7N) of well number 7N-38E-23DBA3 (fig. 2) designates the township north or south; the second (38E), the range east or west; and the third (23), the section number in which the well is located. Letters (DBA) following the section number indicate the well's location within the section and are assigned in counterclockwise order beginning with the northeast quarter. The first letter (D) denotes the 1/4 section (160-acre tract), the second (B) denotes the 1/4-1/4 section (40-acre tract), and the third (A) denotes the 1/4-1/4-1/4 section (10-acre tract). The last number (3) is a serial number assigned when the site was inventoried.

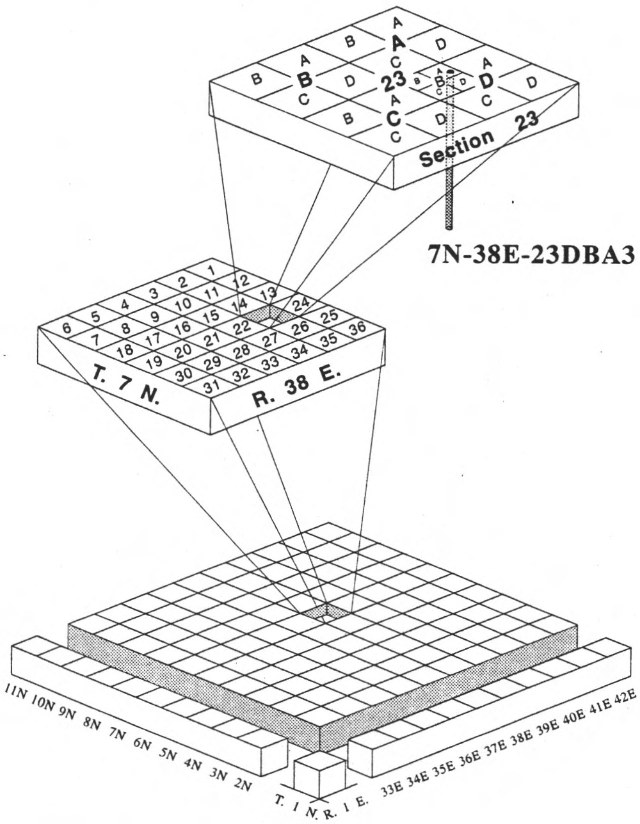


Figure 2. Well-numbering system.

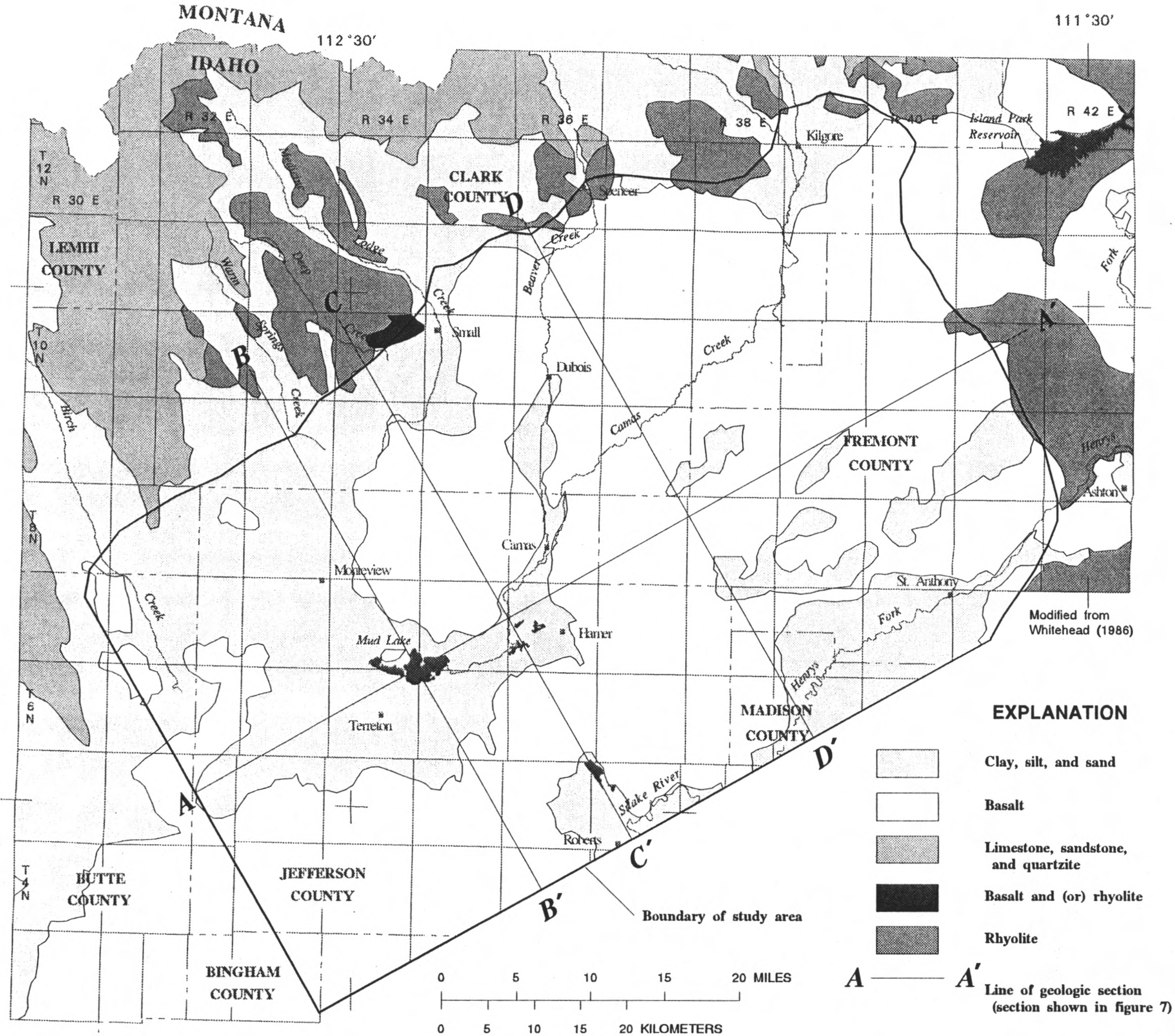


Figure 3. Surficial geology.

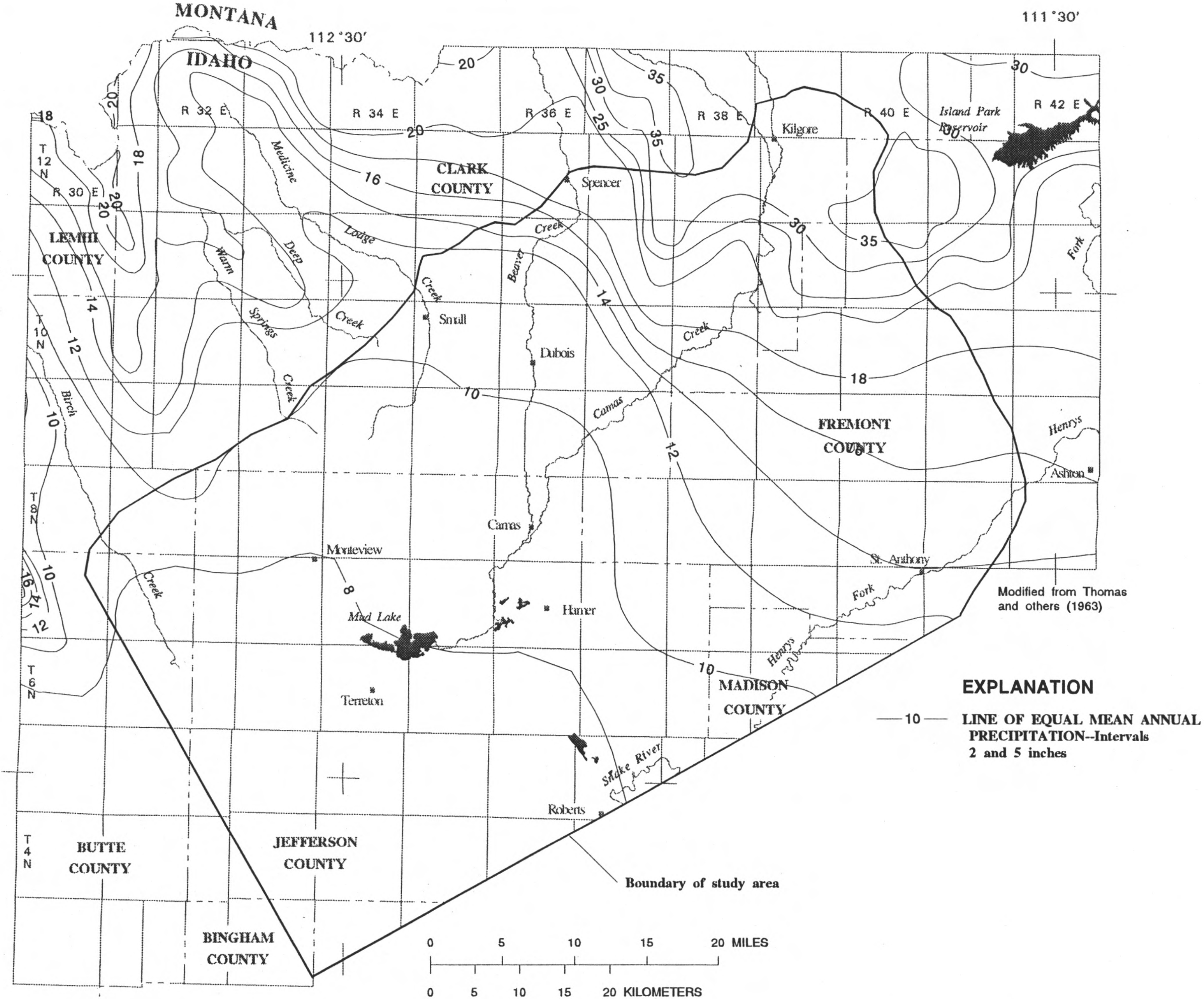


Figure 4. Mean annual precipitation, 1930–57.

Physiography and Precipitation

The study area lies entirely within the eastern Snake River Plain (fig. 1). Basalt, overlain by a discontinuous veneer of fine-grained sediments in much of the area, is the predominant rock (fig. 3). Sediments (clay, silt, sand, and gravel) are present where tributary valleys intersect the northwestern margin of the plain, along the Henrys Fork and Snake River, and around Mud Lake. The surface of the plain ranges from about 4,790 ft above sea level at Mud Lake to about 6,300 ft near Kilgore. The plain is bounded to the northwest by the Beaverhead and Centennial Mountains and to the northeast by Big Bend Ridge (fig. 1). Peaks in the surrounding mountains reach heights of 11,000 ft.

Several streams flow into the study area (fig. 1). Clockwise from the northwest corner, water in Birch, Warm Springs, and Deep Creeks is diverted for irrigation or percolates into the subsurface outside of the study area and rarely flows onto the plain. Medicine Lodge Creek flows southward onto the plain where its water then sinks into the subsurface. Camas Creek and its primary tributary, Beaver Creek, flow southwestward across the plain to Mud Lake. Henrys Fork of the Snake River and the Snake River flow through the southern part of the study area. Mud Lake, the terminus of a closed basin (no natural surface-water outlet), holds 61,000 acre-ft of water at its maximum capacity. Several smaller lakes vary in size from year to year depending upon the supply of surface and ground water. In unusually wet years, the area around Mud Lake becomes flooded in the spring when the lake cannot contain flow from Camas Creek.

Precipitation on areas within and adjacent to the study area determines the supply of surface and ground water in the study area. Mean annual precipitation on the plain during 1930–57 ranged from about 8 in. near Mud Lake to about 35 in. near Kilgore (fig. 4). Annual precipitation at Dubois (fig. 5) ranged from 9.39 in. during 1988 to 20.6 in. during 1983 and averaged 13.9 in. (National Climatic Data Center, monthly reports, January 1980 through December 1990). Monthly precipitation at Dubois for the 1980–90 period ranged from zero during September 1987 and October 1988 to 4.47 in. during May 1980. Annual precipitation in the surrounding mountains sometimes exceeds 60 in., mostly in the form of snow.

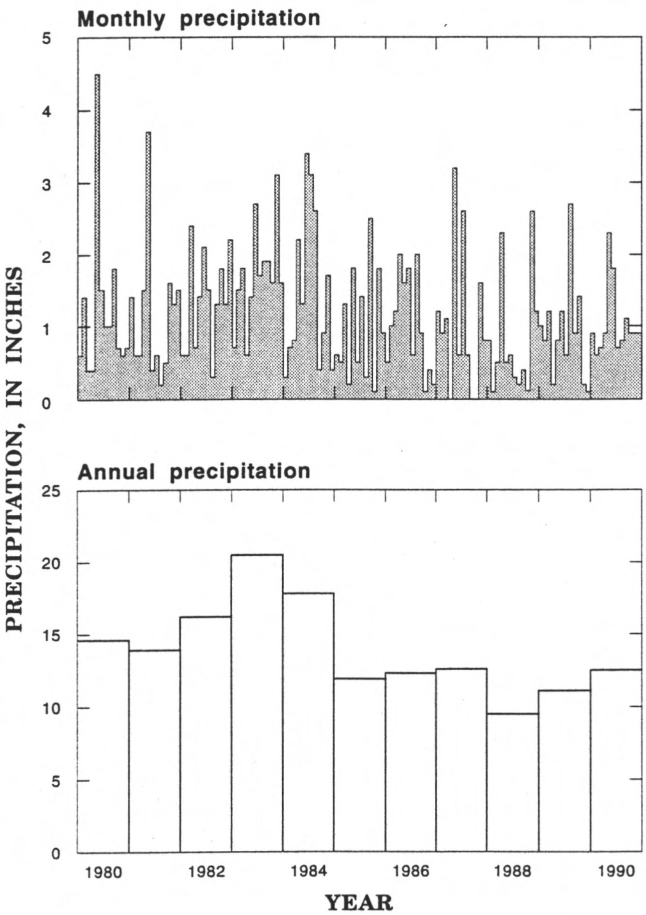
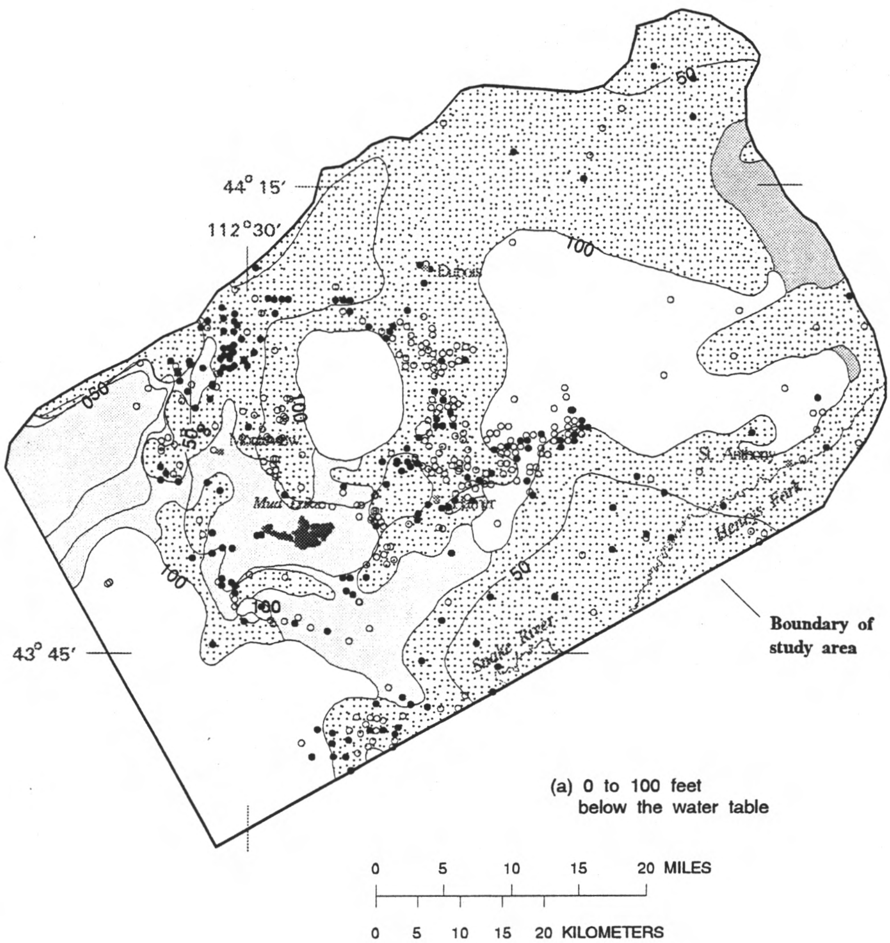


Figure 5. Precipitation at Dubois, Idaho, 1980–90.

Previous Studies

Studies of the entire eastern Snake River Plain that included descriptions of geohydrologic conditions in the Mud Lake area were made by Russell (1902), Stearns and others (1938), Mundorff and others (1964), Whitehead (1986), Bigelow and others (1987), Goodell (1988), Lindholm and others (1988), and Kjelstrom (in press). Several studies that included development of analog or numerical ground-water flow simulation models of the eastern Snake River Plain aquifer included descriptions of aquifer properties in the Mud Lake area. Electrical analog model studies are described in reports by Skibitzke and da Costa (1962), Norvitch and others (1969), and Mantei (1974). Numerical model studies are described in reports by deSonneville (1974), Newton (1978), and Garabedian (1992).

Among the numerous studies that included descriptions of geohydrologic conditions in the Mud



EXPLANATION

- Clay, silt, and sand
- Sand and gravel
- Basalt
- Rhyolite
- Dense, older basalt

— 50 — Line of equal thickness of saturated basalt. Intervals 50, 100, and 500 feet

● Well with driller's log that fully penetrates indicated depth interval

○ Well with driller's log that partially penetrates indicated depth interval

Figure 6. Distributions of predominant rock types for selected depth intervals below the water table.

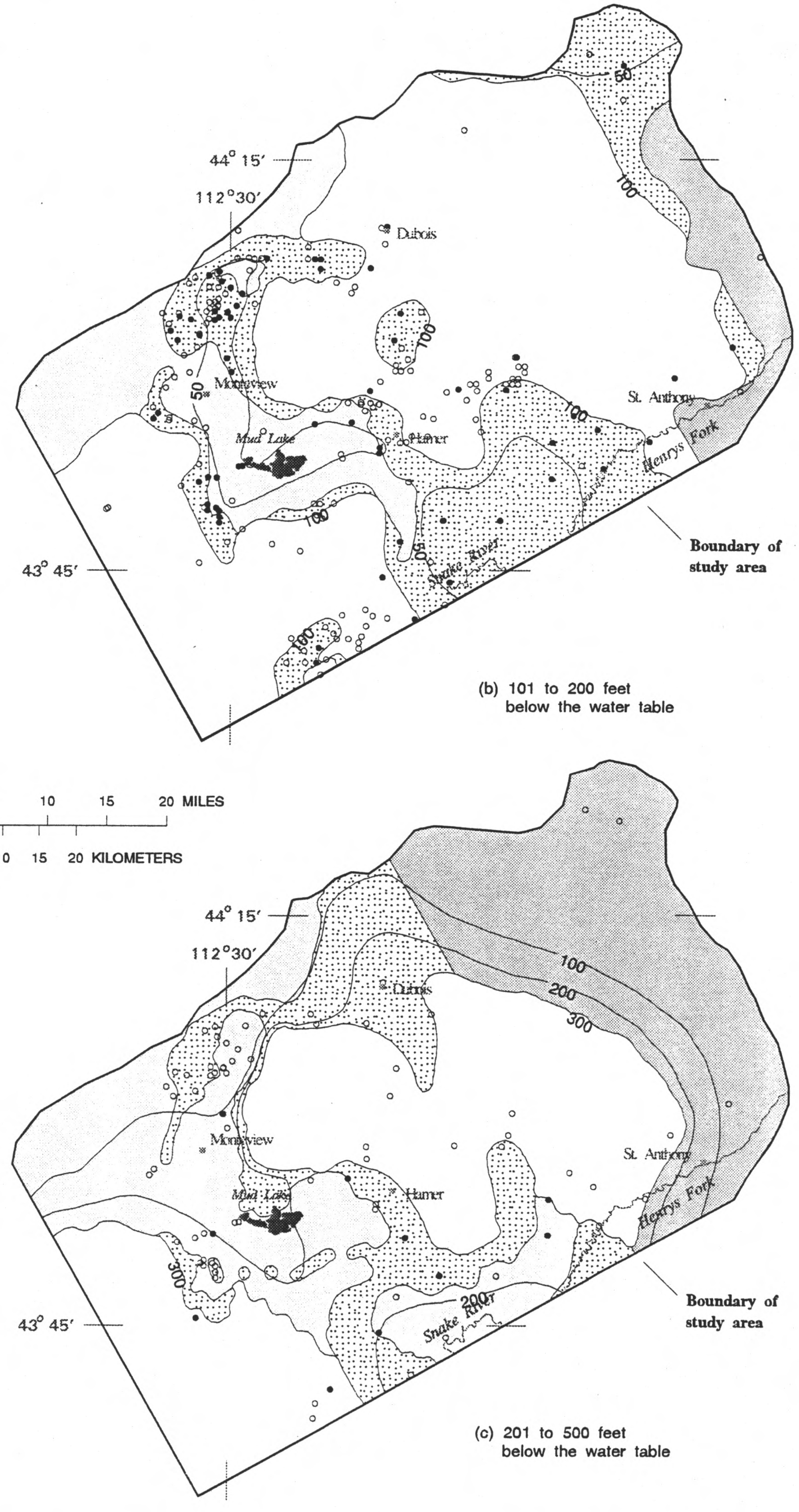


Figure 6. Distributions of predominant rock types for selected depth intervals below the water table—Continued.

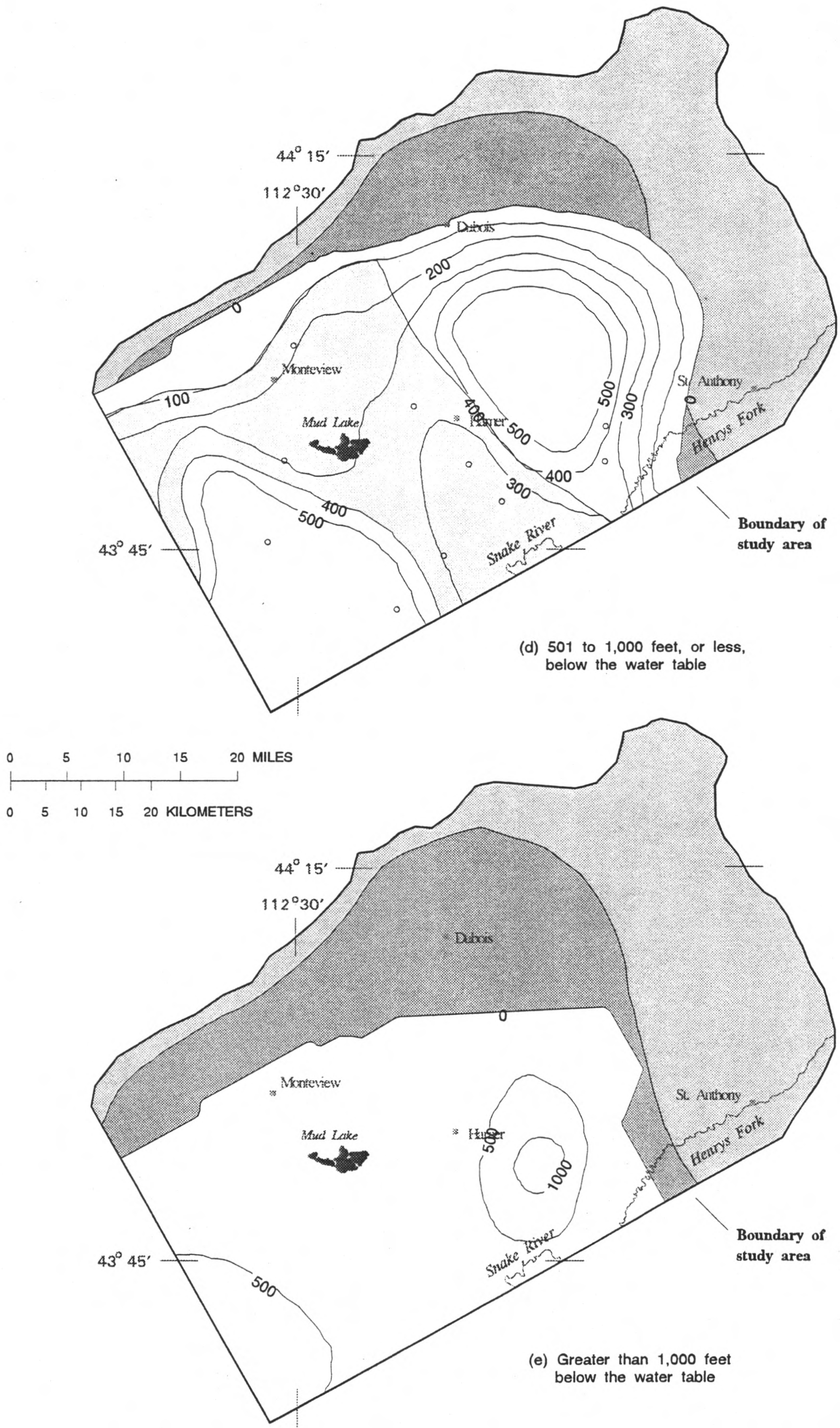


Figure 6. Distributions of predominant rock types for selected depth intervals below the water table—Continued.

Lake area exclusively, the most comprehensive is by Stearns and others (1939). Geology and hydrology were characterized by Luttrell (1982) to provide background for a numerical ground-water flow model by Johnson and others (1984). The effect of conversion from subirrigation to sprinkler irrigation on recharge to the Snake River Plain aquifer was described in a report by King (1987). A surface-water budget model for Beaver Creek, Camas Creek, and the Mud Lake area is described in a report by Brockway and Robison (1988). Basic data collected as part of this study were reported by Spinazola and others (1992).

Studies that describe geohydrologic conditions at the Idaho National Engineering Laboratory (INEL) by Barraclough and others (1967), Robertson (1977), Ackerman (1991), Anderson (1991), and Cecil and others (1991) include information that relates to conditions in the Mud Lake area. The boundary of the INEL crosses the southwestern boundary of the study area; most facilities on the site are in Butte County.

Acknowledgments

Messrs. Alan Robertson, Robert Sutter, Anthony Morse, Robert Harmon, and James Johnson of the IDWR provided data on streamflow diversions and return flows, land use, well permits, and drillers' logs. With few exceptions, landowners, canal companies, and water users in the study area allowed access to their property for survey work and hydrologic measurements. Mr. Paul Bean graciously allowed use of his wells for an aquifer test. Utah Power and Light Company provided records used to determine ground-water withdrawals and assisted in correlating those records with field locations. Although many individuals offered information and assistance during this study, special thanks are extended to Mr. Donald Shenton, Watermaster of Water District 31.

GEOHYDROLOGY

The first part of this report presents a description of the geohydrologic framework of the aquifer system in the Mud Lake study area. The geohydrologic framework was developed from evaluation of the data compiled during the study and includes descriptions of the geologic and hydrologic settings. The geologic setting includes descriptions of the surface and subsurface rocks that store and transmit water. The hydrologic setting includes descriptions of surface water and ground water. Most of the geohydrologic data presented in the first part of this report were used as, or contributed to, inputs to the numerical model described in the second part, or were used to compare with model results.

Geologic Setting

The surface of the eastern Snake River Plain consists of volcanic rocks and alluvial and windblown sediments (fig. 3). Volcanic rocks consist mainly of basalt and rhyolite; sediments consist mainly of clay, silt, and sand. Mountains that surround the plain are composed of consolidated sedimentary and metamorphic rocks, mainly limestone, sandstone, and quartzite. Detailed descriptions of the area geology and geologic history are provided in reports by Stearns and others (1939) and Luttrell (1982).

Basalt flows from numerous vents on the plain and erosion and deposition of sediments over several hundreds of thousands of years have resulted in a complex subsurface geology. For example, 40 separate basalt flows were identified to a depth of several hundred feet below land surface at the INEL several miles southwest of the study area (Anderson, 1991, p. 1). Flows were identified mostly in the unsaturated zone from natural gamma and drillers' logs and core samples. In the study area, mapping separate basalt flows in the saturated or unsaturated zone was impractical because of lack of natural gamma logs and core samples and the poor distribution of drillers' logs for wells that penetrated more than 200 ft below the water table. Instead of detailed mapping of individual basalt flows and sediment layers, distributions of sediments, basalt, rhyolite, and older basalt were prepared for several depth intervals below the water table (fig. 6). The maps in figure 6 were modified from earlier maps (Garabedian, 1992, pl. 5) by inclusion of all drillers' logs on file with the IDWR. Most modifications were for intervals for which drillers' logs were abundant—those from 0 to 200 ft below the water table. Below 200 ft, the number of drillers' logs and modifications to the original maps decreased. Maps show basalt thickness and the predominant rock type among several volcanic and sedimentary rock types present in the subsurface. Correlations with geologic sections (Whitehead, 1986, sheet 2) were used to approximate the boundary between older basalt and rhyolite.

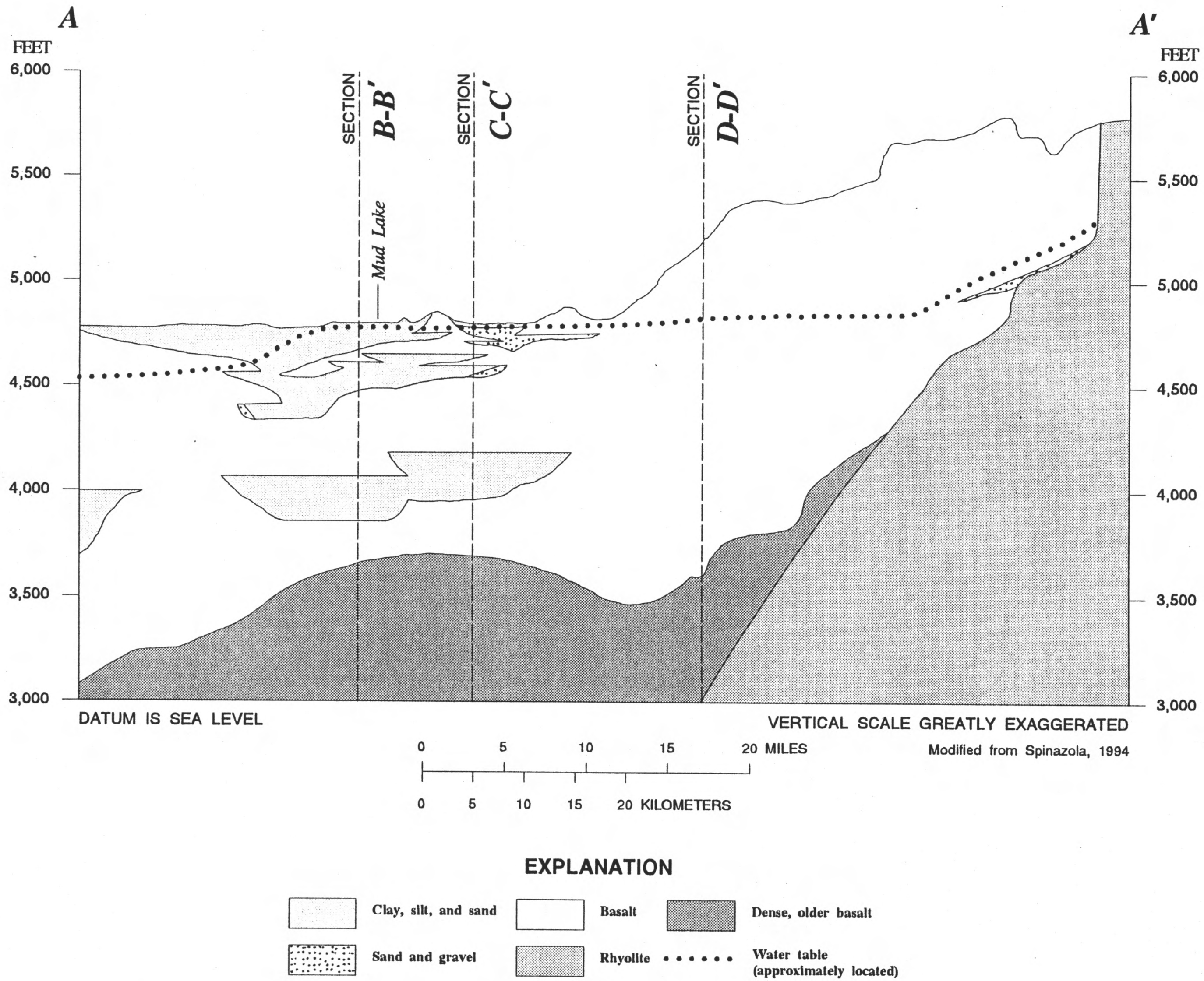


Figure 7. Generalized geologic sections *A–A′*, *B–B′*, *C–C′*, and *D–D′*.

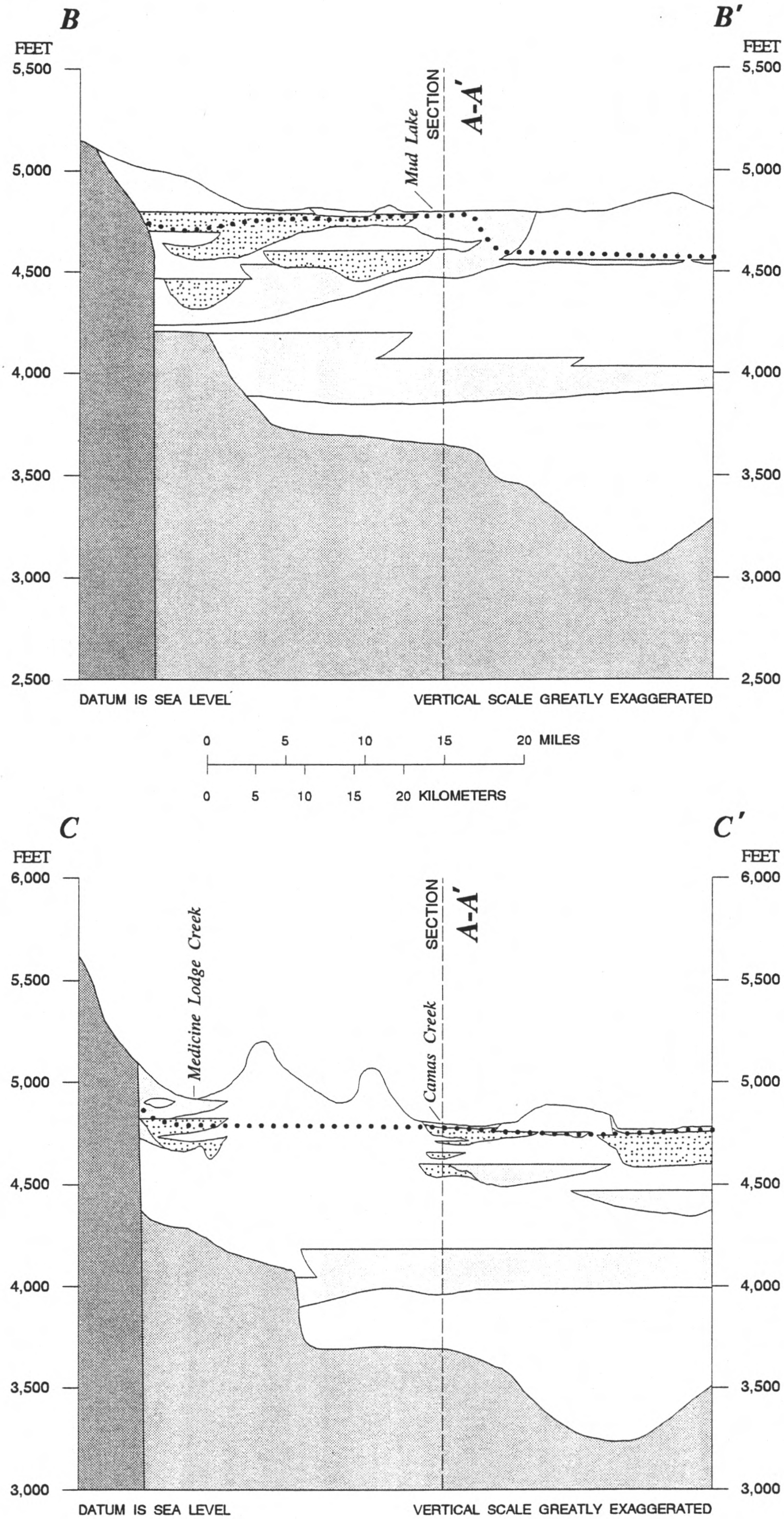


Figure 7. Generalized geologic sections *A–A′*, *B–B′*, *C–C′*, and *D–D′*—Continued.

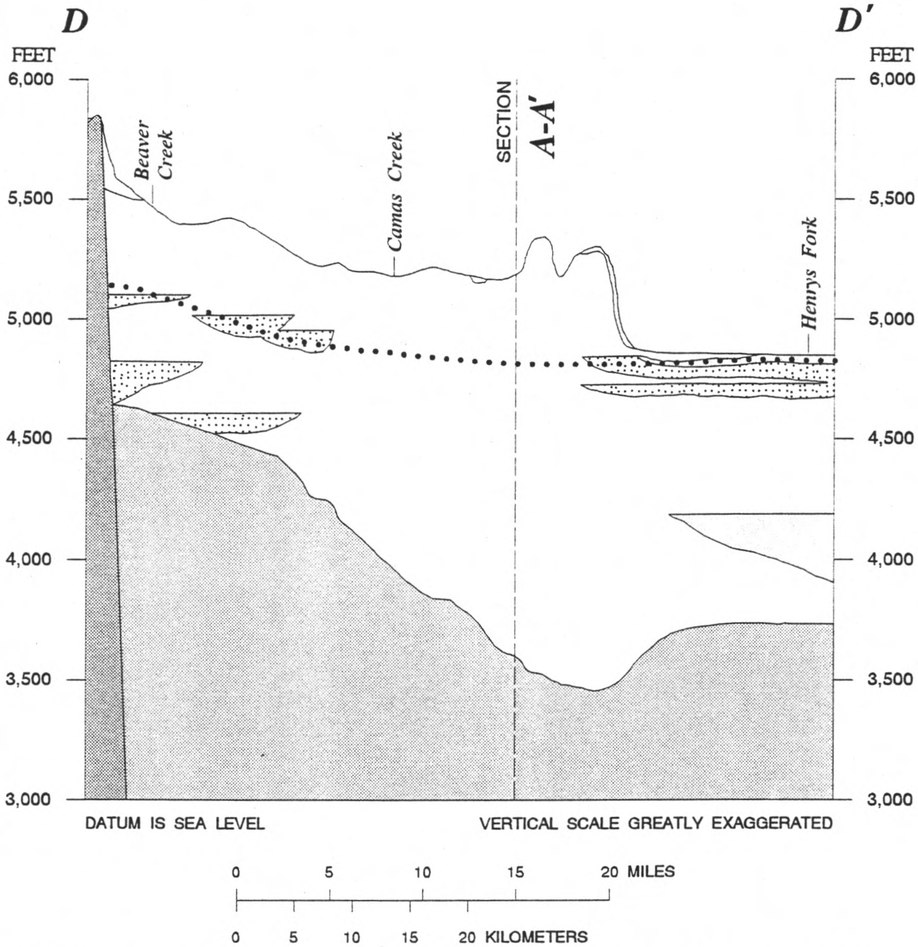


Figure 7. Generalized geologic sections *A–A'*, *B–B'*, *C–C'*, and *D–D'*—Continued.

Layers of basalt flows predominate on and under the plain (fig. 7). Individual basalt flows average 20 to 25 ft in thickness and cover areas of as much as 100 mi^2. Rubble and clinker zones are present at the top of most flows. The remainder of a flow may consist entirely of dense basalt; but vesicles, formed as gases escaped while the molten lava cooled, usually are present. The volume and distribution of vesicles often are greatest near the top of an individual flow. The interconnection of vesicles within individual and among adjacent flows is highly variable. Although basalt flows that underlie the plain are highly fractured, surficial sediments preclude mapping fault traces on much of the plain. Traces of several faults were mapped in the Kilgore area (Whitehead, 1986, sheet 1). Total thickness of basalt is less than 4,000 ft (Whitehead, 1986, sheet 2).

Sediment layers that consist mainly of sand and gravel underlie channels of the Henrys Fork and Snake River and are present in alluvial fans that extend southward from the northwestern margin of the plain (fig. 6a). Sediment layers that consist mainly of clay, silt, and sand are present in lakebeds that underlie the area around Mud Lake. Sediments are not a predominate rock type at depths greater than 1,000 ft below the water table (fig. 6e). Total thickness of sediments in the study area is 0 to less than 1,000 ft (Whitehead, 1986, sheet 2).

Basalt interbedded with sediments is most prevalent around Mud Lake (fig. 7) and progressively decreases from southwest to northeast (fig.7, section *A–A'*). Sediment interbeds affect local ground-water movement and supply. These effects are discussed in the section "Occurrence and movement."

The aquifer system described in this report is composed of saturated volcanic rocks and sediments. The top of the aquifer system is the water table. Several feet to several hundred feet of unsaturated volcanic rocks and sediments separate land surface from the water table (fig. 7). Minimum aquifer thickness is about 500 ft (fig. 6a,b,c). At depths greater than 500 ft below the water table, the aquifer is present where basalt thickness is greater than zero (fig. 6d,e). The effective base of the aquifer system is dense, older

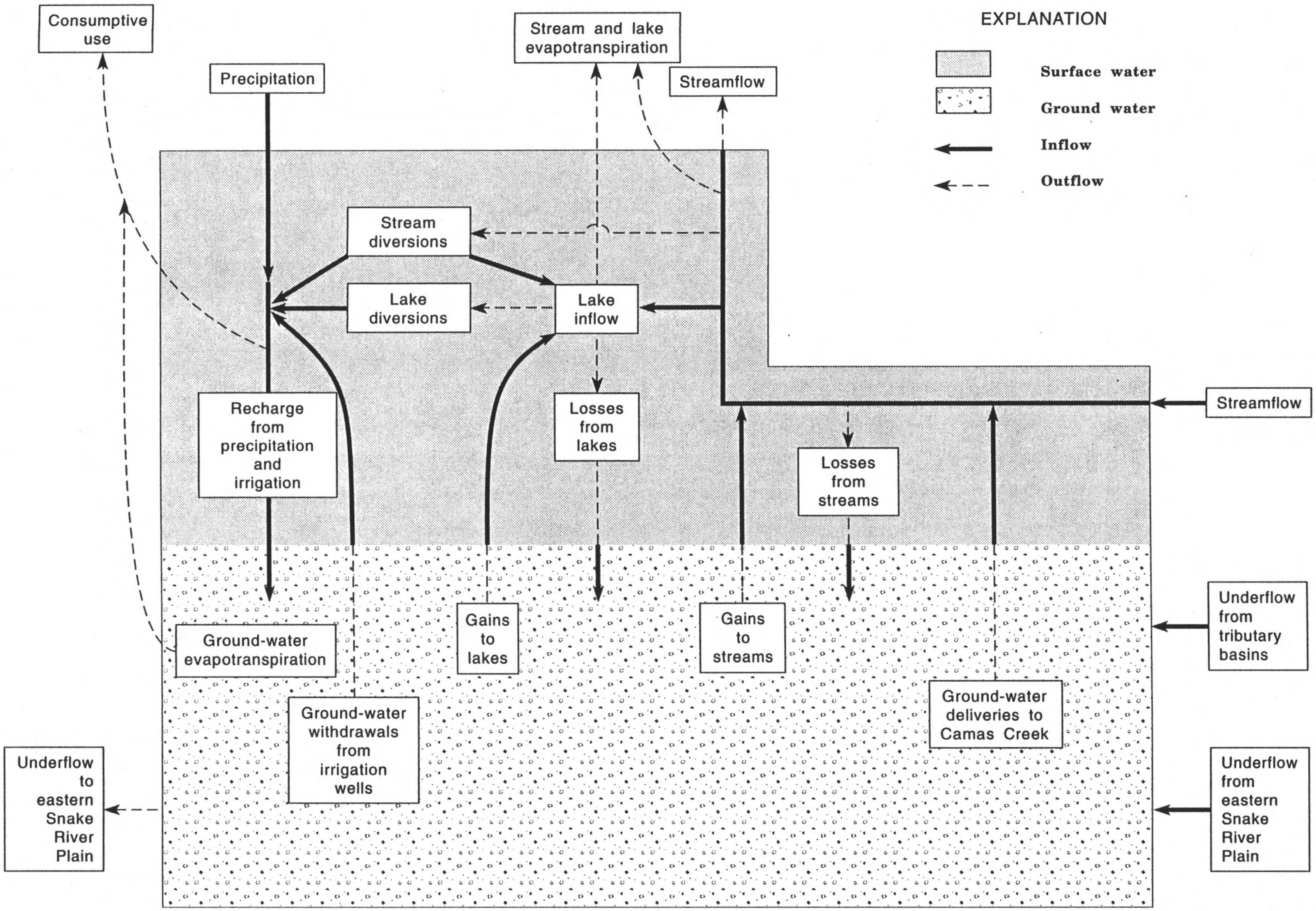


Figure 8. Relations between surface water and ground water in the Mud Lake study area.

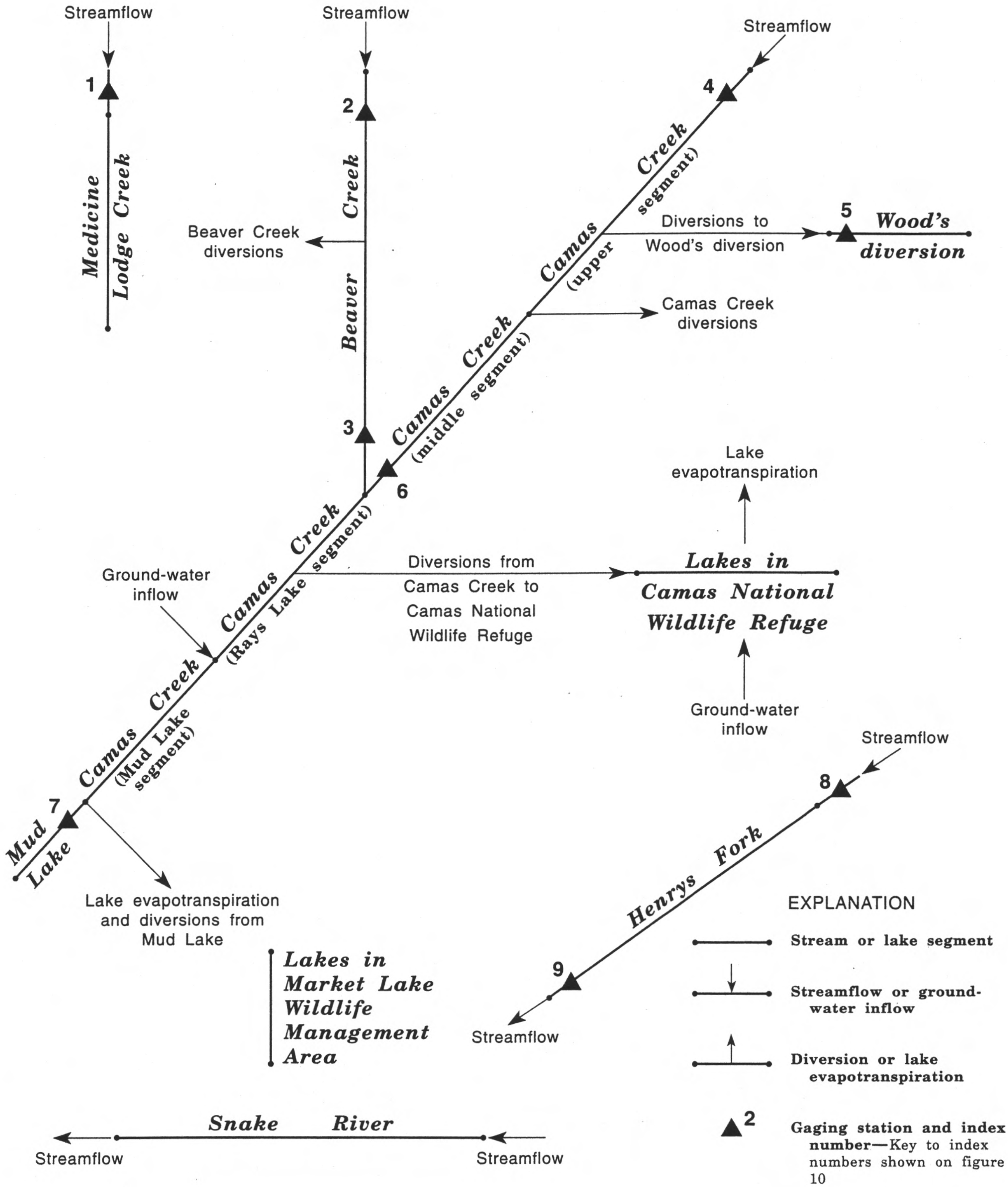


Figure 9. Stream and lake segments, streamflow, ground-water inflow, diversions, and lake evapotranspiration.

basalt or rhyolite that underlies all other rocks at depth (Whitehead, 1986, sheet 2). Maximum thickness of the aquifer system is about 2,000 ft (fig. 6).

Hydrologic Setting

Surface water and ground water enters and leaves the study area in several different forms (fig. 8). Surface water enters as streamflow at the study-area boundary. Ground water enters as recharge from precipitation and irrigation, underflow from tributary basins, and underflow from the eastern Snake River Plain aquifer system adjacent to the study-area boundary. Surface water leaves the study area as streamflow at the study-area boundary and by evaporation from streams and lakes. Ground water leaves the study area as underflow to the eastern Snake River Plain aquifer system adjacent to the study-area boundary. Both surface water and ground water leave the study area as consumptive use by plants. Consumptive use includes evaporation and transpiration associated with the growth of native vegetation and crops.

Almost all surface- and ground-water use in the study area is for irrigation or for maintaining lakes in wildlife refuges. Water is diverted from several streams and from Mud Lake to irrigate crops and from Camas Creek to maintain lakes in the Camas National Wildlife Refuge. Most ground-water use for irrigation is near points of withdrawal. Some ground water flows naturally, or is pumped, into Camas Creek or into canals. Water that flows into Camas Creek augments natural streamflow that is stored in Mud Lake for irrigation. Canals transfer water as much as several miles from the wellheads. Water transferred by canals is used to irrigate crops or to fill lakes on Camas National Wildlife Refuge or Mud Lake WMA. Lakes in Market Lake WMA are supplied entirely by natural inflow of ground water.

SURFACE WATER

Streamflow, ground-water inflow, streamflow and lake diversions, and lake evapotranspiration (ET) are major elements that affect the amount of water that flows in streams and is stored by lakes in the study area (fig. 9). Measurements or estimates of these elements were compiled for streams and lakes for the 1980–90 period and are discussed in the following sections.

Streamflow

Streamflow at or near the study-area boundary was measured regularly or intermittently at gaging stations on Medicine Lodge Creek, Beaver Creek, Camas Creek, Henrys Fork (fig. 10), and Snake River (fig. 1). Water in Birch, Warm Springs, and Deep Creeks usually percolates into the subsurface or is diverted for irrigation upstream from the study-area boundary. Flow data for Medicine Lodge Creek, Beaver Creek, Camas Creek, Henrys Fork, and the Snake River (figs.11 and 12) were obtained from measurements on file at the USGS office in Boise, Idaho, or were estimated. When streamflow gaging measurements were missing, estimates were made and data were restored from regression relations between mean daily streamflow measured at a nearby station with continuous record and mean daily streamflow at the station with missing record. When data at a nearby station were unavailable, less accurate methods were employed to estimate streamflow. Therefore, the accuracy of many streamflow estimates is unknown. Downstream from the gaging station, all water in Medicine Lodge Creek percolates into the subsurface. Water that remains after diversions are made from Beaver and Camas Creeks enters Mud Lake. Henrys Fork enters the study area near Ashton and leaves near Rexburg; the Snake River enters at Lorenzo and leaves near Lewisville.

Ground-water inflow

Water for irrigation from flowing and pumped wells enters Mud Lake by way of Camas Creek; water from pumped wells enters lakes on the Mud Lake WMA and Camas National Wildlife Refuge by way of canals. Ground-water inflow to Mud Lake (fig. 13) was obtained from records on file with the IDWR. Flow into lakes on the wildlife refuges was calculated from electrical power consumption records for wells described in the section "Discharge."

Streamflow and lake diversions

Diversions for irrigation are made from Beaver Creek, Camas Creek, Mud Lake, Henrys Fork, and the Falls and Teton Rivers (figs. 14 and 15). Diversions from Camas Creek are made at Wood's diversion (fig. 9) to reduce high flows in Camas Creek and decrease the threat of floods downstream. All flow to

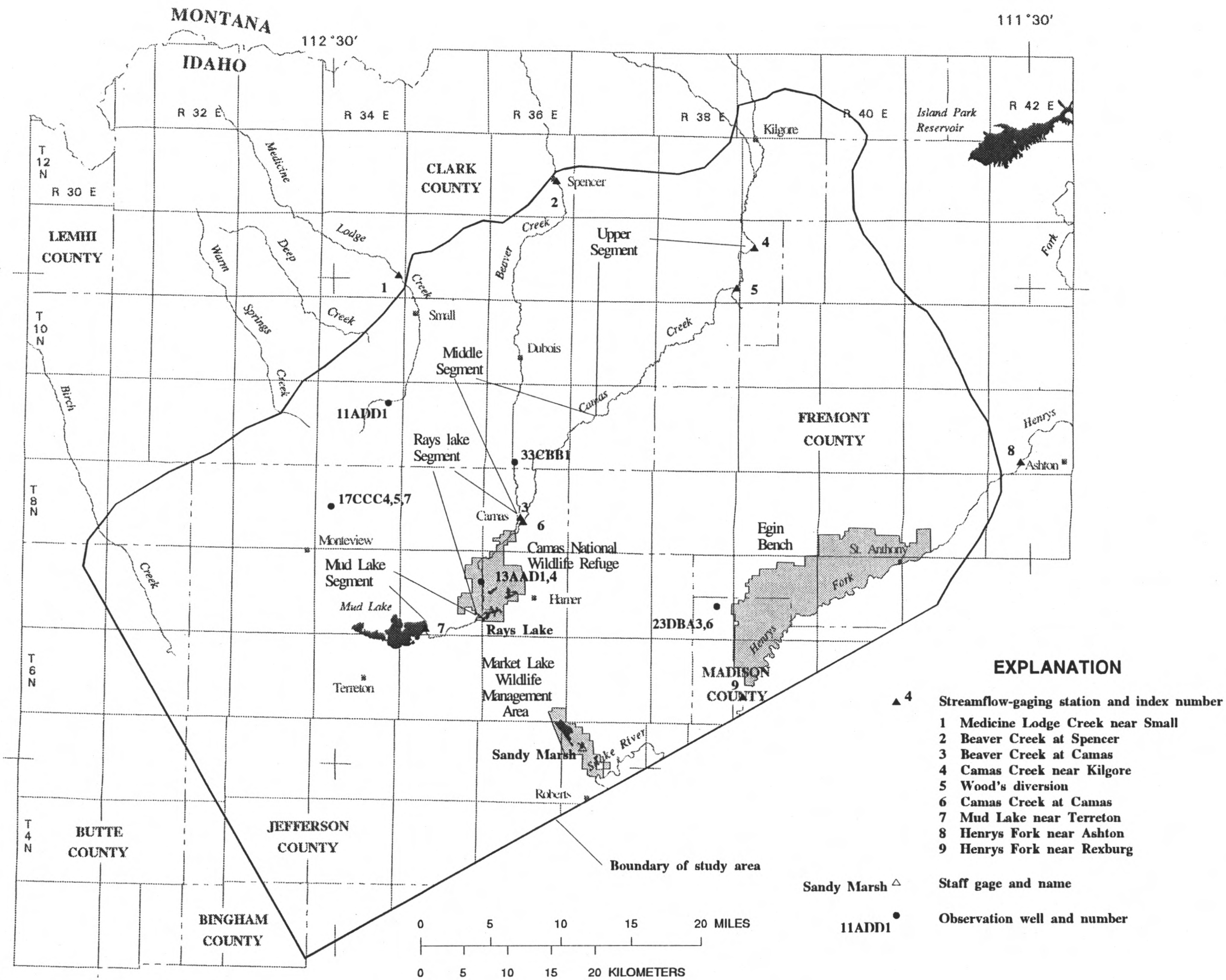


Figure 10. Locations of streamflow-gaging stations, selected stream segments, staff gages, and observation wells.

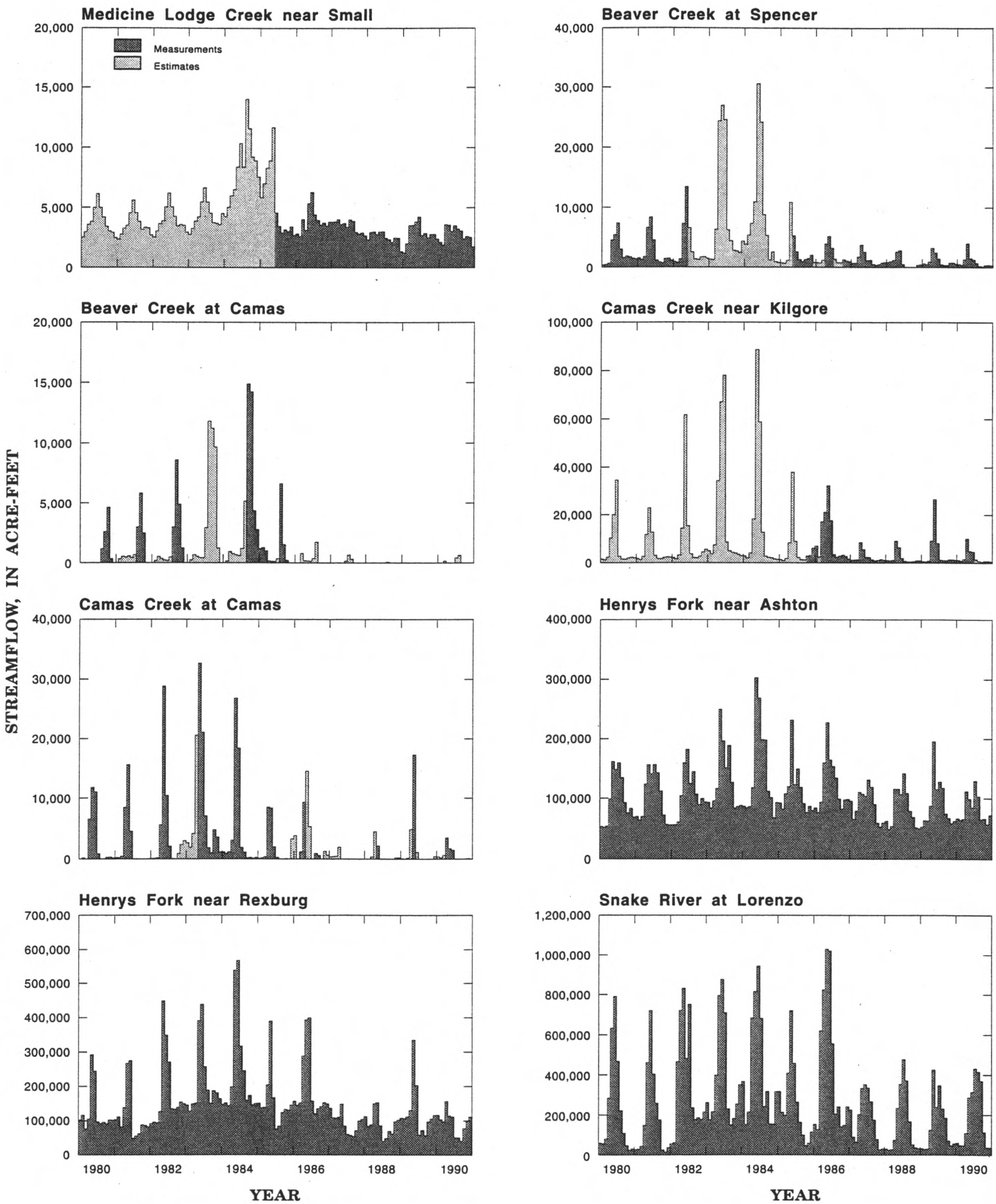


Figure 11. Monthly streamflow at selected gaging stations, 1980–90. (Station locations shown on figures 1 and 10)

Wood's diversion percolates into the subsurface. Diversions from Camas Creek and Warm Creek fill lakes on the Camas National Wildlife Refuge. Warm Creek is an ephemeral stream, distributary from Camas Creek, located mostly in the eastern part of the Camas National Wildlife Refuge (fig. 1), and terminates at a lake on the refuge. Diversions are made from Mud Lake to surrounding areas during the irrigation season or to playas on the INEL, west of the study area, during flood conditions in Mud Lake and Camas Creek. Records on file with the IDWR separate diversions from Mud Lake by source—surface-water sources are natural streamflow from Camas Creek; ground-water sources are from ground-water inflow. Elimination of irrigation diversions from Mud Lake that are attributed to ground-water sources was one of several water-use alternatives examined in a companion report (Spinazola, 1994).

Lake evapotranspiration

Water enters the atmosphere by evaporation from the surface of streams, Mud Lake, and lakes on wildlife refuges and by transpiration from phreatophytes in and adjacent to streams and lakes. ET is the combined effect of evaporation and transpiration. ET from streams was not determined but was assumed to be small in relation to the magnitude of measurements and estimates of streamflow. Lake ET volumes (fig. 16) and rates for Mud Lake were obtained from a report by Brockway and Robison (1988, appendix B, p. 12–13) and from records on file with the IDWR. ET rates for Mud Lake were applied to lake areas to determine lake ET volumes for the Camas National Wildlife Refuge and Market Lake WMA (fig. 16).

Table 1. Measured streamflow losses from the middle segment of Camas Creek

[All measured losses were calculated from streamflow measurements made by the watermaster of Water District No. 31 (Donald Shenton, written commun., 1990), except for measured loss on 5-1-90, which was calculated from streamflow measurements made by U.S. Geological Survey personnel]

Measurement date	Measured loss (acre-feet per month)	Measurement date	Measured loss (acre-feet per month)
8-25-62	2,230	6-26-89	2,300
7-16-63	2,090	5- 1-90	1,320
7- 6-66	1,450	6- 9-90	2,200
6- 8-88	1,360		

Losses and gains

Losses from and gains to streams and lakes are a measurement of the relation between surface water and ground water. A stream or lake loses water to the aquifer system through a hydraulic connection when stage in the stream or lake is greater than the ground-water level. Conversely, a stream or lake gains water from the aquifer when the ground-water level is greater than the stage in the stream or lake. Medicine Lodge, Beaver, and Camas Creeks, Mud Lake, Henrys Fork, Snake River, and lakes on wildlife refuges are hydraulically connected to the aquifer system and, therefore, can lose water to or gain water from ground water. As noted previously, all flow in Medicine Lodge Creek and Wood's diversion is lost to ground water in the study area. Losses and gains for Beaver Creek and Camas Creek (figs. 17 and 18) were calculated by subtraction of measured or estimated flow at downstream gaging stations (Beaver Creek at Camas and Camas Creek at Camas) from measured or estimated flow at upstream gaging stations (Beaver Creek at Spencer and Camas Creek near Kilgore). Streamflow diversions between the downstream and upstream stations were added to the difference for each stream reach.

Losses and gains between gaging stations at Camas Creek near Kilgore and Camas Creek at Camas were subdivided into upper and middle segments (figs. 10, 17, and 18). Observations in 1989 and 1990 verified by conversations with local farmers and ranchers indicated that flow in the upper segment of Camas Creek was perennial and flow in the middle segment was intermittent. Flows at the upstream and downstream gaging stations were used to apportion streamflow losses and gains to upper and middle segments of Camas Creek as follows:

- When no monthly flow was recorded at the downstream gaging station, losses or gains were apportioned entirely to the upper segment.
- When monthly flow was recorded at the downstream gaging station, losses or gains were apportioned to each segment as a percentage of length of each segment divided by the total length of both segments.

Losses or gains for the upper segment of Camas Creek were assumed to be representative of conditions

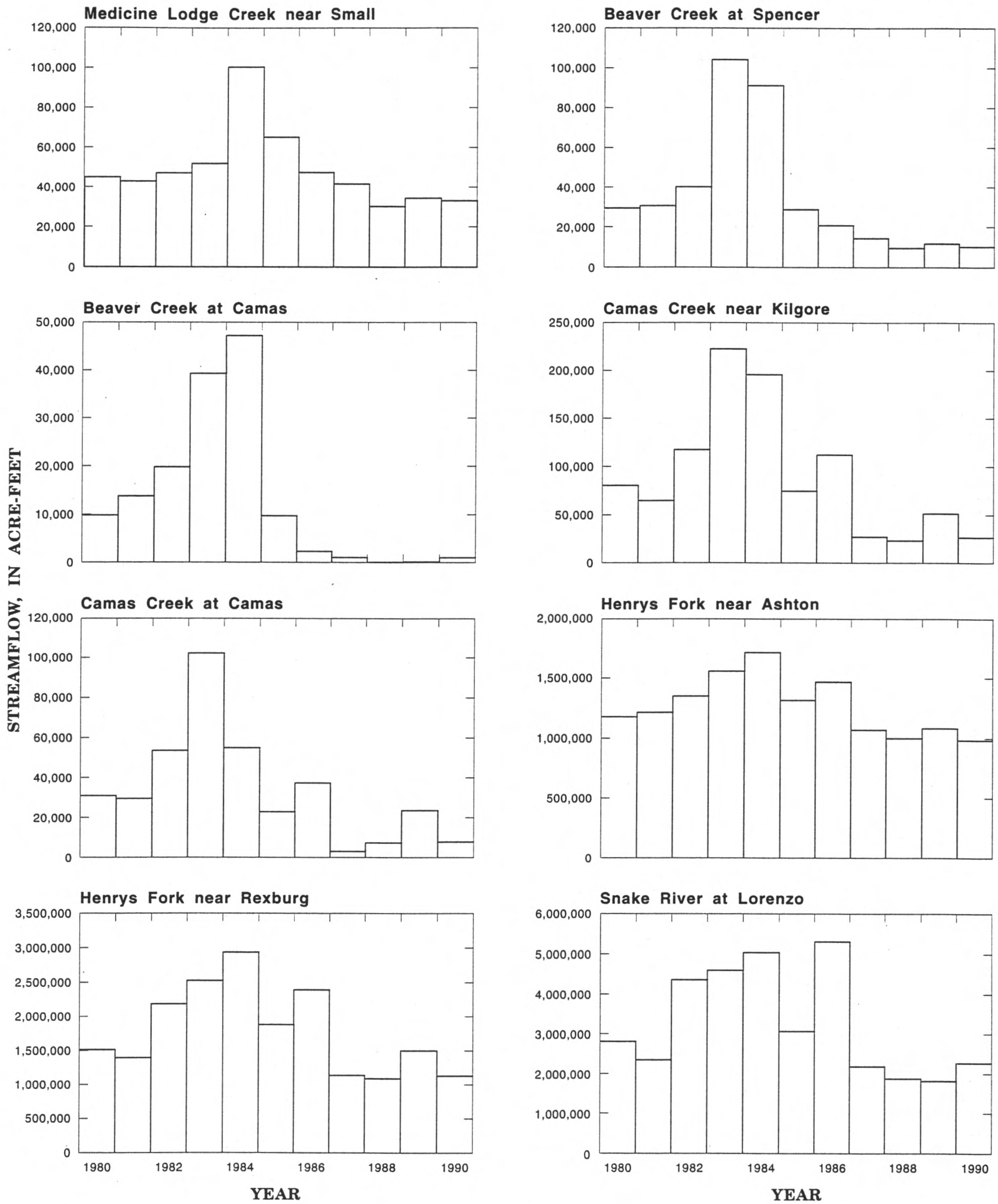


Figure 12. Annual streamflow at selected gaging stations, 1980–90. (Station locations shown on figures 1 and 10)

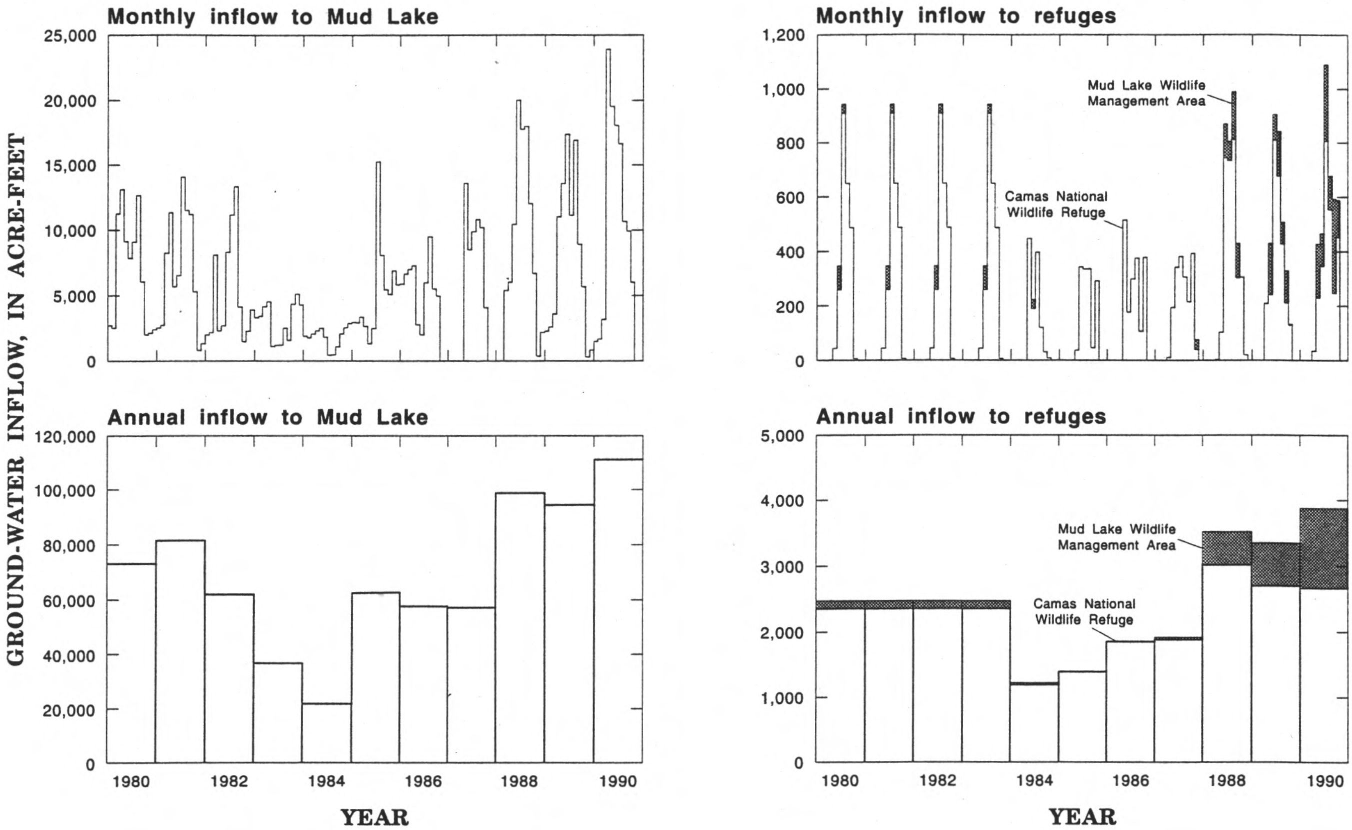


Figure 13. Ground-water inflow to Mud Lake, Mud Lake Wildlife Management Area, and Camas National Wildlife Refuge, 1980–90.

between the boundary of the study area and the middle segment of Camas Creek.

Several miscellaneous measurements conducted specifically to measure streamflow losses or gains show that the middle segment of Camas Creek consistently lost to ground water over a span of 28 years (table 1). The measured losses were converted from instantaneous to monthly rates for comparison with the losses shown in figure 17.

Miscellaneous streamflow measurements and the relation between stream stage and ground-water levels were evaluated to determine whether declines in ground-water levels adjacent to the middle segment of Camas Creek could affect streamflow losses in this segment. Relations among stream stage, ground-water levels, and streamflow losses can be determined by using the equation (McDonald and Harbaugh, 1988, p. 6–5):

$$qriv = (k * l * w / m) * (st - gw\ level), \tag{1}$$

where

- $qriv$ = streamflow loss or gain, in acre-feet per month;
- k = hydraulic conductivity of the streambed material, in feet per month;
- l = length of the stream segment, in feet;
- w = width of the stream segment, in feet;
- m = thickness of the streambed material, in feet;
- st = altitude of the stream surface, in feet; and
- $gw\ level$ = altitude of the ground-water surface, in feet.

Streamflow losses become independent from the altitude of the ground-water surface when the ground-water surface is far enough below the stream that only a narrow saturated connection exists between the streambed and the ground-water surface (McDonald and Harbaugh, 1988, p. 6–8, 6–11). Water levels in

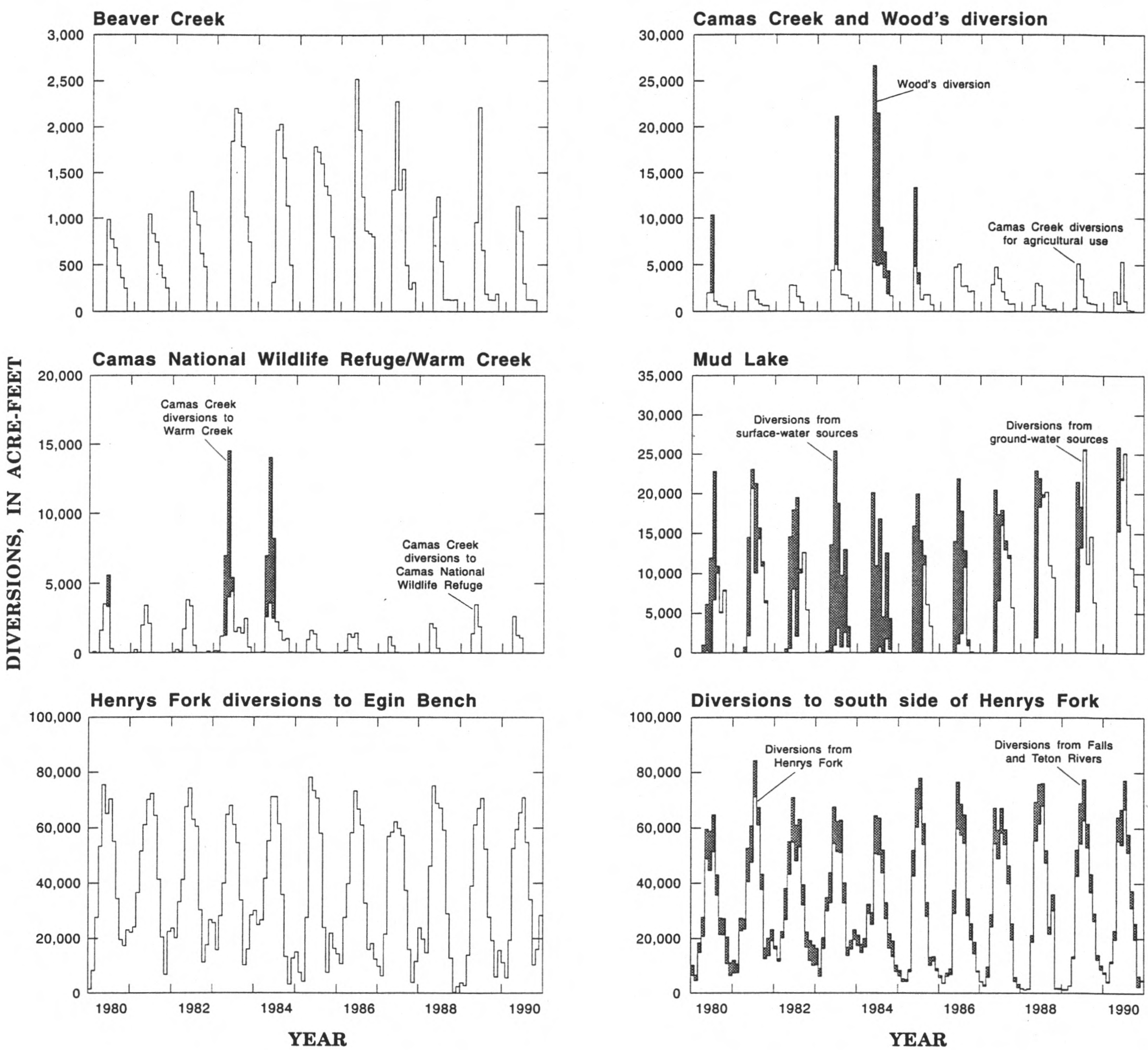


Figure 14. Monthly diversions, 1980–90.

long-term observation well 9N-36E-33CBB1 (fig. 19) are more than 70 ft below land surface. This well is less than 2 mi from Camas Creek (fig. 10); land surface at the well is near stream stage in Camas Creek. The large depth to water indicates that if there is a saturated connection between the stream and the aquifer along the middle segment of Camas Creek, the connection is narrow. The history of measured losses, the narrow saturated connection, and the large depth to ground water indicate that stream stage may be the only factor that controls stream losses in the middle segment of Camas Creek.

Losses and gains for the Rays Lake segment of Camas Creek, the Mud Lake segment of Camas Creek, and Mud Lake (figs. 17 and 18) were calculated with a water balance model (Brockway and Robison, 1988).

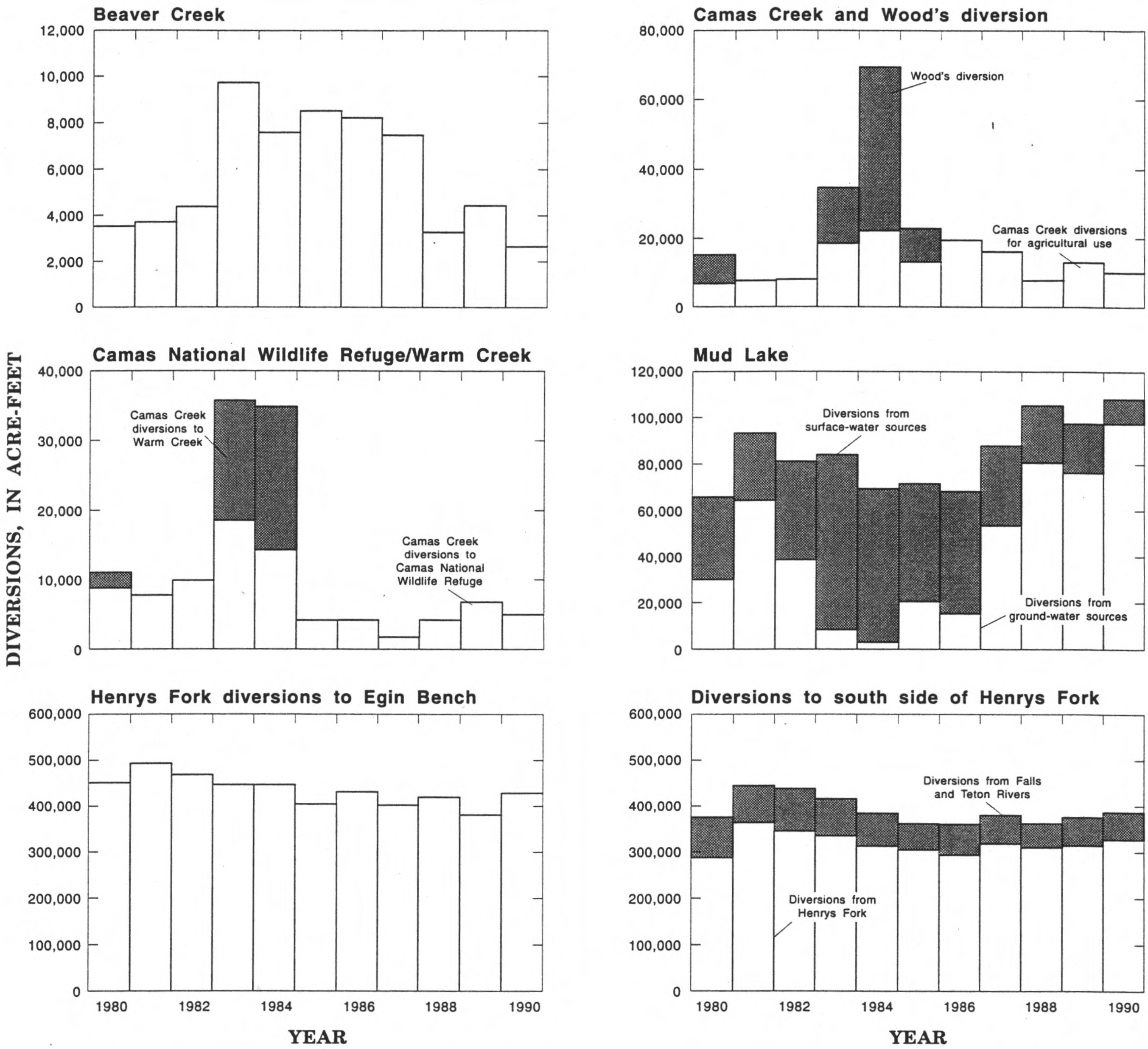


Figure 15. Annual diversions, 1980–90.

The model required monthly streamflow data (fig. 11), ground-water inflow to Mud Lake (fig. 13), stage and volume of Mud Lake, and diversions (fig. 14) and lake ET (fig. 16) from Mud Lake to calculate a water balance for that part of the basin downstream from gaging stations Beaver Creek at Camas and Camas Creek at Camas. Surface-water losses and gains were a residual in the water balance. The data set from Brockway and Robison (1988, appendix B, p. 1–6) was modified with records on file with the IDWR to extend the data through December 1990. Streamflow data used in the model for gaging stations

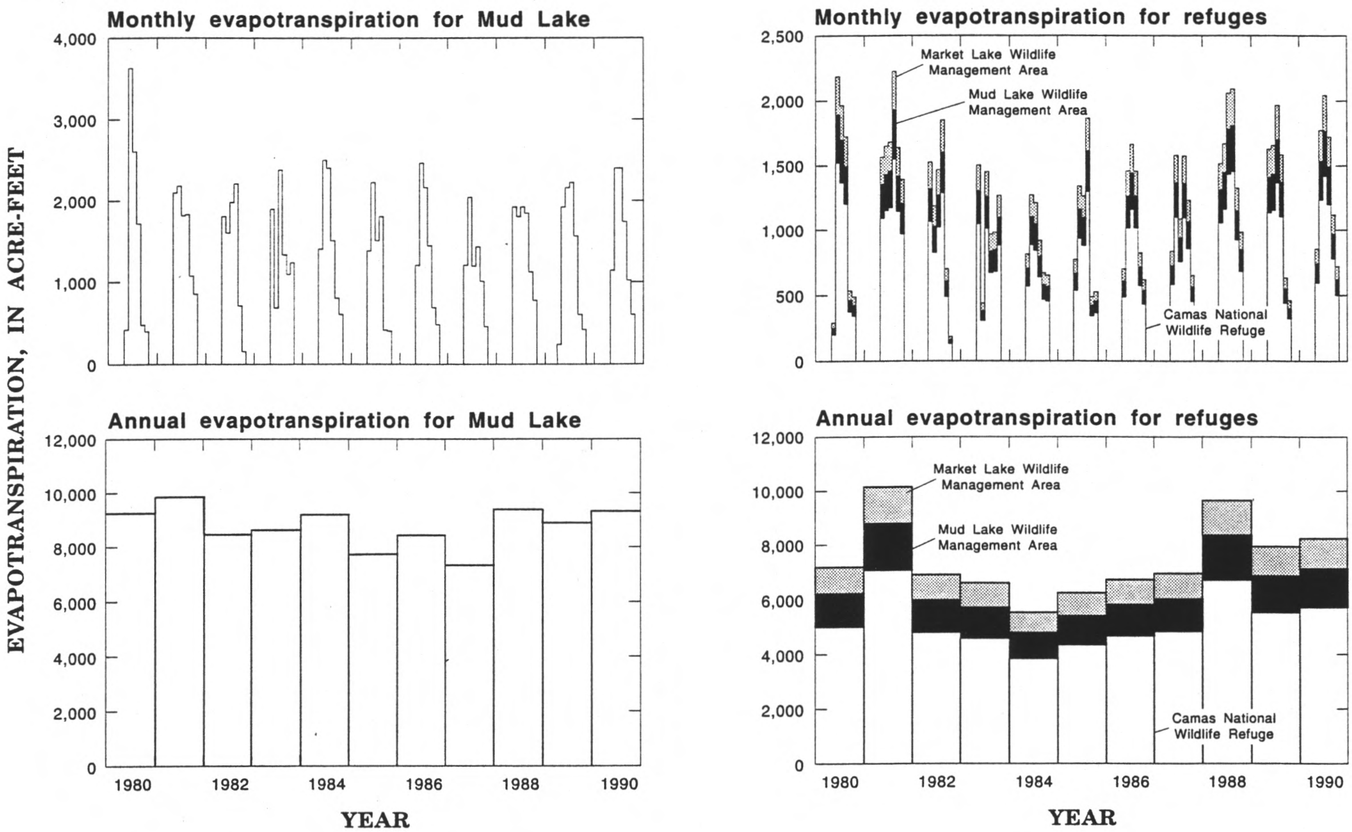


Figure 16. Lake evapotranspiration, 1980–90.

Beaver Creek at Camas and Camas Creek at Camas were derived as described in preceding paragraphs. Contents of Mud Lake were calculated with the algorithm described by Brockway and Robison (1988, appendix A) from records of stage for the gaging station Mud Lake near Terreton (records on file with the USGS in Boise, Idaho).

Flow in Camas Creek increased in a downstream direction from the Rays Lake reach to Mud Lake as a result of stream and lake gains from ground water (figs. 17 and 18). Ground-water levels adjacent to the stream became progressively closer to and then exceeded land surface as the stream neared Mud Lake. Ground-water levels were several tens of feet below land surface in well 9N-36E-33CBB1 (fig. 19), near the middle segment of Camas Creek, and were sometimes within a few feet of the stage in Rays Lake in well 7N-35E-25DAC1, adjacent to the lake (fig. 20). Camas Creek flows through Rays Lake on the Camas National Wildlife Refuge.

Losses and gains for Henrys Fork (figs. 17 and 18) were calculated by subtraction of the flow at the Henrys Fork near Rexburg gaging station from the flow at the Henrys Fork near Ashton station. Tributary inflow and irrigation return flows were added to the difference; irrigation diversions from Henrys Fork (figs. 14 and 15) were subtracted. Streamflow data were obtained from records on file with the USGS. Irrigation diversion and return flow data were obtained from records on file with the IDWR.

Losses and gains for the Snake River (figs. 17 and 18) were calculated by extension of the regression between losses and gains for the Lorenzo-Lewisville reach of the Snake River and water levels in well 4N-38E-12BBB1 (Kjelstrom, in press). The Snake River at Lorenzo gaging station (fig. 1) is about 10 mi

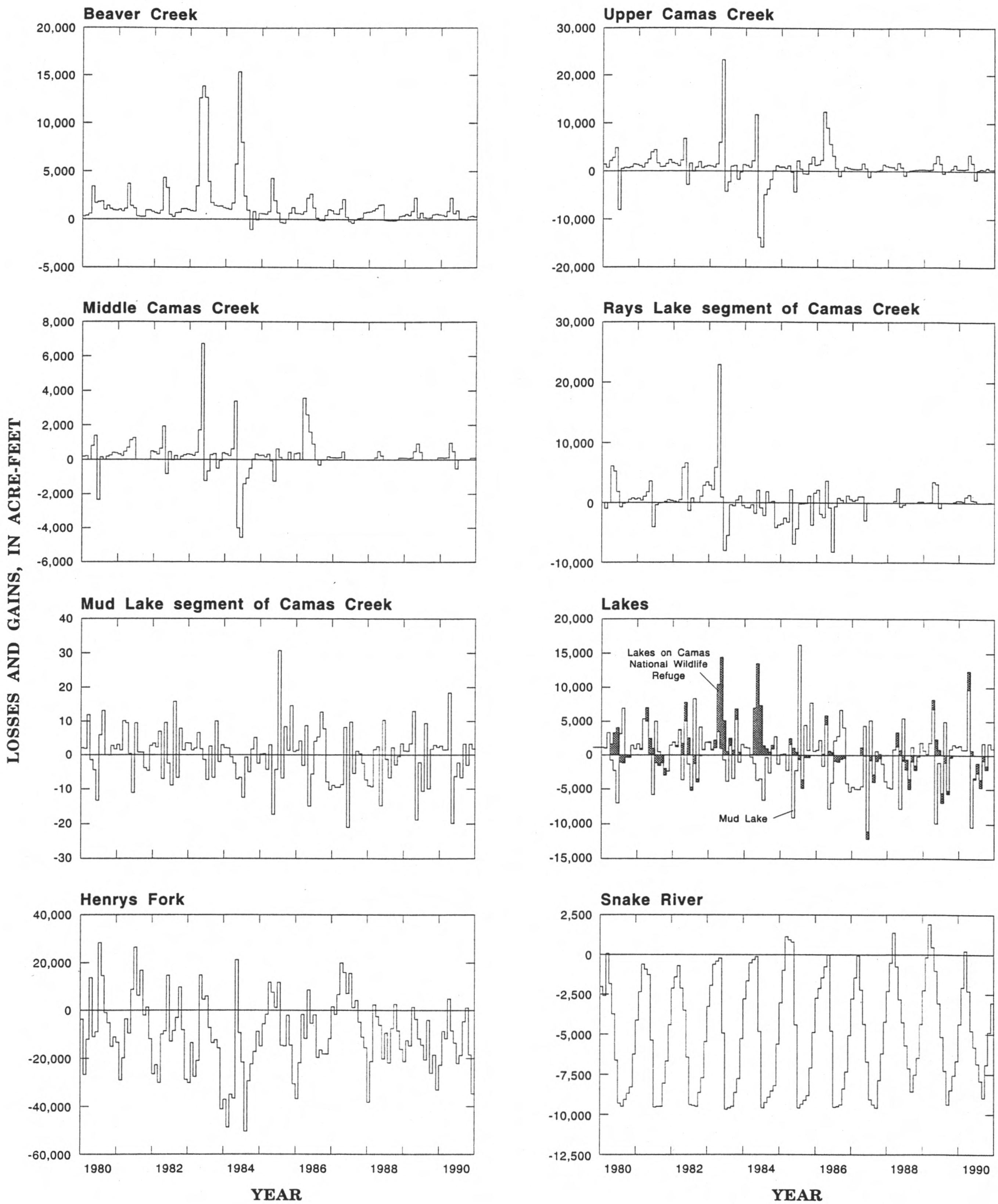


Figure 17. Monthly losses and gains (-) for streams and lakes, 1980–90.

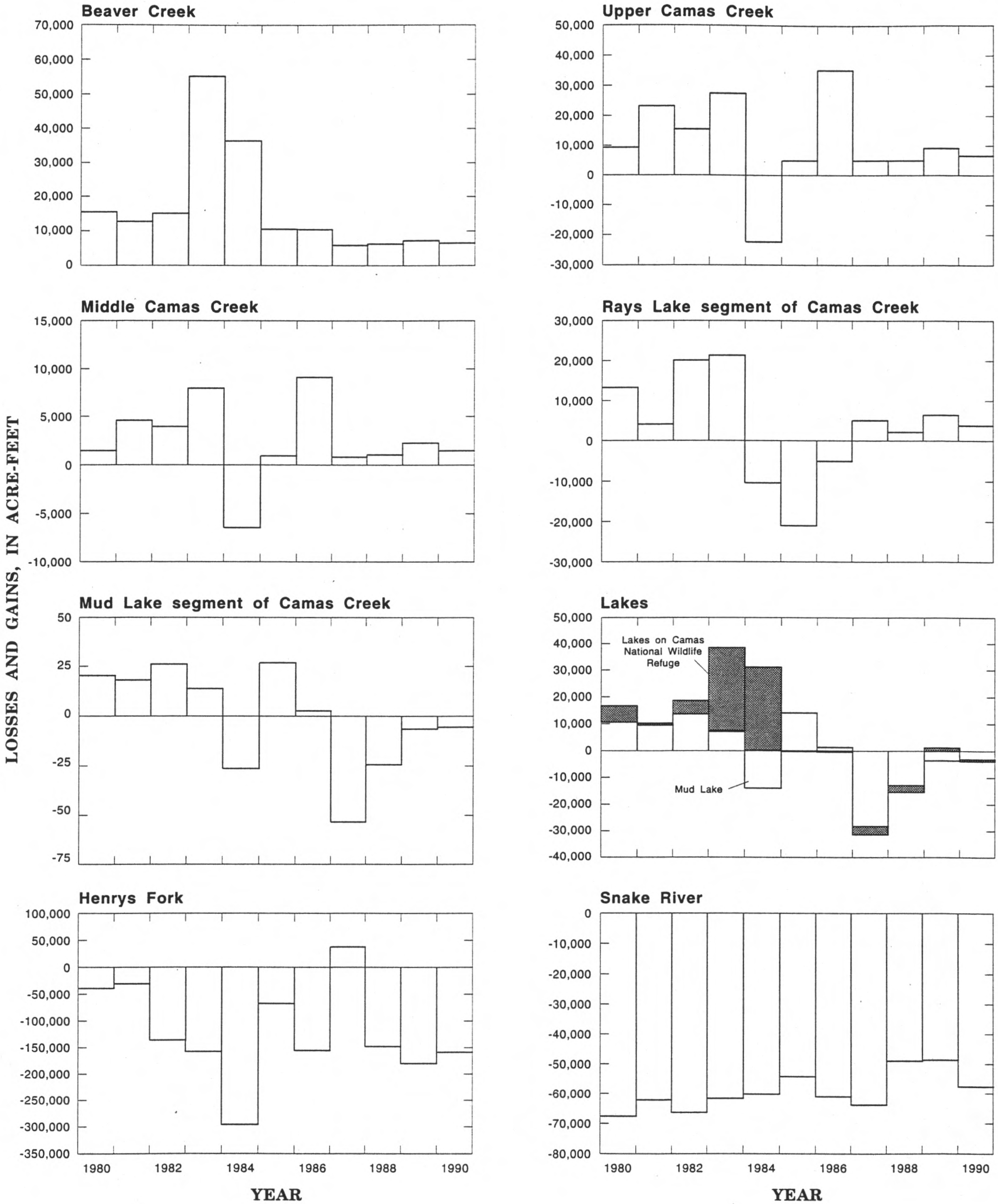


Figure 18. Annual losses and gains (-) for streams and lakes, 1980–90.

upstream from the boundary of the study area; the Snake River near Lewisville station (fig. 1) is about 6 mi downstream from the boundary. Regression results produced an r^2 value of 0.4 at the 0.001 probability level. Losses and gains obtained by regression were multiplied by 48 percent, the length of the reach of the Snake River in the study area divided by the length of the Lorenzo-Lewisville reach.

Mud Lake and a few smaller lakes are part of the Mud Lake WMA. As described previously, Mud Lake is supplied by natural flow in Camas Creek and ground-water inflow by way of Camas Creek. The smaller lakes are supplied by wells (fig. 13, "monthly and annual inflow to refuges"). Losses and gains were calculated for Mud Lake (figs. 17 and 18) but were not calculated for the smaller lakes on the WMA because ground-water inflow to Mud Lake alone is between 100 and 1,000 times greater than ground-water inflow to the smaller lakes (fig. 13).

Losses and gains for Camas National Wildlife Refuge (figs. 17 and 18) and Market Lake WMA were calculated as the difference between the sum of streamflow diversions (figs. 14 and 15) and ground-water inflow (fig. 13) to the refuges minus lake ET (fig. 16). Lakes on Market Lake WMA are supplied solely by unmeasured ground-water inflow. Therefore, gains to lakes on Market Lake WMA were considered to equal lake ET. Minor irrigation diversions were considered to have little effect on lake ET from lakes on Market Lake WMA.

Staff gages and observation wells were installed at selected locations to provide a qualitative depiction of losses and gains in support of the calculated losses and gains for Camas National Wildlife Refuge (figs. 17 and 18) and Market Lake WMA (fig. 16). Staff gages were named for the lakes in which they were installed (fig. 10). Observation wells were installed within 50 ft of the staff gages. Measurements at gages and in wells were made biweekly. Stage in Rays Lake was greater than the water level in well 7N-35E-25DAC1 in 1990 (fig. 20) and indicates that Rays Lake lost water to the aquifer. Water levels in observation well 5N-37E-8DCC1 were greater than the stage in Sandy Marsh and indicate that the lake gained water from the aquifer during most of 1990.

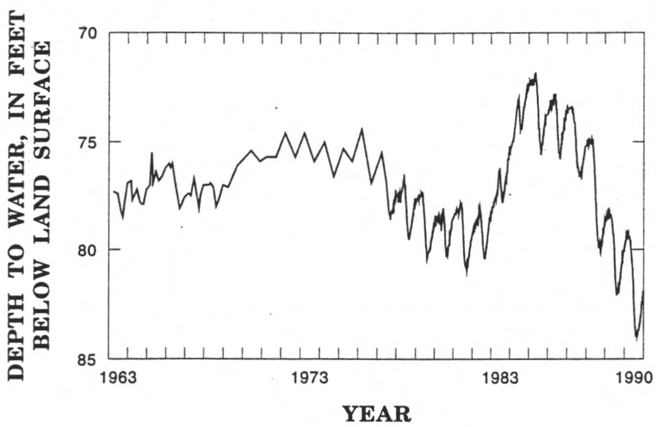


Figure 19. Depth to water in well 9N-36E-33CBB1, 1963–90. (Well location shown on figure 10)

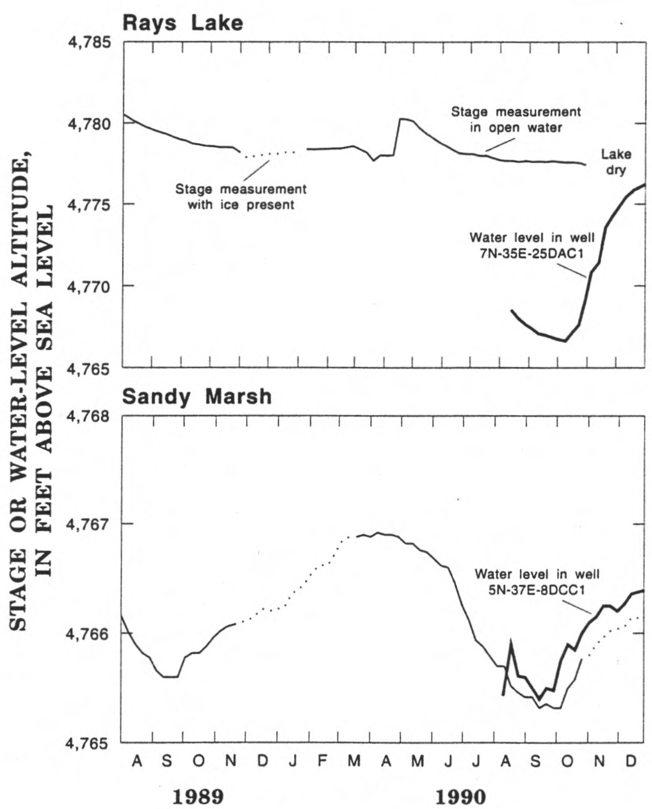


Figure 20. Stage measurements at selected sites and water levels in nearby wells, 1989–90. (Sites shown on figure 10)

GROUND WATER

Water-level measurements in wells were used to help define the occurrence and movement of ground water. Measurements and estimates described in sections that follow were used to derive values for

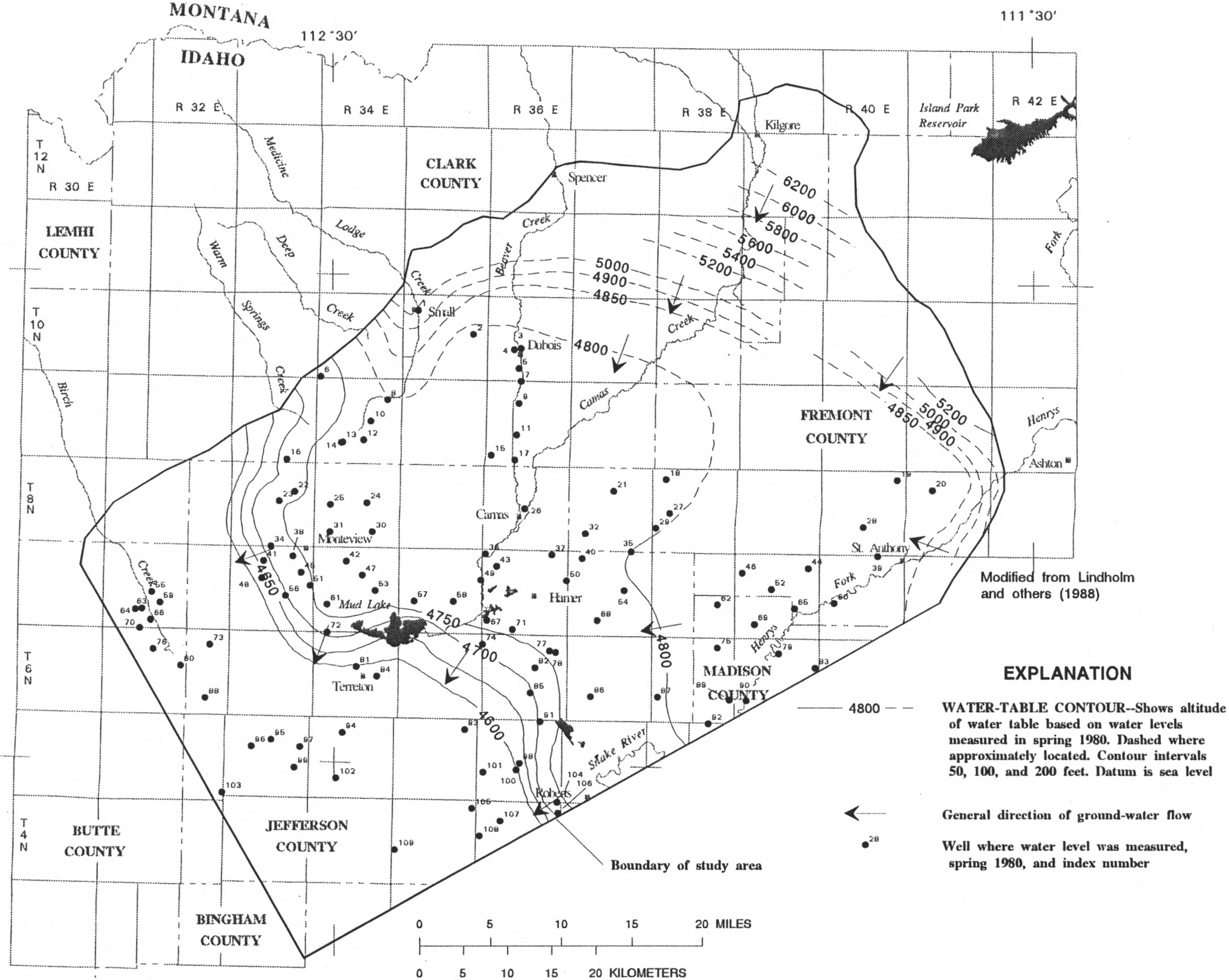


Figure 21. Water table, spring 1980.

recharge, discharge, and aquifer properties. Recharge includes infiltration of precipitation and irrigation, stream and lake losses, and underflow from basins tributary to the Snake River Plain and from the eastern Snake River Plain aquifer system adjacent to the study area. Discharge includes underflow to the eastern Snake River Plain aquifer system adjacent to the study area, stream and lake gains, ground-water ET, withdrawals from wells, and flowing wells. Aquifer properties influence the ability of the aquifer system to transmit and store water.

Occurrence and movement

Ground water is present under unconfined and confined conditions in the study area. Generally, water nearest the land surface is unconfined, and the water table defines the top of the aquifer. "The water table is that surface in an unconfined water body at which the pressure is atmospheric" (Lohman and others, 1972, p. 14). Layers of low-permeability sediments and basalt that underlie the area between Monteview and Roberts interfinger with more permeable layers to the southwest and northeast (fig. 21), retard the lateral and vertical movement of water, and provide the geologic setting for locally confined conditions. "Confined ground water is under pressure significantly greater than atmospheric" (Lohman and others, 1972, p. 7). An aquifer is confined where the hydraulic head in the aquifer is above the base of an overlying confining bed that is composed of rocks of distinctly lower permeability than those that comprise the aquifer.

Ground water generally moves perpendicular to water-table contours from areas of higher water-level altitude, as much as 6,200 ft above sea level in the northeastern part of the study area near Kilgore, to lower water-level altitude, almost 4,500 ft near the southwestern corner of the study area (fig. 21). The water-table gradient is about 30 ft/mi in the area between the 4,600- and 4,700-ft contours. That area coincides with a zone of rocks in the subsurface that is predominantly sediments and sediments interbedded with basalt (figs. 6 and 7). Low permeability in this zone results in a steeper water-table gradient. The water table closely resembles the configuration of the land surface at altitudes greater than 4,700 ft. The water-table gradient is as low as about 3 ft/mi between the 4,700- and 4,900-ft contours and steepens to about 120 ft/mi where the water-table altitude is greater than 4,900 ft.

Water-level measurements in wells located close to one another but completed at different depths in the aquifer system indicate areas of downward or upward water movement. In general, water moves downward along the margins of the plain. Water-level altitudes in wells 7N-38E-23DBA3 (open from 127 to 202 ft) and DBA6 (open from 467 to 472 ft) (fig. 22) near Egin Bench (fig. 10); and wells 8N-34E-17CCC7 (open from 40 to 50 ft), CCC4 (open from 511 to 545 ft), and CCC5 (open from 602 to 607 ft) (fig. 22) in the sediments downstream from the mouth of Medicine Lodge Creek (fig. 10) indicate downward water movement from recharge at land surface.

The zone of low-permeability sediments and basalt that underlies the area between Monteview and Roberts impedes the horizontal movement of water and contributes to conditions that produce vertical water movement. Locally, water moves downward along the southwestern margin of the zone and moves upward along the northeastern margin. Measurements made in 1968 show that water moves downward southwest of Mud Lake (Ralston and Chapman, 1969, p. 8–9). Water levels northeast of Mud Lake in wells 7N-35E-13AAD1 (open from 322 to 327 ft) and AAD4 (open from 862 to 867 ft) (fig. 22), located on Camas National Wildlife Refuge (fig. 10), indicate upward water movement. Upward movement along the northeastern margin also is indicated by ground-water discharge to streams, lakes, and flowing wells when water-level altitudes in the aquifer system are higher than land-surface altitudes. Discharge to the Rays Lake and Mud Lake segments of Camas Creek and to Mud Lake and lakes on Camas National Wildlife Refuge is indicated by stream and lake gains (figs. 17 and 18). Ground water also discharges to lakes in the Market Lake WMA. Water levels in observation well 5N-37E-8DCC1 in relation to stage in Sandy Marsh (fig. 20) in Market Lake WMA (fig. 10) indicate that the lake gains from ground water. Flowing irrigation wells (described in the "Discharge" section) indicate upward ground-water movement in T. 7 N., R. 35 E.

Recharge

Recharge to the aquifer system is from infiltration of precipitation and applied irrigation water in excess of consumptive use by plants, underflow from basins tributary to the Snake River Plain and from the eastern Snake River Plain aquifer system through part of the southeastern boundary of the study area, and stream

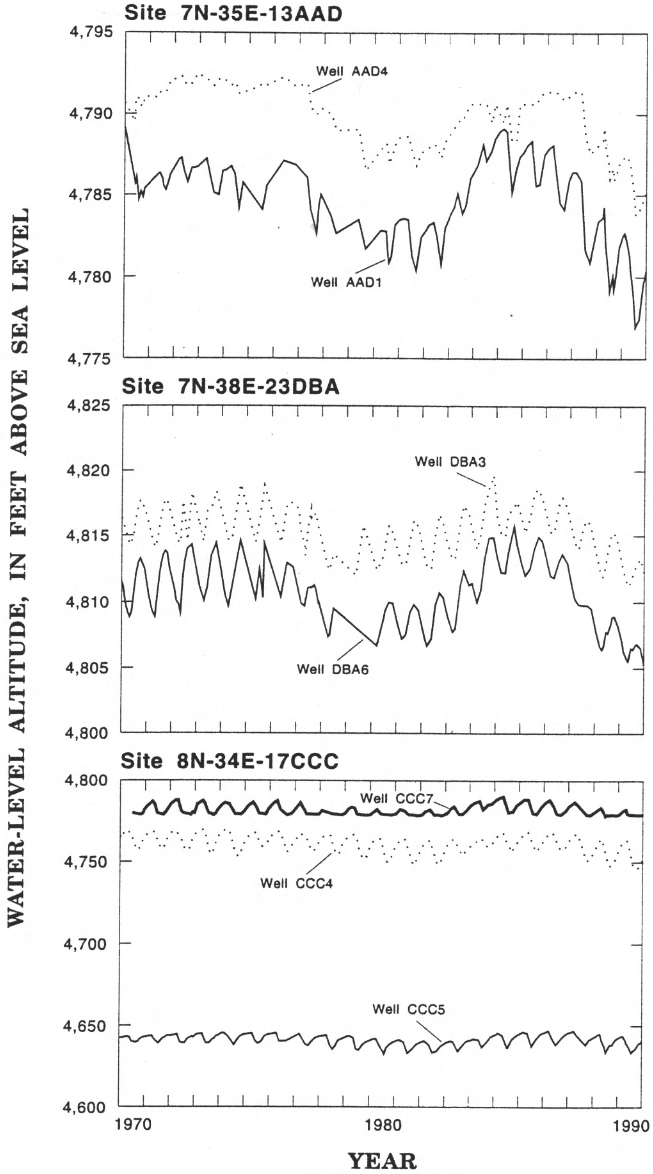


Figure 22. Water levels in wells at selected sites, 1970–90. (Water levels were measured on a continuous or less frequent basis; well locations shown on figure 10)

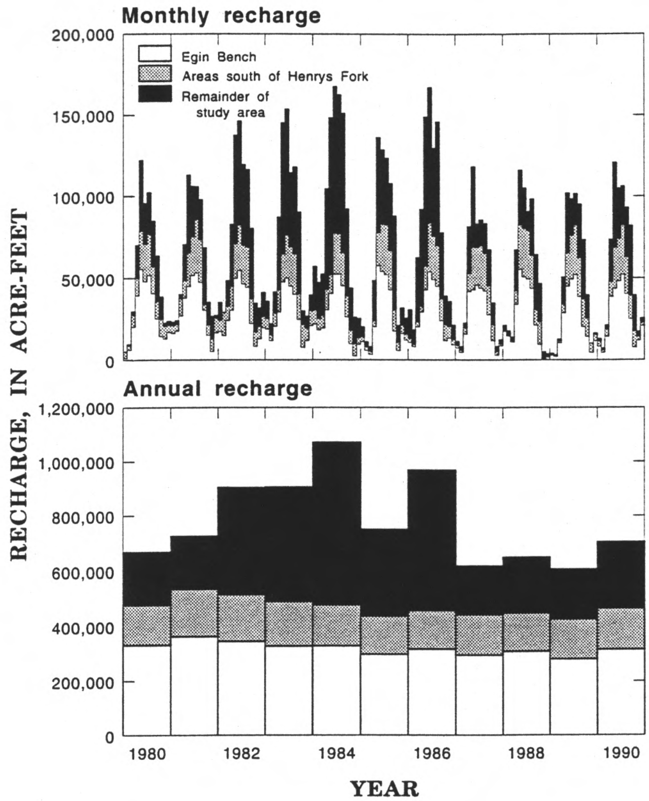
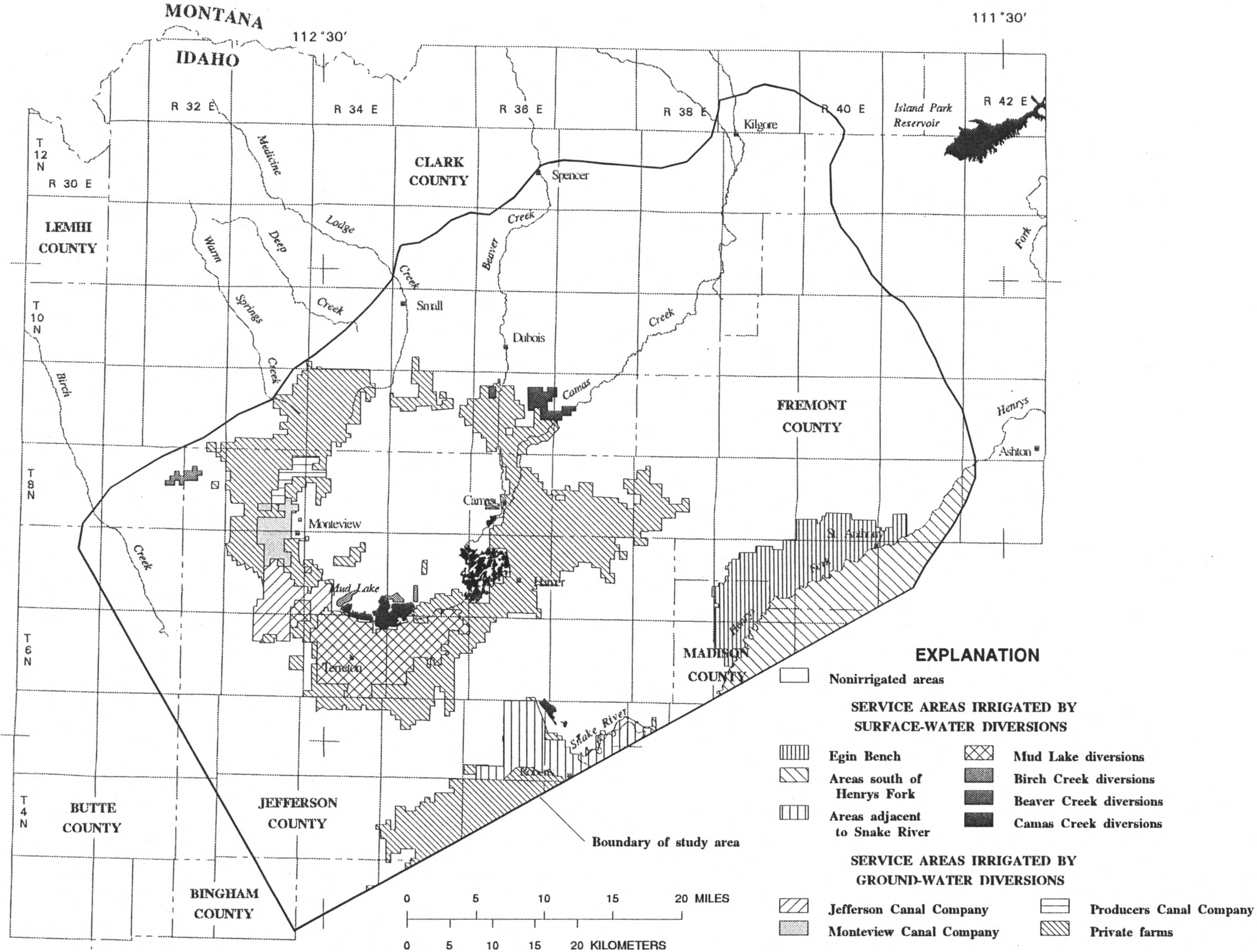


Figure 23. Recharge from precipitation and irrigation for selected parts of the study area, 1980–90.

and lake losses. Recharge from precipitation and irrigation (fig. 23) was calculated using different methods for different parts of the study area. Recharge from precipitation alone was calculated for nonirrigated areas (fig. 24). Recharge from precipitation and irrigation was calculated for irrigated areas. Recharge from precipitation, irrigation, and underflow frequently was derived from single average values used in previous studies; these values were indexed to derive monthly and annual recharge values from 1980 to 1990. Methods used to estimate stream and lake losses were described earlier in the section "Losses and Gains." Underflow from the eastern Snake River Plain aquifer system was calculated as a residual by the numerical model and is described in the section "Calibration Results."

The method used to calculate monthly recharge solely from precipitation was modified from a method used to calculate average annual recharge from precipitation by Garabedian (1992, p. 15–17). Garabedian assumed that recharge varied as a function of precipitation, soil thickness, and infiltration capacity of the soil cover, and that most recharge is from snowmelt during winter and spring when ET rates are low. Garabedian mapped generalized soil zones on the plain and assigned to each a single value of average annual precipitation and recharge rate. Garabedian's method was modified as follows to utilize the variation in precipitation within each soil zone. First, precipitation was interpolated from the mean annual precipita-



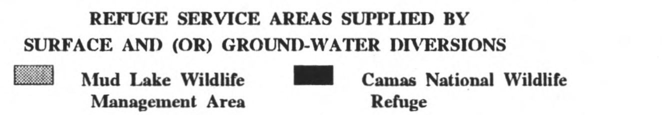


Figure 24. Irrigated areas generalized from satellite imagery, 1986.

tion map (fig. 4) at points 1 mi apart in a regular grid. Precipitation at each grid point was assumed to represent precipitation for the area between grid points. A recharge factor was calculated by division of recharge rate by average annual precipitation for each soil zone (Garabedian, 1992, p. 23). Finally, average annual recharge was calculated by multiplication of the precipitation at each point in the grid by the recharge factor for the soil zone to which it belonged.

Monthly recharge from precipitation was calculated by multiplication of average annual recharge at each point in the grid by a monthly precipitation factor. The monthly precipitation factor was calculated by multiplication of the percent of total annual precipitation that was recorded at Dubois in each month (fig. 5) by a weighting factor. The weighting factor was calculated by division of total October through March precipitation for the water year of interest by average October through March precipitation from 1930 to 1957, the period represented by the precipitation map (fig. 4). The factor weighted October-March precipitation because recharge was assumed to be greatest during the winter and spring months when ET rates are lowest (Garabedian, 1992, p. 16).

Monthly recharge from precipitation and irrigation was calculated for Egin Bench, irrigated areas south of Henrys Fork, areas irrigated with diversions from the Snake River, areas irrigated with diversions from Birch Creek, Beaver Creek, Camas Creek, and Mud Lake, and areas irrigated with ground water (fig. 24). Recharge from precipitation and ground-water inflow was calculated for smaller lakes on the Mud Lake WMA. Different methods were used to calculate recharge for each of these areas.

Monthly recharge on Egin Bench and irrigated areas south of the Henrys Fork (fig. 23) was calculated with the equation:

Recharge (acre-feet per month) =
diversions (acre-feet per month) *
[average recharge (acre-feet per year)/
average diversions (acre-feet per year)]. (2)

Monthly streamflow diversions to Egin Bench (fig. 14) were obtained from records on file with the IDWR; average recharge was 320,176 acre-ft/yr (King, 1987, p. 18) and average annual diversions were 435,000 acre-ft/yr (King, 1987, p. 16). Average recharge was calculated from diversion, return flow, evaporation, infiltration, and consumptive use data (King, 1987, p. 18). Ground-water ET provides part of the consumptive use in subirrigated areas on Egin Bench.

Monthly streamflow diversions to areas south of the Henrys Fork (fig. 14) were obtained from IDWR records. Average recharge was about 130,000 acre-ft/yr for an area of 21,800 acres using a recharge rate of 5.97 ft/yr (Garabedian, 1992, p. 15); average diversions were 337,400 acre-ft/yr (S.P. Garabedian, USGS, written commun., 1989). Recharge to areas irrigated with diversions from the Snake River was assumed to be a constant 5.5 ft/yr, the average of the recharge rate for Garabedian's area 6 of 8.2 ft/yr and for his area 7 of 2.79 ft/yr (Garabedian, 1992, p. 15). Areas 6 and 7 coincide with irrigated areas adjacent to the Snake River (Garabedian, 1992, p. 13). Average recharge for areas south of the Henrys Fork and areas irrigated with diversions from the Snake River was calculated from data for diversions, return flows, and ET (Garabedian, 1992, p. 11).

Recharge for areas irrigated with diversions from Beaver Creek, Camas Creek, Mud Lake, and areas irrigated with ground water (fig. 24) was calculated with the equation:

Recharge (acre-feet per month) =
precipitation (acre-feet per month) +
irrigation applications (acre-feet per month) -
consumptive water use (acre-feet per month). (3)

Monthly precipitation was calculated by multiplication of the grid of monthly precipitation (described in a preceding paragraph) by irrigated area attributed to the area around each grid point. Irrigation applications in areas irrigated with diversions from Camas Creek and Mud Lake (fig. 24) were obtained from records on file with the IDWR. Applications for areas irrigated with ground water were calculated from electrical power consumption records as described in the "Discharge" section. Consumptive water use for all irrigated areas was calculated as the product of alfalfa reference ET and average crop coefficients. An average monthly alfalfa reference ET (fig. 25) was calculated from values cited for the Dubois Experiment Station and the Hamer 4 NW weather stations (Allen and Brockway, 1983, p. 134). Average monthly crop coefficients (fig. 26) were derived by weighting mean crop coefficients (Allen and Brockway, 1983, p. 64) in proportion to the number of acres for each major crop harvested each year from 1980 through 1990 (Idaho Agricultural Statistics Service, annual reports).

The recharge rate calculated for nearby ground-water-irrigated areas also was used for areas irrigated with diversions from Birch Creek because Birch Creek diversion and canal-loss data were not adequate to calculate recharge. Crops raised with diversions from Birch Creek are similar to those raised in nearby areas irrigated with ground water. Irrigation applications, consumptive water use, and recharge rates in nearby areas irrigated with ground water were assumed to be similar to those in areas irrigated with diversions from Birch Creek.

Recharge from smaller lakes on the Mud Lake WMA was calculated as the difference between the sum of monthly precipitation and ground-water inflow (fig. 13) minus lake ET (fig. 16). Recharge was not allowed to be negative but was set to zero where ET exceeded ground-water inflow. Any recharge from Camas National Wildlife Refuge or Market Lake WMA was represented by losses from lakes described earlier in this report. A map that shows the distribution of recharge from precipitation and irrigation is presented in the section "Boundary Conditions."

Monthly underflow into the study area (fig. 27) was calculated from single values of underflow for each tributary drainage basin indexed by reference water levels in an observation well with the equation:

Underflow (acre-feet per month) =
0.95 (dimensionless) *
long-term underflow (acre-feet per year) *
monthly reference water level (feet per month) /
annual reference water level (feet per year). (4)

A dimensionless constant of 0.95 was obtained by a process of trial-and-error to produce annual results for 1980 nearly equal to reported values of long-term underflow for each basin (Garabedian, 1992, p. 19). Water levels in well 9N-34E-11ADD1, near the mouth of Medicine Lodge Creek (fig. 10), exhibit seasonal and annual trends similar to levels in other wells throughout the study area (figs. 19 and 22) and, therefore, were considered representative of all tributary basins. The well is located at the edge of the irrigated area near the margin of the plain, and water levels in the well represent unconfined conditions that respond more to changes in tributary underflow from Medicine Lodge Creek than to recharge from precipitation and irrigation. A single monthly reference water level near the middle of each month was selected from measurements in this well, and the sum of the monthly reference measurements was averaged for each calendar year to determine the annual reference water level.

Discharge

Discharge from the aquifer system includes underflow to the eastern Snake River Plain aquifer system across the southwestern boundary and part of the southeastern boundary of the study area, stream and lake gains, ground-water ET, withdrawals from wells for irrigation, and discharge from flowing wells. Estimates of underflow to the eastern Snake River Plain aquifer system were obtained from the model produced by Garabedian (1992) and are described in the section "Calibration Results"; stream and lake gains were described in the section "Losses and Gains." Ground-water ET occurs where plant roots intercept the water table and was assumed to be part of lake ET described earlier.

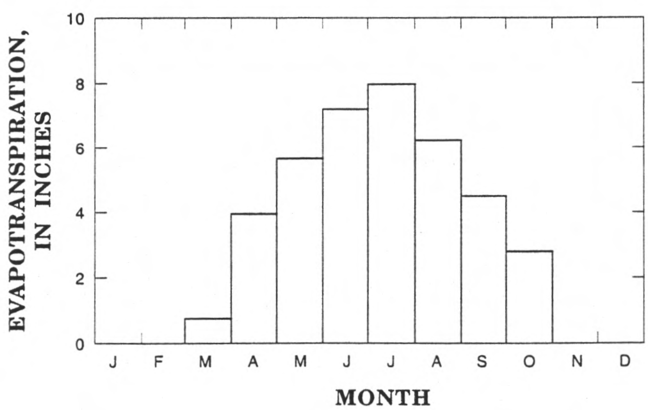


Figure 25. Average monthly alfalfa reference evapotranspiration rate, 1980–90.

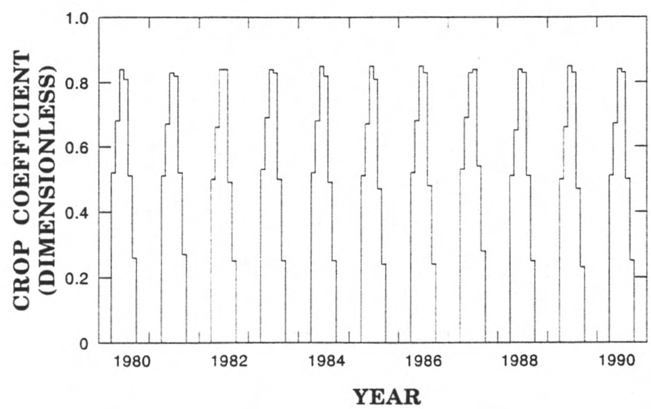


Figure 26. Average monthly crop coefficients, 1980–90.

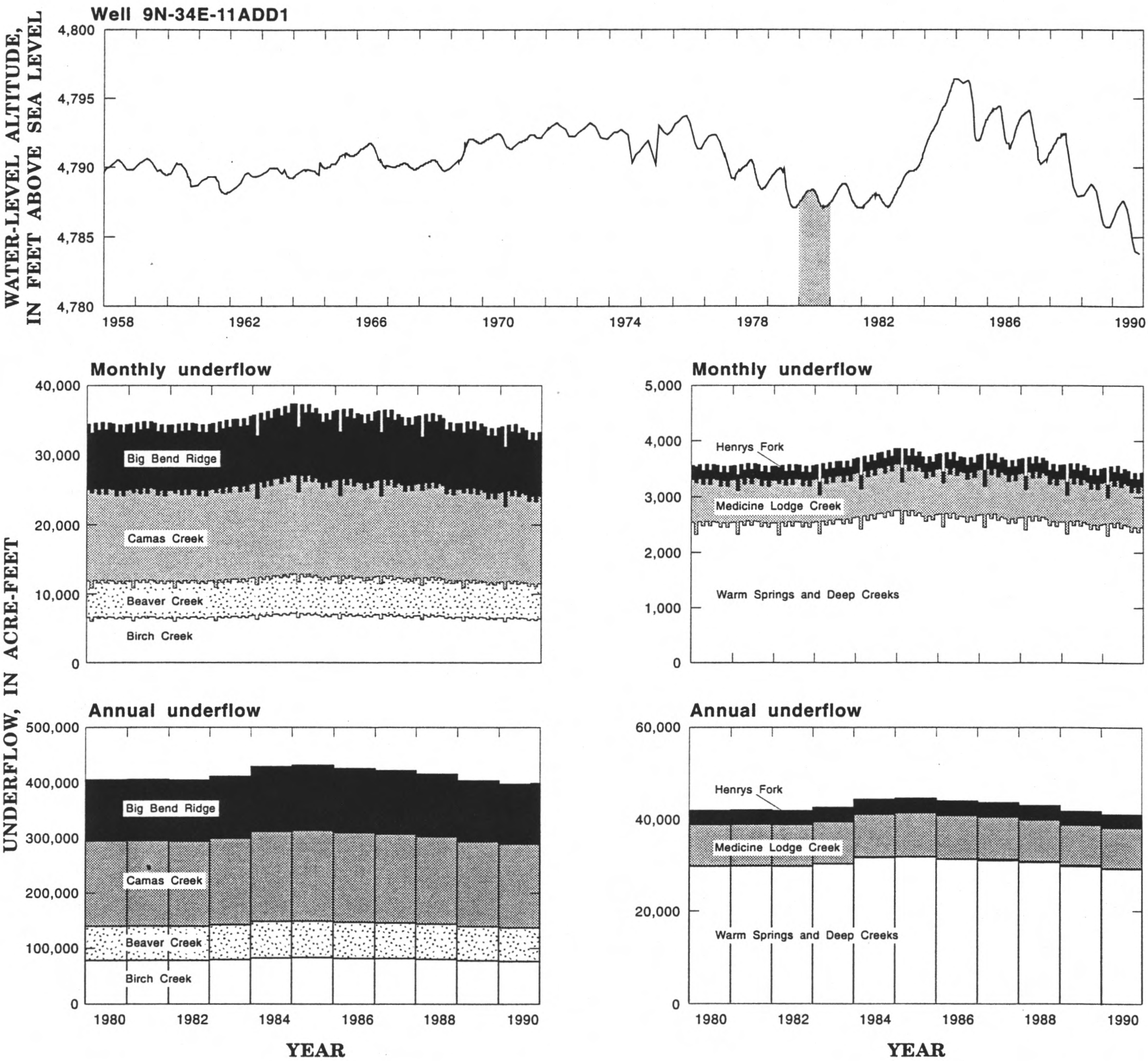


Figure 27. Water level in well 9N-34E-11ADD1, 1958–90, and calculated underflow from tributary basins, 1980–90. (Well location shown on figure 10)

Records on file with the IDWR indicate that ground-water withdrawals for irrigation were made as early as 1896. Widespread ground-water development began in the 1930's, and the number of well permits issued by the IDWR grew exponentially from about 1950 through the early 1980's, then decreased to about the rate that occurred before 1950 (fig. 28). Ground-water withdrawals were calculated from electrical power consumption records starting in 1983, the first year for which records were available in a computer-readable form.

Monthly withdrawals from January 1983 through December 1990 for irrigation wells (fig. 29) that served private farms or wildlife refuges were calculated from electrical power consumption records, irrigation system characteristics, and hydrologic measurements and estimates. A computer tape of monthly electrical power consumption records from 1983 through 1990 was provided by Utah Power and Light Company (Dwight Searle, written commun., 1991). The irrigation system and typical system head (in parentheses) associated with each well were recorded as flood (0 ft), mist pivot (116 ft), handline sprinkler (139 ft), roller sprinkler (139 ft), or impulse pivot (162 ft) (Spinazola and others, 1992, table 2). Values for typical system head were obtained by canvassing irrigation supply companies in the area and were assumed to be uniform for all wells in each irrigation system category. Factors required to estimate withdrawals included determination of pump efficiency from a sample of wells and determination of a nonpumping water level and drawdown at each well. Pump efficiency was determined to be 70 percent from a sample of 20 wells measured during the summer of 1990. Water levels were measured in nonpumping wells in April 1989 and in nonpumping and pumping wells in August and September 1989 (Spinazola and others, 1992). April nonpumping water levels were estimated for unmeasured wells with the kriging statistical technique (Skrivan and Karlinger, 1980). A single value of pumping drawdown in wells was assigned to different parts of the study area (fig. 30) on the basis of differences between August and September pumping water levels minus April nonpumping water levels. The single value was chosen to provide one representative value of drawdown from among the range of drawdown values in each area. The representative drawdown value was assigned only to wells that did not have both pumping and nonpumping measurements. The arithmetic sum of system head, nonpumping water level, and drawdown equals total head. Withdrawals were calculated by multiplication of electrical power consumption by pump efficiency divided by total head.

Ground water that is pumped or flows from wells to serve areas other than private farms is transported in canals to respective service areas (fig. 24). Seepage measurements and reports indicated that part of the water transported by canals returned to the aquifer as canal losses. The treatment of canal losses associated with the transport of ground water was simplified in this study. Ground-water withdrawals were calculated by the method described in the preceding paragraph and then were reduced by a percentage to represent average measured or reported canal loss. The simplification resulted in the same net amount of water withdrawn from the aquifer system but under-represented ground-water withdrawals near wells that served canals and recharge below canals. These effects were considered to compensate one another because they occurred over relatively small areas of less than 5 mi^2 in all but one case.

Wells that serve Mud Lake or areas irrigated by diversions from Mud Lake, Jefferson Canal Company, Monteview Canal Company, and Producers Canal Company (fig. 29) deliver water to canals that transfer water to service areas north and west of Mud Lake (fig. 24). The canals connect networks of wells and traverse as much as several miles of basalt between the wellheads and service areas. Seepage measurements conducted on canals in 1990 indicated that losses from the canals were as high as 50 percent for some canal reaches. Average loss was about 36 percent.

Canals deliver water to Mud Lake by way of Camas Creek from a network of wells east of Mud Lake (fig. 29). Water flows naturally into these canals from some wells during part of the year and is pumped from all wells during the irrigation season in most years. The amount of ground water delivered by canals to Mud Lake (fig. 13) was compiled from Brockway and Robison (1988, appendix B, p. 5–6) and from records on file with IDWR. Most canals traverse only a few miles of sediments between the wellheads and

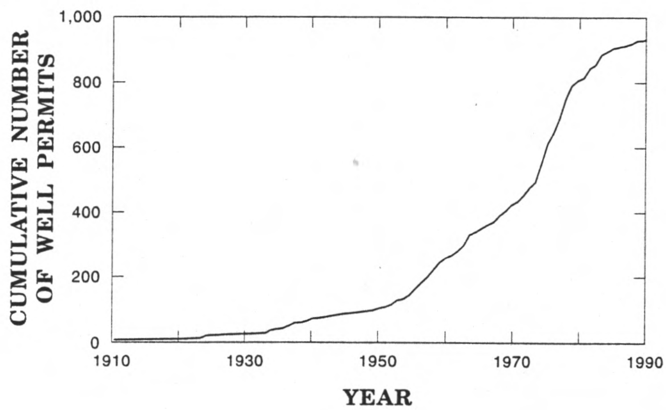


Figure 28. Cumulative number of well permits issued in the study area, 1910–90.

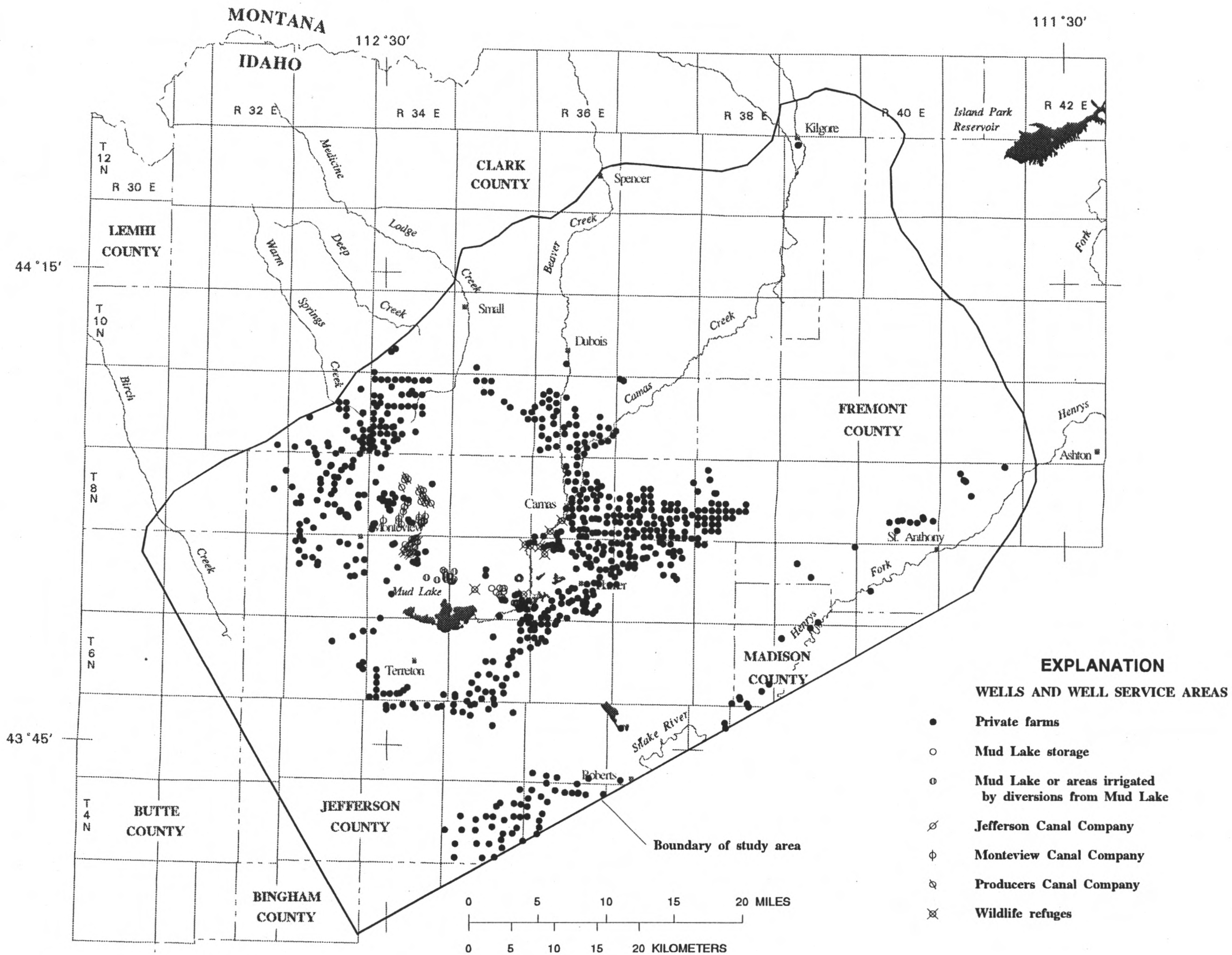


Figure 29. Locations of irrigation wells, 1990.

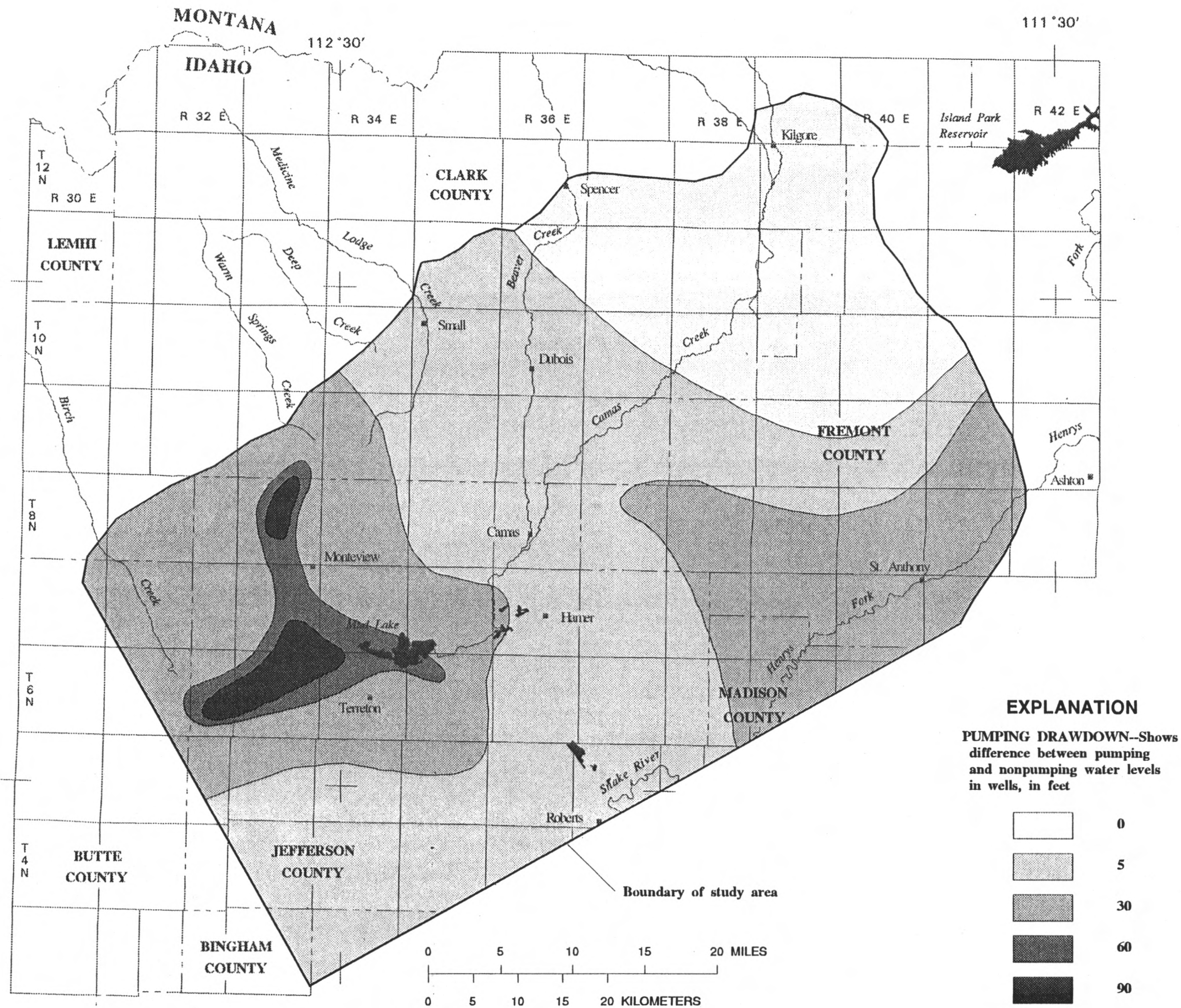


Figure 30. Estimated distribution of pumping drawdown.

Camas Creek. One canal traverses a few miles of basalt. Records on file with the Watermaster of Water District 31 indicate that seepage from these canals averaged about 18 percent during the 1980's. These canal deliveries, and any undiverted streamflow in Camas Creek, are stored in Mud Lake and diverted to service areas south of Mud Lake (fig. 24) during the irrigation season. Ground-water withdrawals on wildlife refuges were not reduced by canal losses because refuge canals were considered to be part of the lake complex in refuge service areas. Canal losses from stream or lake diversions were included in the calculation of surface recharge within respective service areas as described in the preceding section.

Total ground-water withdrawals from wells in all service-area categories increased from about 240,000 acre-ft in 1983 to about 370,000 acre-ft in 1990 (fig. 31). Withdrawals shown in figure 31 were reduced by canal losses for some wells as described in preceding paragraphs. Maps that show distributions of ground-water withdrawals are presented in the section "Boundary Conditions."

Flowing wells were measured by the Watermaster of Water District 31 at Buck Springs in T. 7 N., R. 35 E., sec. 13; at the Owsley wells in T. 7 N., R. 35 E., sec. 25; and at the Holley wells in T. 7 N., R. 35 E., sec. 26 (fig. 32). Wells drilled to depths of 40 to 250 ft in these areas commonly flow during the winter but are pumped once flows decline when the irrigation season begins in the spring. No measurements were made at the Owsley and Holley wells in 1984 and early 1985 because high streamflows in Camas Creek precluded access to these sites.

Aquifer properties from field data

Aquifer properties, which influence the ability of an aquifer to transmit and store water, were determined from aquifer and specific-capacity tests. Results from the few aquifer tests available for the study area (fig. 33) provided values of transmissivity, storage

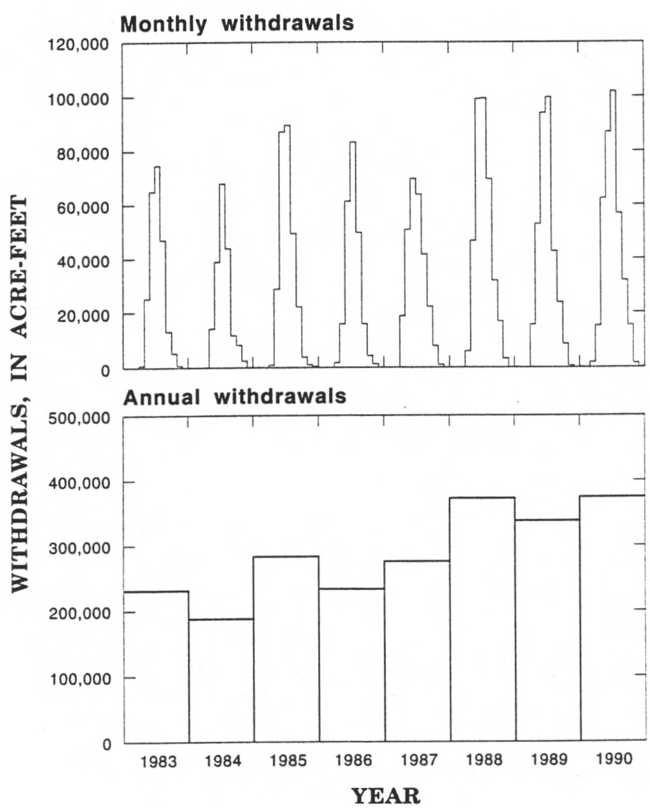


Figure 31. Ground-water withdrawals, 1983–90.

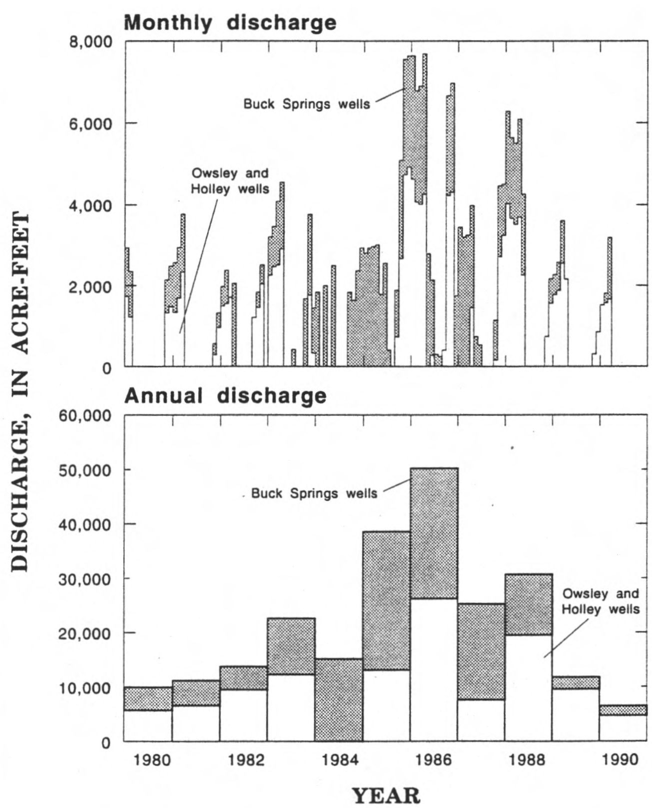


Figure 32. Discharge from Buck Springs (T. 7 N., R. 35 E., sec. 13), Owsley, and Holley (T. 7 N., R. 35 E. secs. 25 and 26) flowing wells, 1980–90.

coefficient, and specific yield and were used with other data to calculate hydraulic conductivity. More prevalent data from specific-capacity tests were used to estimate transmissivity and hydraulic conductivity.

Transmissivity and hydraulic conductivity describe the ability of an aquifer to transmit water. Transmissivity values estimated from three aquifer tests in Jefferson County ranged from about 480,000 to 2,500,000 ft^2/d (Mundorff and others, 1964, p. 155). Wells tested were 106 to 300 ft deep, completed in basalt, and used for irrigation. Hydraulic conductivity was calculated by division of transmissivity by well depth. Under the assumption that the wells fully penetrate the aquifer and the water table was near land surface, hydraulic conductivity as determined from wells tested ranged from about 1,600 to 22,000 ft/d. If the preceding assumptions were incorrect, hydraulic conductivity would be greater.

Storage coefficient and specific yield describe the ability of an aquifer to store and release water. Storage coefficient for most confined aquifers ranges from 0.00001 to 0.001; specific yield of most unconfined aquifers ranges from about 0.1 to 0.3 (Lohman, 1972, p. 8). Storage coefficient and specific yield as determined from the three wells tested in Jefferson County ranged from 0.0008 to 0.19.

An aquifer test was made prior to the irrigation season in April 1990 to determine aquifer properties of sediments, mainly sand and gravel. The test included three irrigation wells in T. 8 N., R. 34 E., sec. 27—one pumped well and two observation wells less than 250 ft from the pumped well. Discharge from the pumped well was maintained at 340 gal/min for 30 consecutive hours. A curve-matching method (Neuman, 1975) was used to calculate values of transmissivity, vertical hydraulic conductivity, and specific yield from time and drawdown measurements in observation wells (table 2). Hydraulic conductivity was calculated from transmissivity and open interval. Differences in aquifer properties between observation wells probably are due to heterogeneity of the sediments and differences in well losses between the pumped well and the observation wells.

Basalt and sediments yield large quantities of water to wells in the study area. Well discharges reported by drillers ranged from 20 to 9,000 gal/min; the median was 2,250 gal/min for 73 sites on file with the IDWR. Pumping drawdown below static water level ranged from 0.3 to 137 ft and the median was 4 ft; specific capacity, or well discharge divided by drawdown, ranged from 3 to 4,490 (gal/min)/ft and the median was 625 (gal/min)/ft. Specific capacity is affected by well construction and development, degree of connection between the well and the aquifer, and the velocity and length of flow up the well casing. Specific capacity is roughly proportional to aquifer transmissivity when the relation between well discharge and drawdown does not change with time (Lohman and others, 1972, p. 11).

Comparisons among median values of well discharge, pumping drawdown, and specific capacity were made for wells completed in sediments and those completed in basalt. Wells for which discharge and drawdown were reported by drillers were grouped by type of rock that composes the aquifer—basalt, sediments, and basalt and sediments (fig. 33). Greater median values of discharge and specific capacity and a lesser median value of drawdown for wells completed in basalt (fig. 34) indicate that wells in basalt typically yield more water than wells in sediments. The range of drawdown values was large in all rock types. The similarity among discharge, drawdown, and specific capacity between wells in basalt and wells in basalt and sediments suggests that basalt contributes large quantities of water to wells completed in basalt and sediments. For the study area in general, wells completed in basalt typically yield an adequate amount of water for intended use with small drawdown. Wells completed in coarse sand and gravel typically yield adequate supplies of water with moderate drawdown; wells completed in silt and clay typically yield inadequate supplies of water for most uses, and drawdown is large.

Specific capacity was used with other well data to estimate transmissivity with the equation (Theis and others, 1963, p. 332, eq. 1):

$$T = 15.32(Q/s)(-0.577 - ln(r^2S/4Tt)), \quad (5)$$

where

T = transmissivity, in feet squared per day;
Q/s = specific capacity of the well, in gallons per minute per foot of drawdown;
r = effective radius of the pumped well, in feet;
S = specific yield, dimensionless; and
t = time of the specific-capacity test, in days.

The equation was modified from the original to allow for different units and was solved by iteration because transmissivity is present on both sides of the equation. Only tests that included measurements of specific

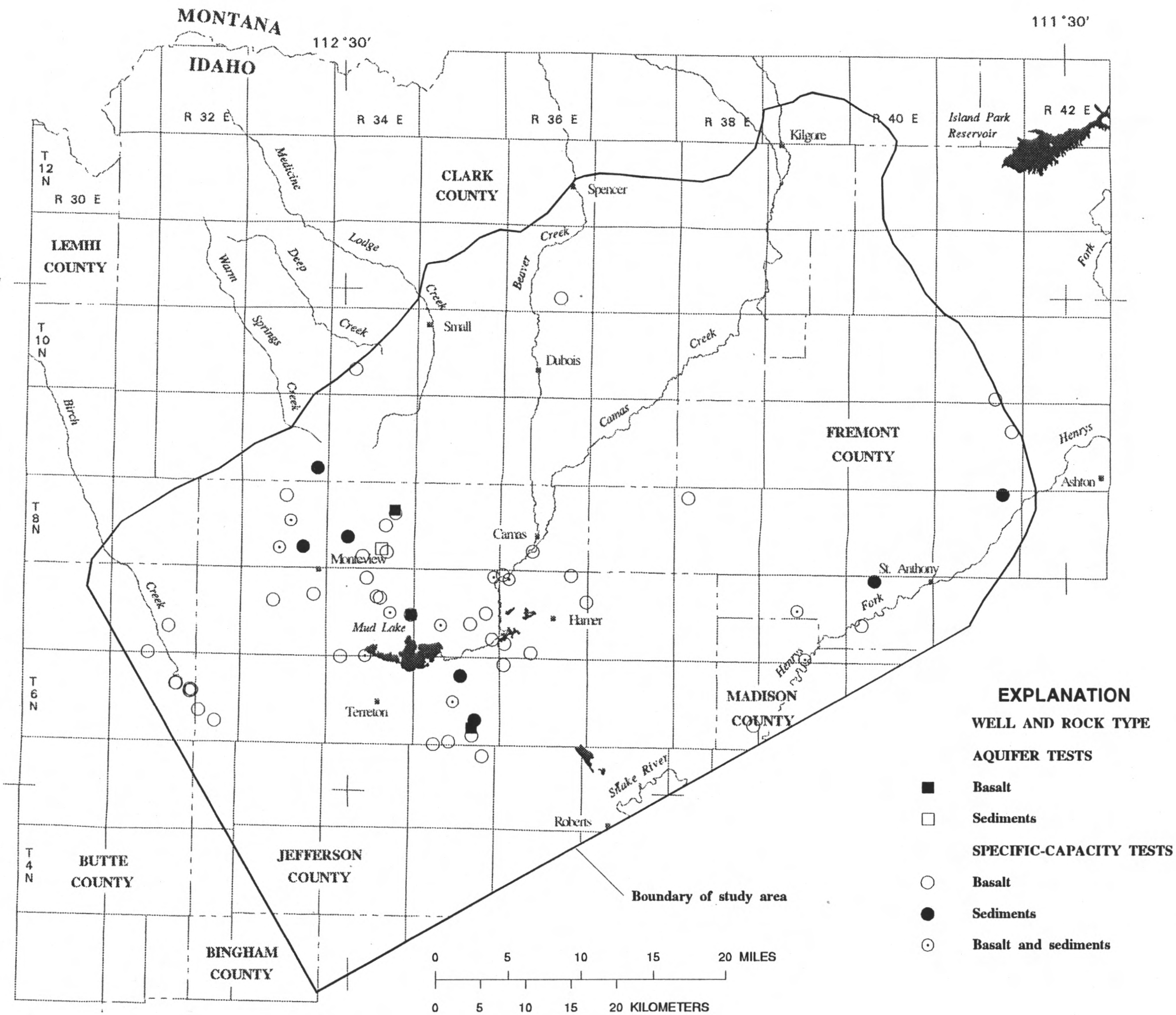


Figure 33. Sites where aquifer and specific-capacity tests were made.

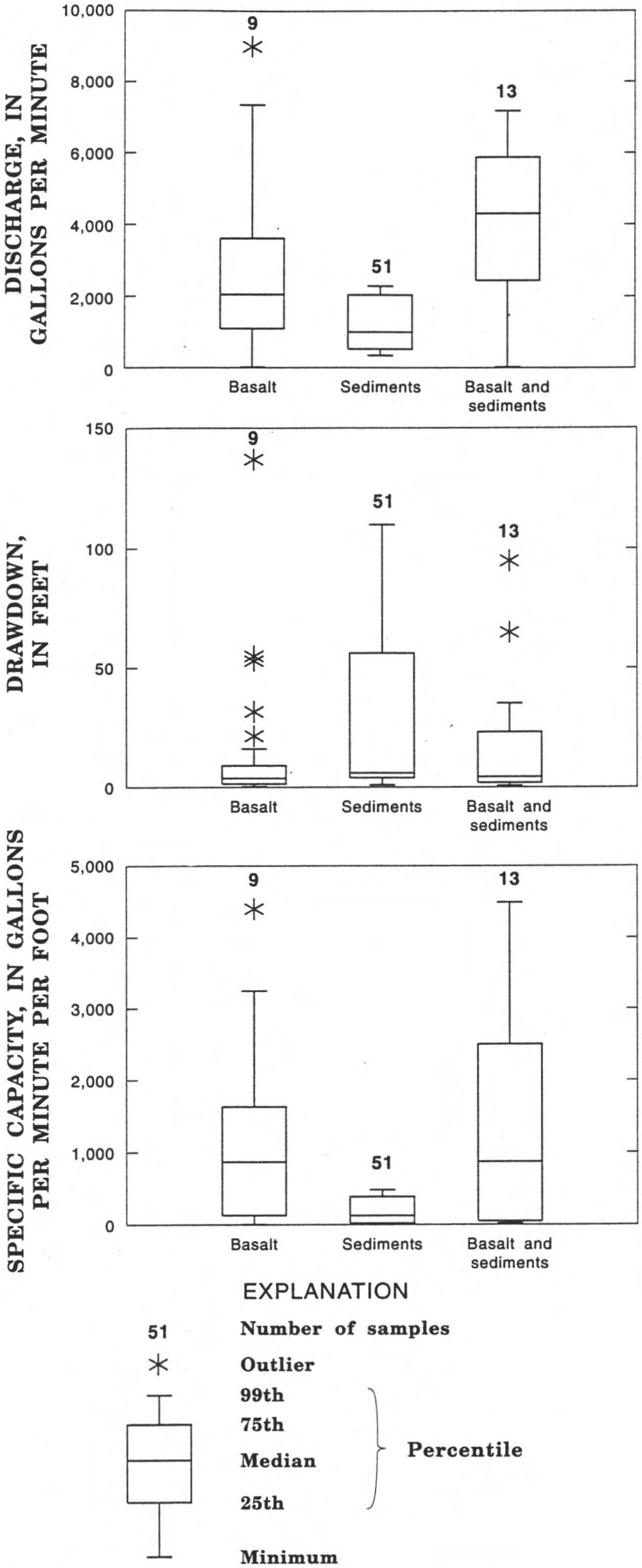


Figure 34. Discharge, drawdown, and specific capacity for wells completed in basalt, sediments, and basalt and sediments.

Table 2. Well data and aquifer properties from an aquifer test

[Well locations are shown on figure 33]

Well number	Well depth (feet)	Open interval (feet below land surface)	Trans-mis-sivity (feet squared per day)	Hy-drau-lic conduc-tivity (feet per day)	Vertical hy-drau-lic conduc-tivity (feet per day)	Specific yield (dimen-sion-less)
8N-34E-27AAD1	73	28–73	6,300	140	4	0.12
8N-34E-27AAD3	61	41–57	5,300	330	1.5	.17

capacity and time were used to estimate transmissivity. An effective well radius of 8 in. and a specific yield of 0.1 were assumed for all tests. Drillers' logs indicate that the drilled diameter of most wells below the water table was 16 in. The value for specific yield was chosen to represent unconfined conditions.

Median transmissivity estimated from wells completed in basalt was 180,000 ft^2/d; from wells completed in sediments, 43,000 ft^2/d; and from wells completed in basalt and sediments, 200,000 ft^2/d (fig. 35). Estimates were about 2 to 10 times greater from wells completed in basalt or in basalt and sediments than from wells completed in sediments. A rough estimate of hydraulic conductivity was obtained by division of transmissivity estimates by the length of the well open to the aquifer system as described on drillers' logs. Median hydraulic conductivity estimated from wells completed in basalt was 1,200 ft/d; from wells completed in sediments, 780 ft/d; and from wells completed in basalt and sediments, 1,500 ft/d. Estimates were as much as four times greater from wells completed in basalt than from wells completed in sediments.

Results of aquifer and specific-capacity tests help to identify relative differences in aquifer properties between rock types. Aquifer tests indicated that trans-missivity of basalt in the study area is about 100 to 1,000 times greater than transmissivity of sediments. Specific-capacity tests indicated the same relation to a lesser degree. Results of the tests indicate aquifer properties in the immediate vicinity of tested wells. Most wells are drilled to procure an abundant supply of water for irrigation, and most aquifer and specific-capacity tests are in areas where the aquifer system is most productive. However, transmissivities in many parts of the aquifer system could be much less than in areas where the tests were made.

Transmissivity and hydraulic conductivity of basalt are as variable as the lithologic character of the rock within individual and among successive flows. Flow centers that consist of dense, fine-grained basalt with few interconnected voids provide few avenues for water movement. Interflow zones between successive flows that contain clinkers, cinders, and rubble provide ample interconnected voids for water movement. Hydraulic conductivity of basalt can range from 1.3×10^{-8} to 13,000 ft/d (Freeze and Cherry, 1979, p. 29). A value of 0.03 ft/d was determined from an aquifer test for an interval of mostly basalt in a test hole about 20 mi southwest of Terreton (Mann, 1986, p. 7). Hydraulic conductivity of basalt in the study area determined from aquifer and specific-capacity tests is typically on the higher end of or exceeds the cited range.

Transmissivity and hydraulic conductivity of sediments are affected by the distribution and proportion of clay, silt, sand, and gravel. Clay and silt are prevalent in the subsurface around Mud Lake. Hydraulic conductivity of clay and silt can range from 1.3×10^{-7} to 13 ft/d (Freeze and Cherry, 1979, p. 29). Clay and silt grade into coarse sand and gravel between Mud Lake and alluvial fans near the northern boundary of the plain. Hydraulic conductivity of sand and gravel can range from about 0.13 to 130,000 ft/d (Freeze and Cherry, 1979, p. 29). Hydraulic conductivity of sediments in the study area determined from aquifer and specific-capacity tests is in the middle part of this range.

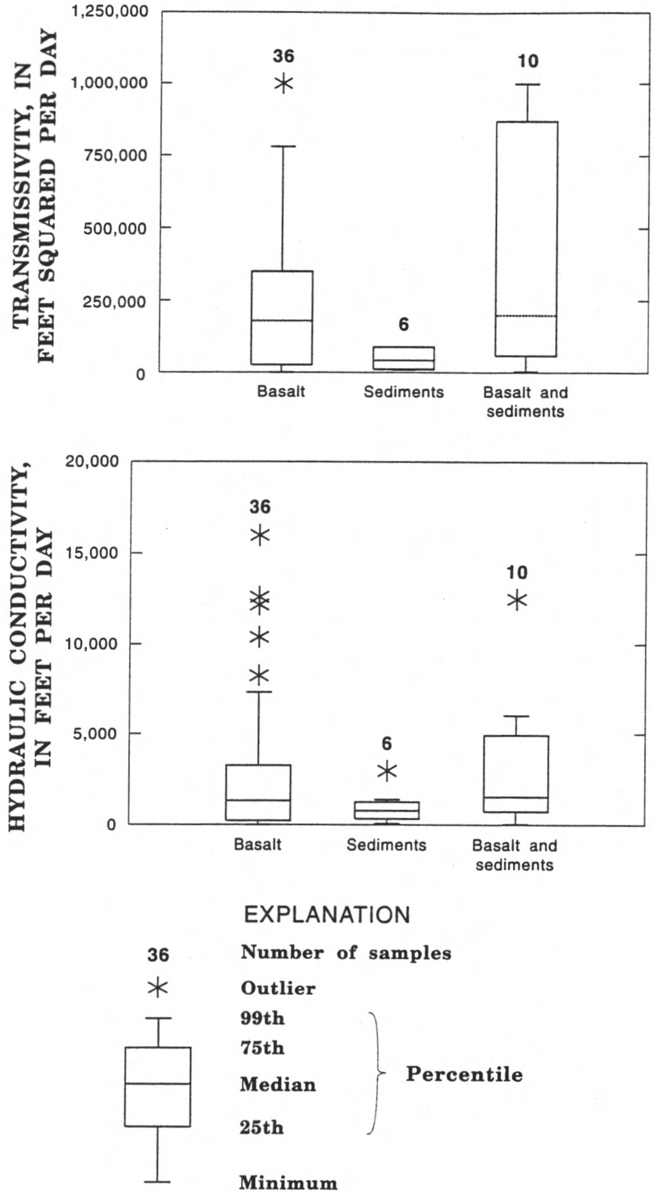


Figure 35. Transmissivity and hydraulic conductivity estimated from specific-capacity tests for wells completed in basalt, sediments, and basalt and sediments.

NUMERICAL MODEL

The USGS modular, three-dimensional, finite-difference ground-water flow model (McDonald and Harbaugh, 1988) is a computer program that was used to simulate surface- and ground-water flow and water levels in the aquifer system from a mathematical synthesis of data sets derived from geohydrologic data. The model was developed with data that describe aquifer geometry and boundaries, recharge, discharge, and aquifer properties of the Mud Lake study area to obtain a better understanding of the geohydrology and to provide a tool to evaluate water-use alternatives. Most of these data were derived independently of the model and were presented in preceding sections. However, some data, called calibration variables, were adjusted from initial estimates and were finalized during model calibration. The model was calibrated to steady-state conditions for the 1980 calendar year. A copy of the model and associated data sets is on file in the USGS office in Boise, Idaho.

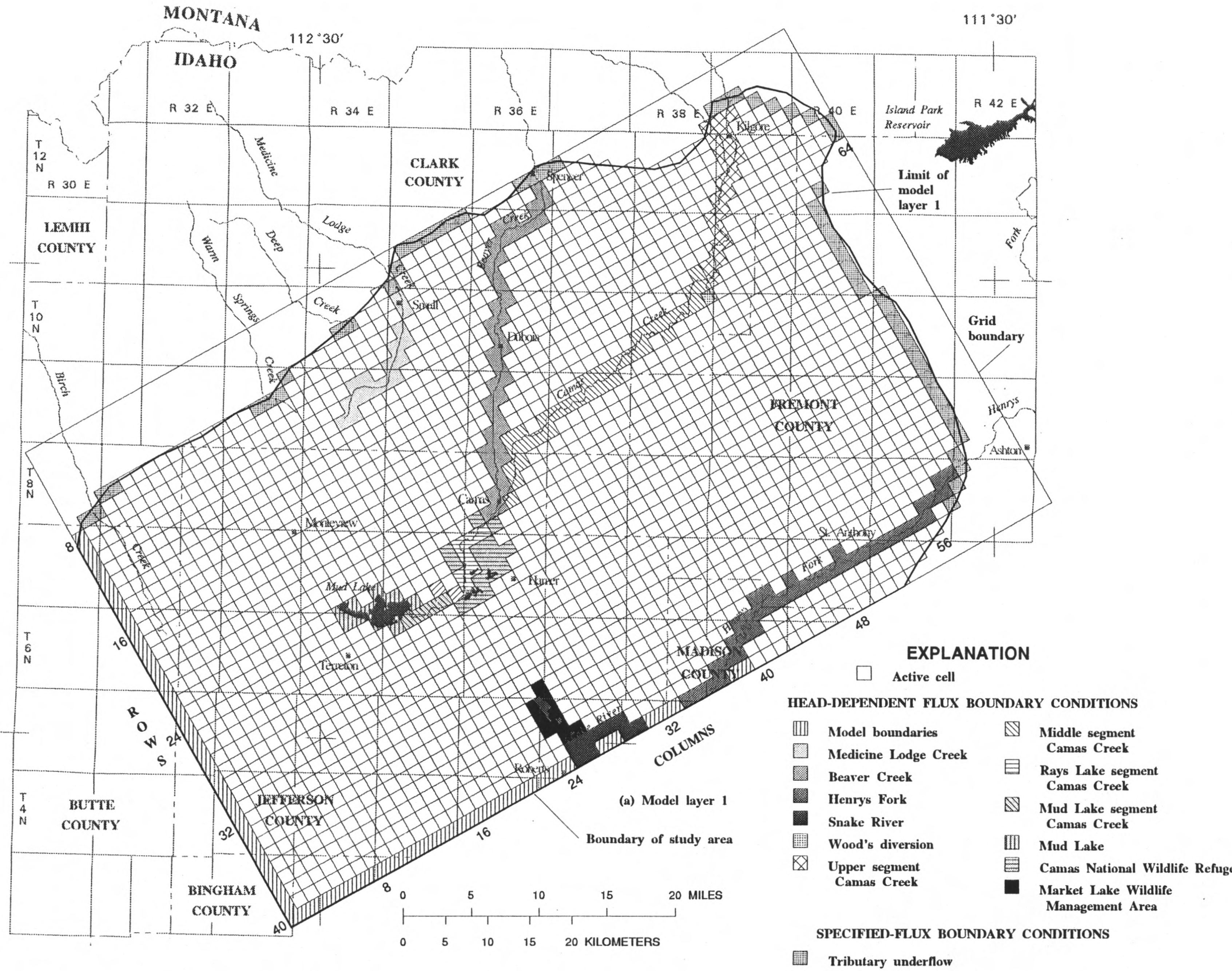


Figure 36. Grid and boundary conditions for model layers 1–5.

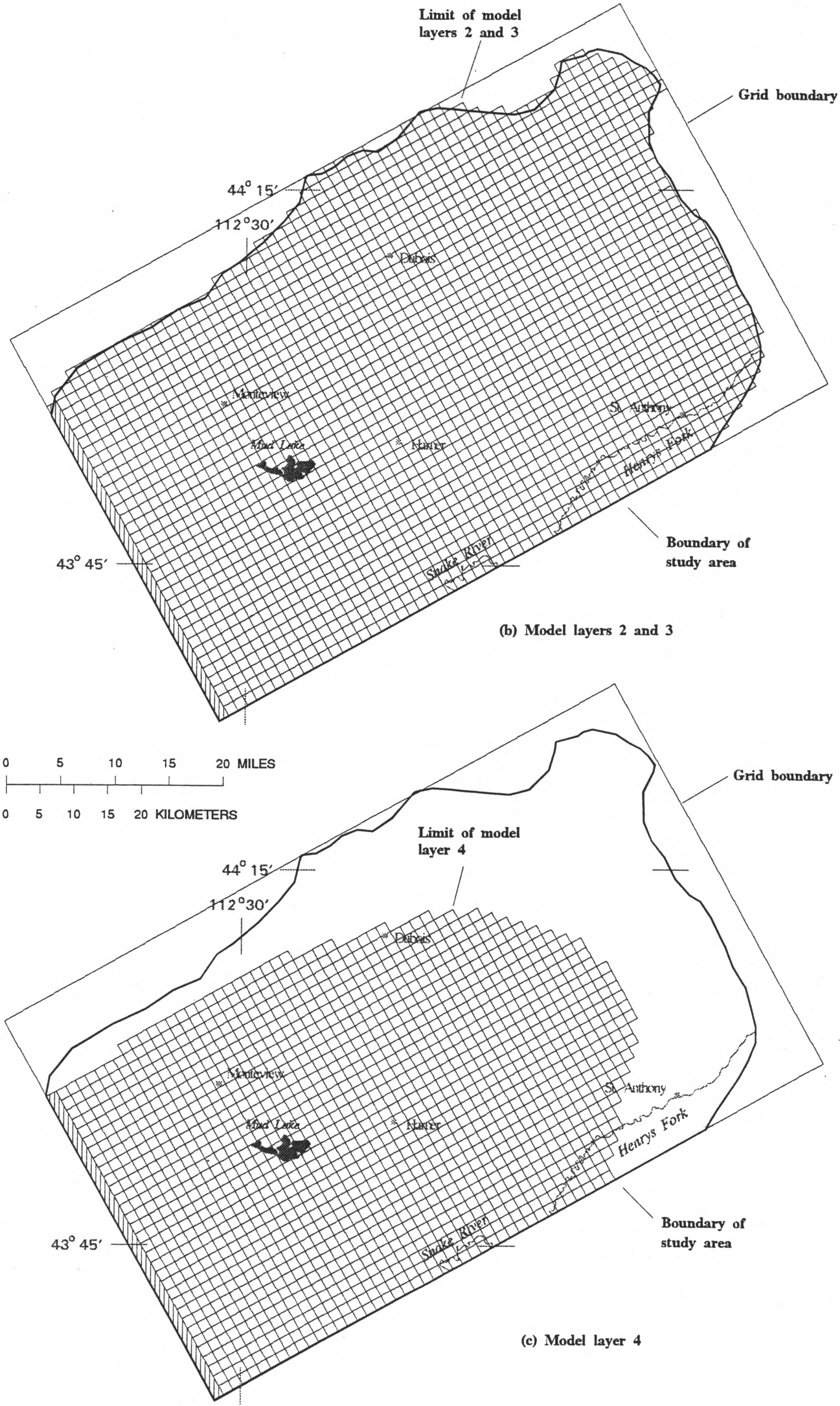


Figure 36. Grid and boundary conditions for model layers 1–5—Continued.

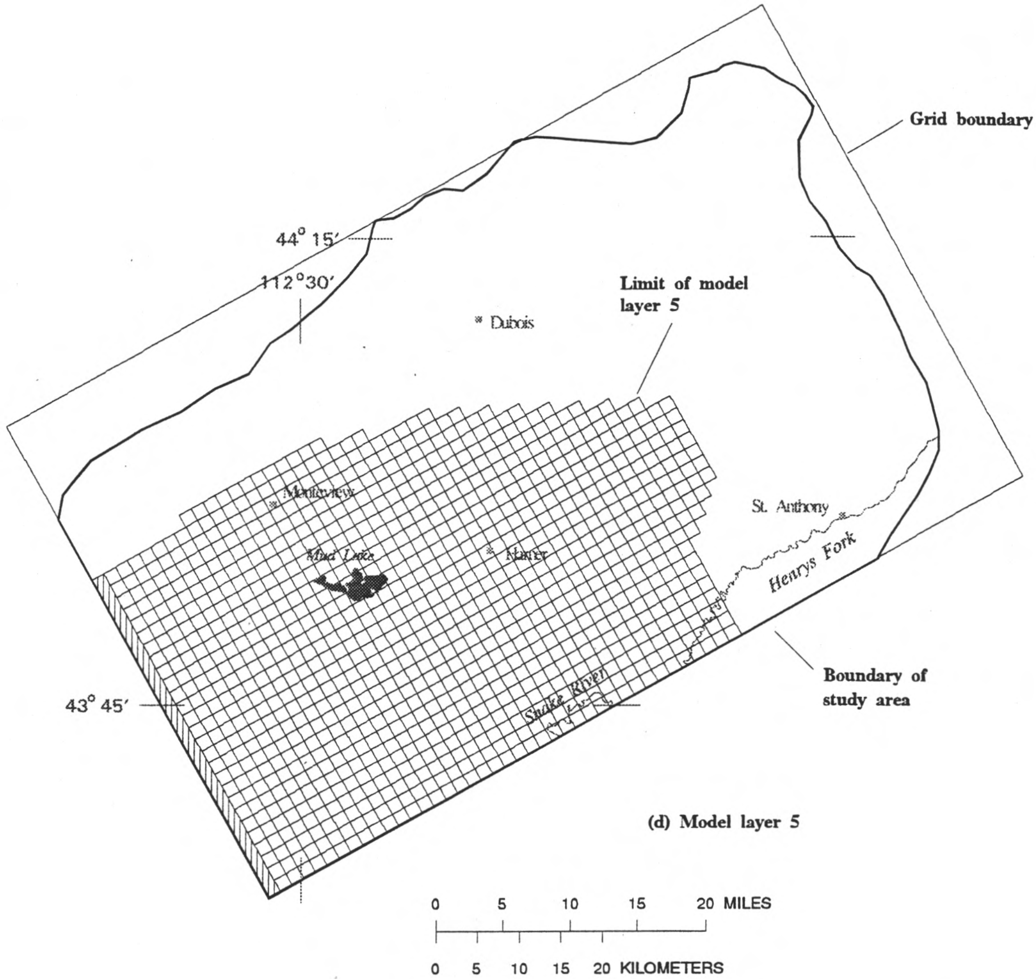


Figure 36. Grid and boundary conditions for model layers 1–5—Continued.

Model Grid

The model is delineated by a grid of cells 40 rows long by 64 columns wide by 5 layers deep and is aligned with the 4-layer model grid used to simulate ground-water flow and water levels in the eastern Snake River Plain aquifer system (Garabedian, 1992, p. 38). The grid was rotated counterclockwise 31° 24' from its origin at the lower left-hand corner at latitude 44° 02' 53" north, longitude 112° 55' 37" west to align the grid with the principal direction of ground-water flow. Grid dimensions were chosen to provide an adequate representation of available data and smaller grid dimensions were not warranted. Cells along rows and columns are 1 mi on a side. The lateral extent of active cells in layers 1, 2, and 3 is identical (fig. 36); the lateral extent of active cells in layers 4 and 5 decreases successively to represent thinning of the aquifer system toward the margin of the plain. Cells in layers 1, 2, and 3 represent constant layer thicknesses of 100, 100, and 300 ft that correspond to the maps of subsurface geology for selected intervals below the water table (fig. 6a,b,c). Cells in layers 4 and 5 are of variable thickness dependent upon the thickness of saturated basalt and sediments (fig. 6d,e) and represent thicknesses of 500 ft or less and 1,000 ft or less, respectively. Active grid cells represent a three-dimensional volume of the aquifer system and were assigned representative values for aquifer properties, boundary conditions, recharge, and discharge, as described in the next two sections.

Aquifer Properties

Hydraulic conductivity and transmissivity values were calibration variables that were assigned by zones to active cells in the model grid. A single, unique vertical conductance value was assigned between active cells in each pair of adjacent model layers.

Maps of saturated basalt thickness and predominant rock type for specific intervals below the water table (fig. 6) were used to develop a horizontal hydraulic conductivity distribution that corresponded to each model layer (fig. 36). Thicknesses of four rock types—basalt, fine-grained sediment, coarse-grained sediment, and rhyolite—in each active grid cell within each layer were calculated using a commercially available geographic information system (GIS). Initial values of hydraulic conductivity then were calculated for each grid cell with the equation:

$$K_{cell} = (K_b{*}T_b + K_f{*}T_f + K_c{*}T_c + K_r{*}T_r) \,/\, (T_b + T_f + T_c + T_r), \qquad (6)$$

where

K = hydraulic conductivity, in feet per day;
T = thickness, in feet;
$cell$ = cell;
b = basalt;
f = fine-grained sediments;
c = coarse-grained sediments; and
r = rhyolite.

Initial values of horizontal hydraulic conductivity were 5,000 ft/d for basalt, 5 ft/d for fine-grained sediments, 125 ft/d for coarse-grained sediments, and 25 ft/d for rhyolite. All values are within ranges for these rock types as described in the section "Aquifer properties from field data." Hydraulic conductivity was reduced by one-third in model layer 4 and two-thirds in model layer 5 to represent decreasing hydraulic conductivity with depth (Garabedian, 1992, p. 42). Within each layer, initial hydraulic conductivity values for each cell were grouped into zones that were assigned a single hydraulic conductivity value for each cell in the zone. Zones represented areas that were composed of a predominant rock type and (or) various proportions of several different rock types. Initial values of hydraulic conductivity specified for each zone were adjusted and finalized (fig. 37) during model calibration. Transmissivity values for layers 2 through 5 were obtained by multiplication of hydraulic conductivity and layer thickness.

Vertical conductance between cells in vertically adjacent layers (table 3) was calculated using a relation between vertical hydraulic conductivity and layer thickness (McDonald and Harbaugh, 1988, chap. 5, p. 13, eq. 51). Vertical hydraulic conductivity values in sediments (mostly sand and gravel) calculated from an aquifer test that included two observation wells (table 2) were 1.5 and 4.0 ft/d. Several model runs were made to associate vertical hydraulic conductivity values with rock types, but the effort was discontinued when no meaningful relation could be established. Vertical hydraulic conductivity likely varies widely due to fractures, basalt density, and lithologic discontinuities, among many other reasons. Vertical hydraulic conductivity was arbitrarily chosen to be a constant 1 ft/d for all active cells in the grid.

Table 3. Vertical conductance between adjacent model layers

Model layers	Vertical conductance (foot per day per foot)
1–2	0.01
2–3	.005
3–4	.0025
4–5	.0013

Boundary Conditions

Four types of boundary conditions were used in the model: (1) no flow, (2) head-dependent flux, (3) free surface, and (4) specified flux. The first three types describe how flow was simulated along external boundaries of the model; the fourth describes how recharge and discharge were assigned to active cells in the model grid.

NO FLOW

No-flow boundaries were specified to represent the natural extent of the aquifer system and a flowline along part of the southeastern boundary of the study area. A no-flow boundary was specified at the limit of each model layer except for model boundaries identified with head-dependent flux boundary conditions (fig. 36). As the name implies, the model does not simulate underflow across a no-flow boundary (Franke and others, 1987, p. 3). A no-flow boundary was specified in all model layers where the aquifer abuts mountains along the northwestern and northeastern margins of the plain. Mountain slopes that directly abut the northwestern margin of the plain were considered to contribute insignificant amounts of underflow to the study area because of their relatively small drainage areas. However, tributary basins that intersect the northwestern margin and Big Bend Ridge (fig. 1), which composes most of the northeastern margin, contribute significant amounts of underflow. Underflow from tributary basins and Big Bend Ridge was represented with a boundary condition described in the section "Specified Flux." No-flow, or streamline, boundaries were specified in layers 2–5 along the southeastern model boundary, which approximates a flowline. A flowline along this boundary is implied by water-table contours that are nearly perpendicular to the boundary (fig. 21). A no-flow boundary was specified south of the Henrys Fork along the southeastern boundary in layer 1. The natural extent of the aquifer is less than 3 mi from the model boundary and ground-water inflow is believed to be negligible in the intervening distance because recharge from the small area of contribution is relatively low. A no-flow boundary also was specified to represent the effective base of the aquifer system along the bottom of all active cells in layer 5 and along the bottom of cells in layers 4 and 3 with no active cells below.

HEAD-DEPENDENT FLUX

Head-dependent flux boundaries (Franke and others, 1987, p. 4) were used to simulate underflow between the modeled area and areas in the eastern Snake River Plain aquifer system adjacent to the model, losses from and gains to streams and lakes, and discharge from flowing wells. Underflow between the modeled area and areas in the eastern Snake River Plain aquifer system adjacent to the model was simulated with the general-head boundary package (McDonald and Harbaugh, 1988, chap. 11). Head-dependent flux boundaries were specified along the southwestern boundary of the model grid in layers 1–5 to simulate underflow from the model area (fig. 36). Head-dependent flux boundaries along the southeastern boundary of the grid in layer 1 were specified to permit simulation of underflow across the boundary in response to changes in recharge and ground-water withdrawals within the grid. Much of the area between the southeastern model boundary and the margin of the plain is irrigated with streamflow diversions and only a small amount of underflow into or out of the study area is possible across this boundary. The general-head package required specifications of (1) conductance of the material between the boundary and a fixed head outside of the model grid and (2) a value for fixed head. Conductance was calculated by multiplication of the hydraulic conductivity assigned to the active cell adjacent to the boundary (fig. 37) by the cross-sectional area of the cell face along the boundary divided by the distance between the boundary and the fixed head. Distance between the southwestern boundary and fixed head was arbitrarily selected to be 1 mi. Distance between the southeastern boundary and the fixed head is 1 to 30 mi and was determined as the distance from the model boundary to the boundary of the plain (Whitehead, 1986, sheet 1) or to the highest water-table altitude encountered before the boundary of the plain was reached (Lindholm and others, 1988). Fixed head for the southwestern boundary of the grid could not be represented adequately from available piezometric data and was treated as a calibration variable. The method used to obtain fixed heads along the southwestern boundary is presented in the section "Model Calibration." Fixed head for the southeastern boundary of the grid was described by the water-table altitude at the boundary of the plain or the highest water-table altitude before the boundary of the plain was reached (Lindholm and others, 1988).

Stream and lake losses and gains were simulated with the streamflow-routing package (Prudic, 1989). Head-dependent flux boundaries were specified for cells in model layer 1 (fig. 36a) that corresponded to stream and lake segments (fig. 9). Each segment is composed of one or more reaches. A reach is that part of a cell occupied by a stream or lake. Losses and gains were calculated for cells that included stream and lake reaches with the equation (Prudic, 1989, p. 7):

$$Q = (H_S - H_a)\ CSTR, \tag{7}$$

where

Q = loss to or gain from the aquifer, in cubic feet per day;

H_S = stage in the stream, in feet;

H_a = water-level altitude in the aquifer, in feet; and

$CSTR$ = conductance of the streambed, in feet squared per day, which is the hydraulic conductivity of the streambed times the product of the width and length of the stream divided by the thickness of the streambed.

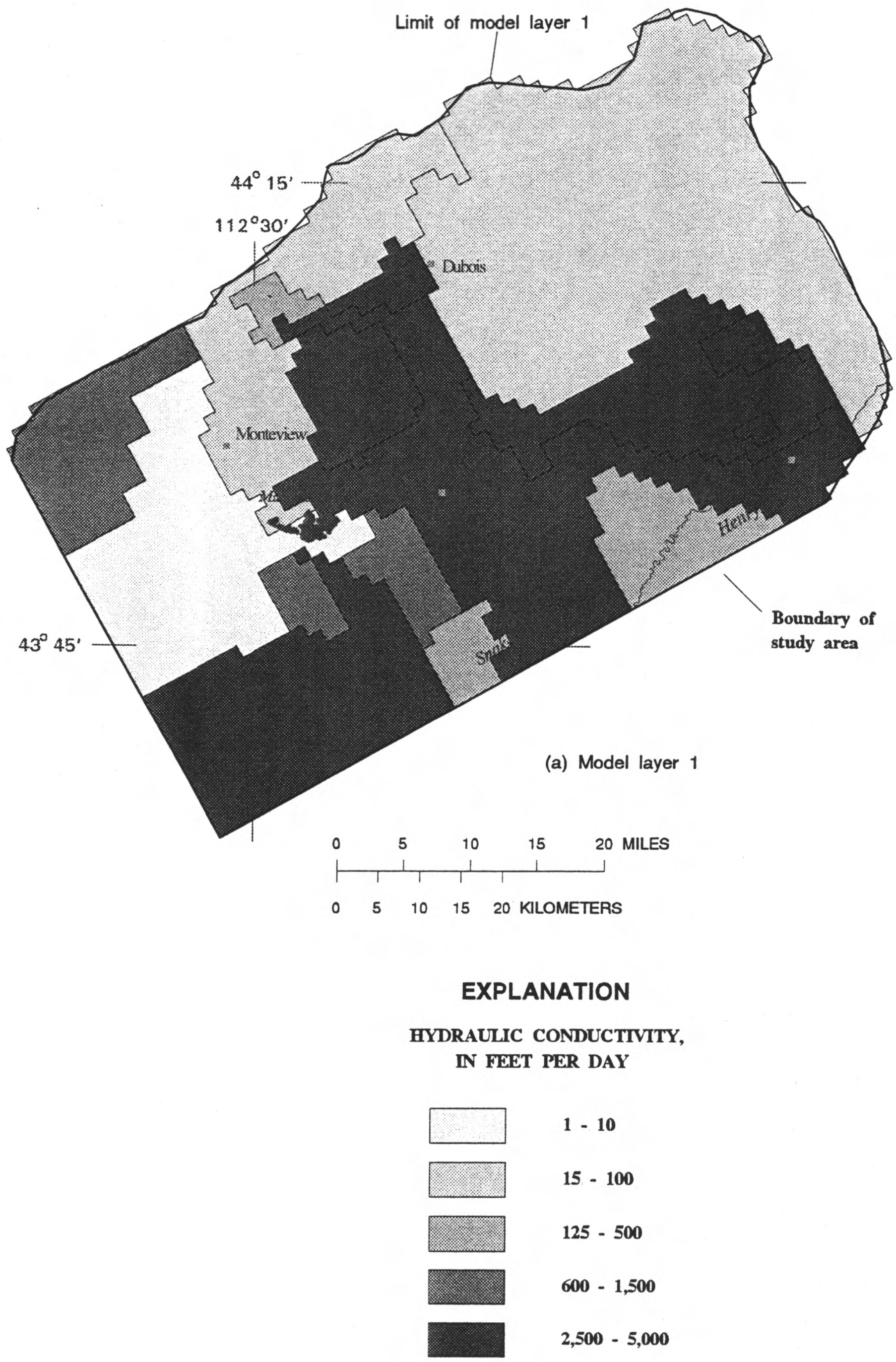


Figure 37. Horizontal hydraulic conductivity zones for model layers 1–5.

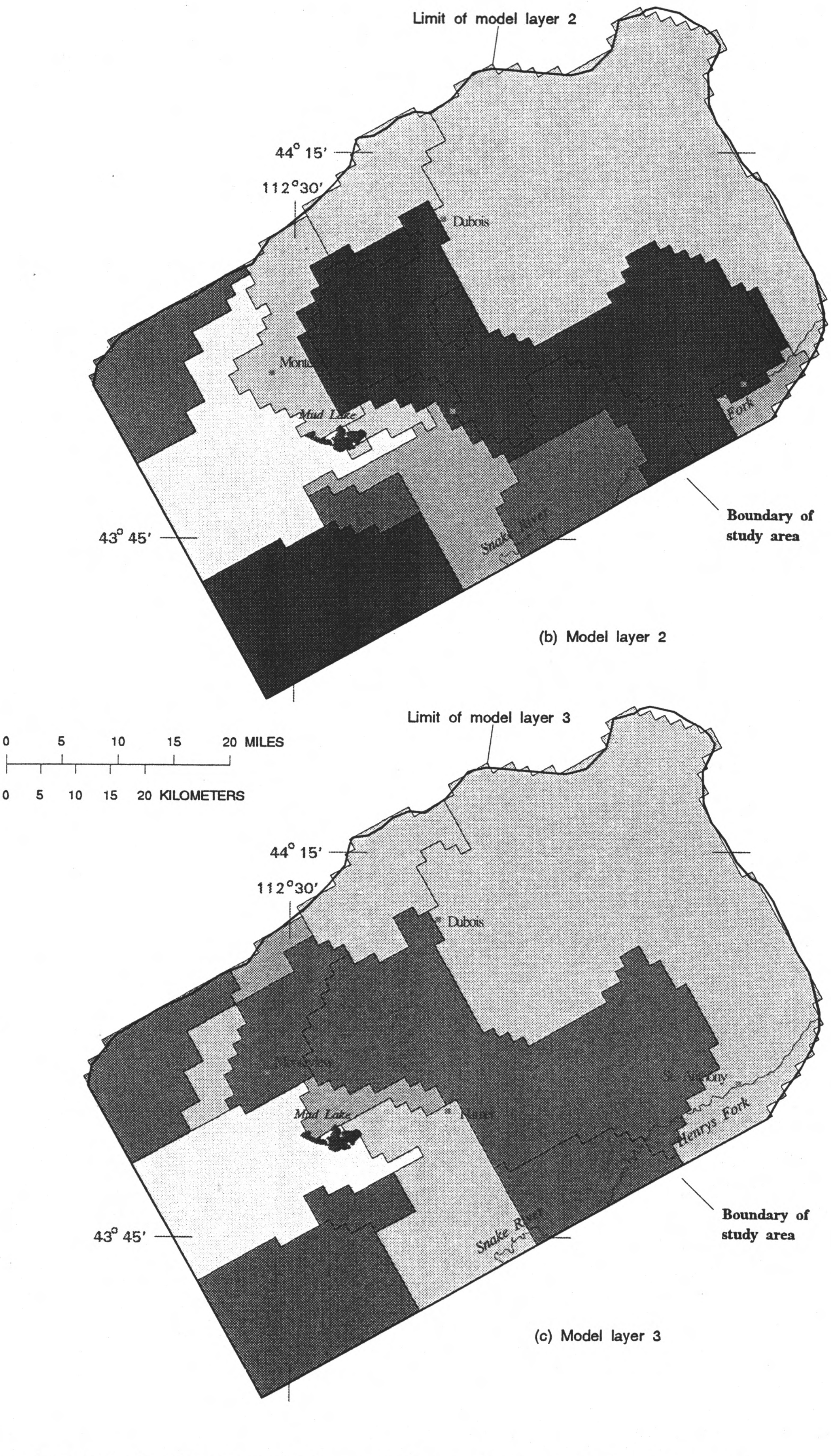
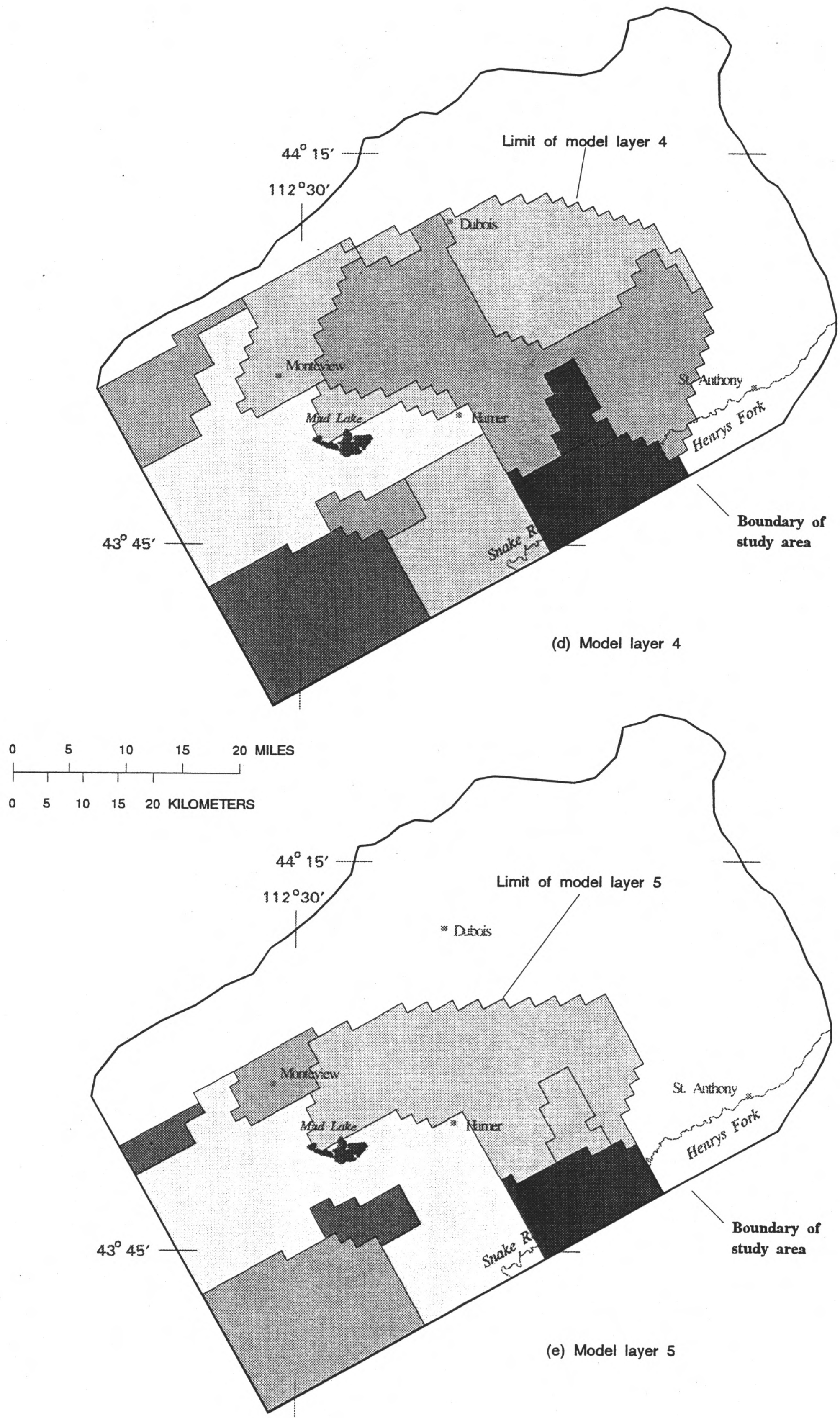


Figure 37. Horizontal hydraulic conductivity zones for model layers 1–5 — Continued.



(d) Model layer 4

(e) Model layer 5

Figure 37. Horizontal hydraulic conductivity zones for model layers 1–5—Continued.

Positive results signify streamflow losses; negative results signify streamflow gains.

Stream stage, water-level altitude, streambed bottom, streambed conductance, streambed top, streamflow, streamflow diversions, and streamflow returns were among the data required by the streamflow-routing package. Stage in the stream was estimated from topographic maps at a point near the middle of each reach. Water-level altitude in the aquifer usually was simulated. The simulated water level in the aquifer was replaced with the altitude of the bottom of the streambed in equation 7 whenever the simulated water level was less than the bottom of the streambed. This condition represents a constant loss from the stream to the aquifer. Altitude of the bottom of the streambed was specified differently for different stream and lake segments.

Altitude of the bottom of the streambed was specified as the simulated water level in the aquifer for reaches in stream segments that represented Medicine Lodge Creek, Beaver Creek, and the upper and middle segments of Camas Creek to approximate the general condition of streamflow losses due to large differences between stage and water level in the aquifer (McDonald and Harbaugh, 1988, chap. 6, p. 10–12). Losses greatly outnumbered gains in these segments (figs. 17 and 18), and differences between stage and water level in the aquifer ranged from several to a few hundred feet. Simulated water level in the aquifer provided initial values for altitude of the bottom of the streambed for most other stream and lake reaches. Initial values were finalized during model calibration. Altitude of the bottom of the streambed was chosen arbitrarily as 5 ft below stage for Henrys Fork and the Snake River and 1 ft below stage for lakes on the Market Lake WMA.

Streambed conductance was specified for each stream reach as defined in equation 7. Stream width and length were replaced by lake area for lake reaches. Stream widths were chosen from field observation: 15 ft for Medicine Lodge and Beaver Creeks, 20 ft for Wood's diversion and most of Camas Creek, 30 ft for the Mud Lake segment of Camas Creek, 100 ft for the Snake River, and 200 ft for Henrys Fork. Stream lengths and lake areas were obtained from digital data processed with a GIS. Streambed thickness was unknown and was chosen arbitrarily to be 1 ft. Equation 7 was rearranged to obtain initial values for hydraulic conductivity in each stream and lake reach using values for stage; water level in the aquifer; streambed width, length, and thickness described previously; and values for reach losses and gains. Stream reach losses and gains were distributed by multiplication of 1980 segment loss or gain (fig. 18) and reach length divided by segment length. Reach and segment lengths were replaced by reach and segment area for lakes. Initial values for streambed hydraulic conductivity were finalized during model calibration.

Streamflow is routed by the model through the network of stream and lake reaches dependent on specified values for streamflow, streamflow diversions, streamflow returns, and simulated losses and gains. Streamflow that enters the study area (fig. 12) was specified at the farthest upstream reach for Medicine Lodge, Beaver, and Camas Creeks and the Snake River. Streamflow (fig. 12) minus streamflow diversions (fig. 15) was specified at the farthest upstream reach for Henrys Fork. Diversions to Wood's diversion (fig. 15) were specified for the upper segment of Camas Creek. Irrigation diversions were specified for Beaver Creek, the middle segment of Camas Creek, and Mud Lake. Diversions to Camas National Wildlife Refuge were specified for the Rays Lake segment of Camas Creek. Lake ET from Mud Lake and lakes on Camas National Wildlife Refuge (fig. 16) also was specified as diversions. Diversions were used to specify ET from Mud Lake and lakes on the Camas National Wildlife Refuge. Ground-water inflow to Mud Lake (fig. 13) was specified as a streamflow return to the Mud Lake segment of Camas Creek. If streamflow in a reach was zero and the water level in the aquifer was less than the altitude of the top of the streambed, no loss was calculated for the reach. Because streambed thickness was chosen to be 1 ft, altitude of the top of the streambed was specified to be 1 ft above the altitude of the bottom of the streambed.

Discharge from flowing wells was simulated with the drain package (McDonald and Harbaugh, 1988, chap. 9). Head-dependent flux boundaries were specified in two model cells that contained flowing wells (fig. 36). Drain cells used to represent flowing wells required specifications for drain altitude and conductance. Altitude was set to 5 ft below simulated water level. Most flowing wells in the study area penetrated less than 100 ft of the aquifer system in 1980. Initial values for drain conductance were determined by division of hydraulic conductivity in the drain cell by 60, the average depth of the aquifer penetrated by flowing wells. Drain conductance was finalized during model calibration.

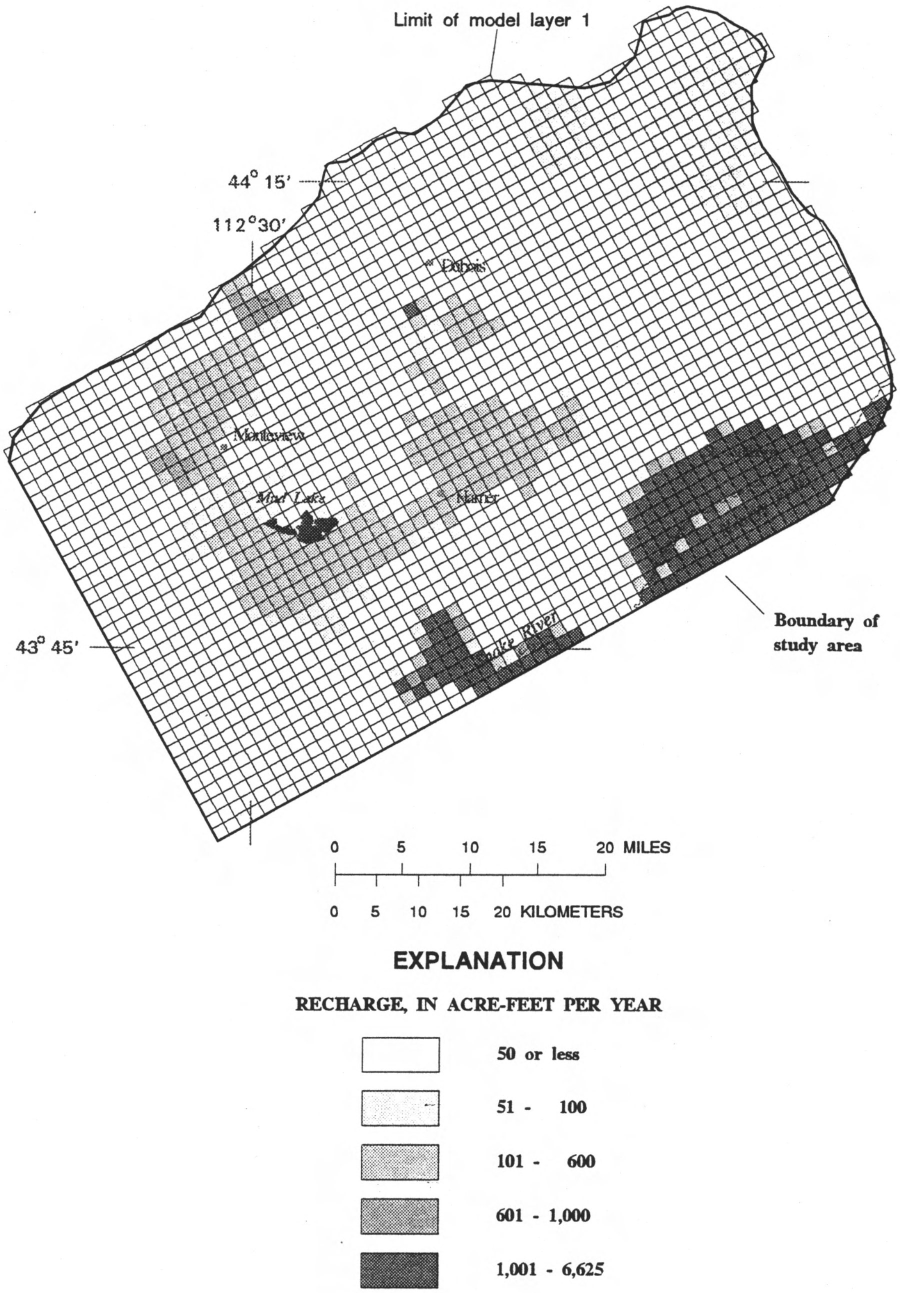


Figure 38. Distributed recharge from precipitation and irrigation, 1980.

FREE SURFACE

A free-surface boundary represented the water table and the top of model layer 1. The position of the free-surface boundary is calculated during model simulations and can move up or down in response to changes in recharge and simulated flow through head-dependent boundaries (Franke and others, 1987, p. 5).

SPECIFIED FLUX

Recharge from precipitation and irrigation, underflow from tributary basins, and withdrawals from wells were assigned to active cells in the model grid through specified-flux boundaries (Franke and others, 1987, p. 4). Recharge was specified in the model with the recharge package (McDonald and Harbaugh, 1988, chap. 7). The method described in the "Recharge" section was used to obtain a value of recharge from precipitation and irrigation for each active cell in model layer 1 for 1980 (fig. 38).

Underflow from tributary basins and ground-water withdrawals from wells were specified in the model with the well package (McDonald and Harbaugh, 1988, chap. 8). The method used to calculate underflow was described in the "Recharge" section. Underflow was allocated to cells in model layer 1 that corresponded to the location of each tributary basin (fig. 36) by division of 1980 underflow (fig. 27) by the number of cells where underflow was presumed to occur on the basis of geologic information.

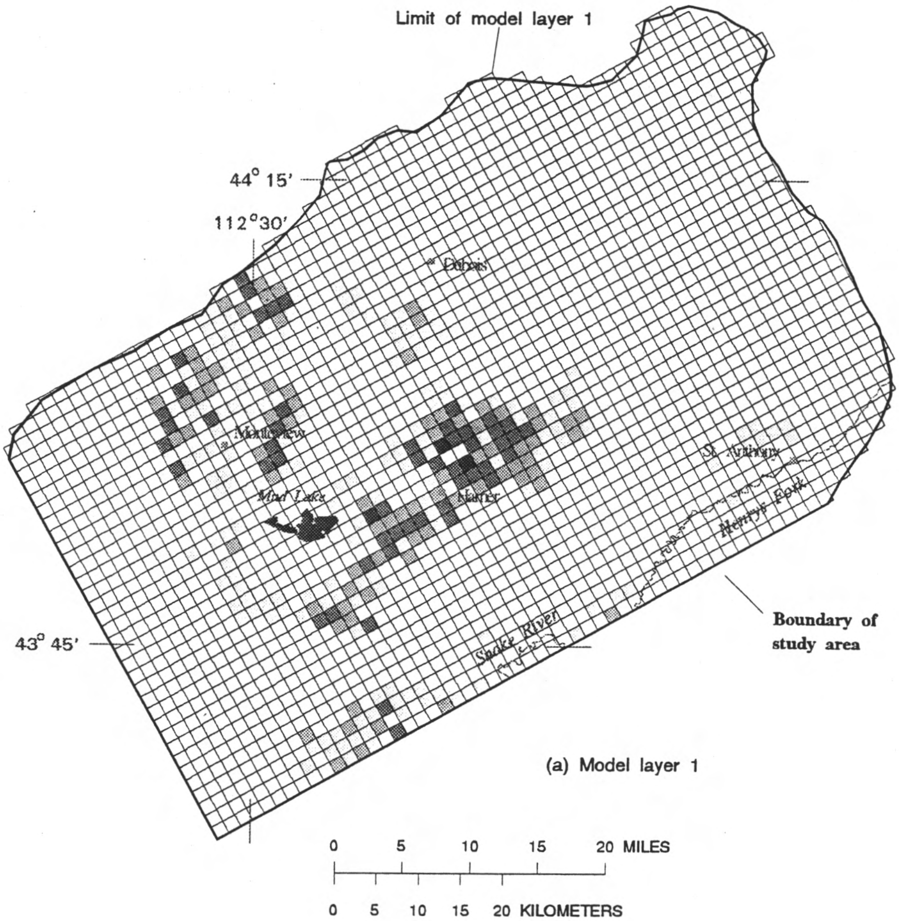
The method used to calculate withdrawals from wells for 1983–90 was described in the "Discharge" section. Withdrawals calculated for 1983 were used for 1980–82 also. Although the number of well permits issued increased during 1980–82 (fig. 28), the change was small relative to the total number of permits on file by 1983. Withdrawals for 1980 were assigned to the row and column that corresponded to well locations (fig. 29). Withdrawals were assigned to specific model layers (fig. 39) on the basis of the depth of the well below the water table and were weighted by the hydraulic conductivity assigned to the cell. For 1980 conditions, withdrawals were specified in 333 cells for layer 1; 239 cells for layers 1 and 2; 60 cells for layers 1, 2, and 3; and 9 cells for layers 1, 2, 3, and 4. No production wells are known to produce water from depths represented by model layer 5.

Model Calibration

The model was calibrated to assumed steady-state hydrologic conditions for calendar year 1980 in one annual increment. Although water-level measurements indicate that the water table rose and declined during 1980 (figs. 19, 22, and 27), levels at the beginning and end of the year were nearly identical, which indicates equilibrium between recharge and discharge with little or no change in ground-water storage.

The trial-and-error method was used to calibrate the model. In this method, intermediate steady-state simulations were produced from data sets derived independently of the model and from initial specifications of calibration variables. Simulation results were evaluated, calibration variables were adjusted within reasonable hydrologic limits, and the process was repeated until adjustment of calibration variables did not improve the correspondence between measured values and final simulation results for ground-water levels, losses and gains for streams and lakes, and discharge from flowing wells. Data derived independently of the model were not changed during calibration. Weighted-average monthly recharge from precipitation and irrigation, tributary underflow, and withdrawals from wells (described in the "Recharge" and "Discharge" sections) were aggregated to produce data sets for the 1980 steady-state simulation.

Calibration variables included fixed heads along the southwestern head-dependent flux boundary, zone values for hydraulic conductivity in model layer 1 and transmissivity in model layers 2 through 5, streambed hydraulic conductivity, altitude of the bottom of the streambed for some stream reaches, and drain conductance for flowing wells. Calibration variables were finalized in stages. First, streamflow and lake losses and gains and discharge from flowing wells were represented by specified-flux boundary conditions while values for fixed head along the southwestern model boundary and hydraulic conductivity and transmissivity were finalized. Results from initial steady-state simulations indicated that ground water along the southwestern head-dependent flux boundary flowed to the study area through some cells and from the study area through others in each model layer. The water-table map (fig. 21) indicates that ground water moves southwestward, out of the study area. Intermediate simulations indicated that fixed heads along the southwestern head-dependent flux boundary controlled the direction of water movement across that



EXPLANATION

GROUND-WATER WITHDRAWALS, IN ACRE-FEET PER YEAR

0

1 - 500

501 - 1,000

1,001 - 2,500

2,501 - 5,000

Figure 39. Distributed ground-water withdrawals for model layers 1–4, 1980.

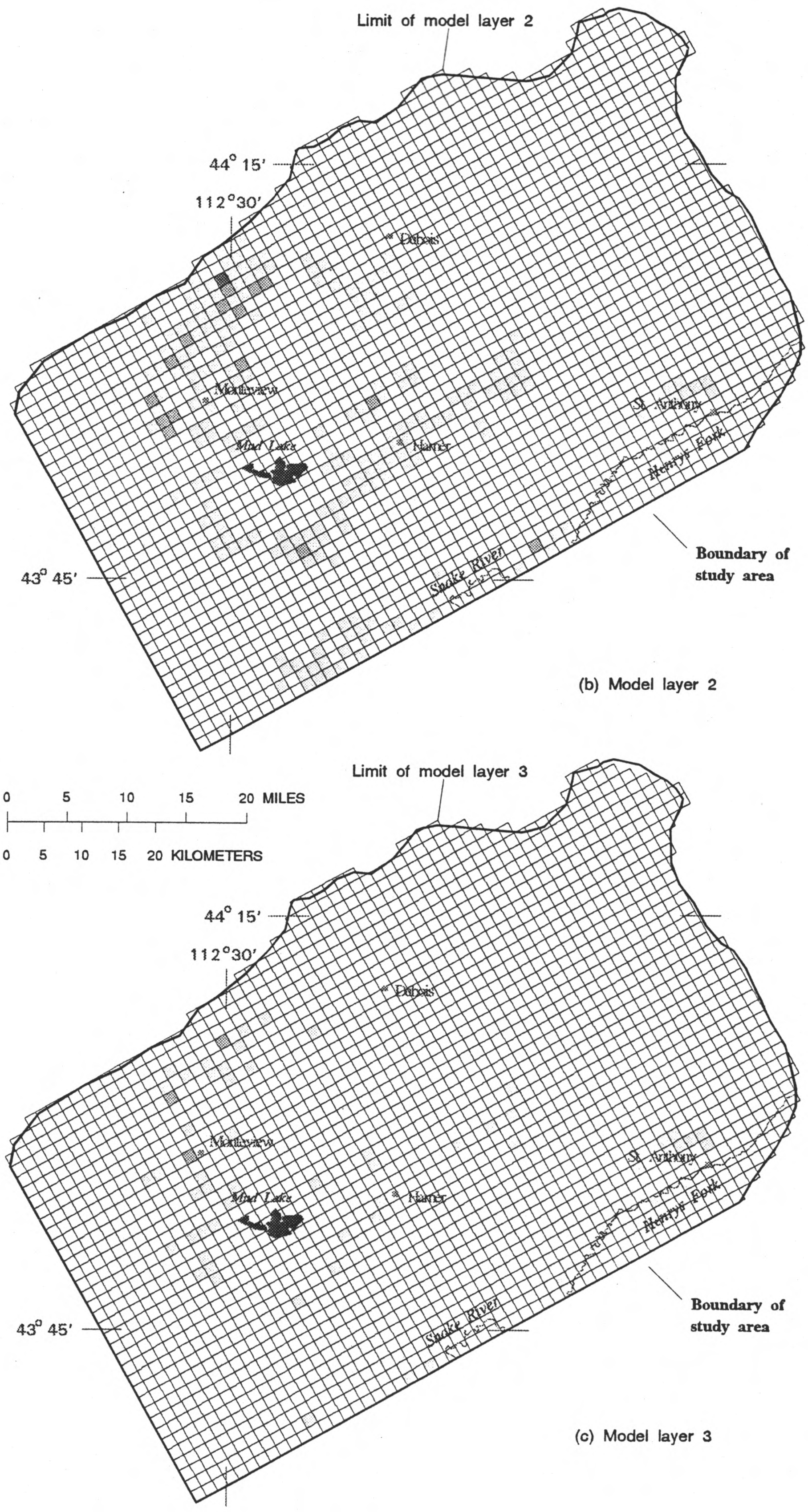


Figure 39. Distributed ground-water withdrawals for model layers 1–4, 1980—Continued.

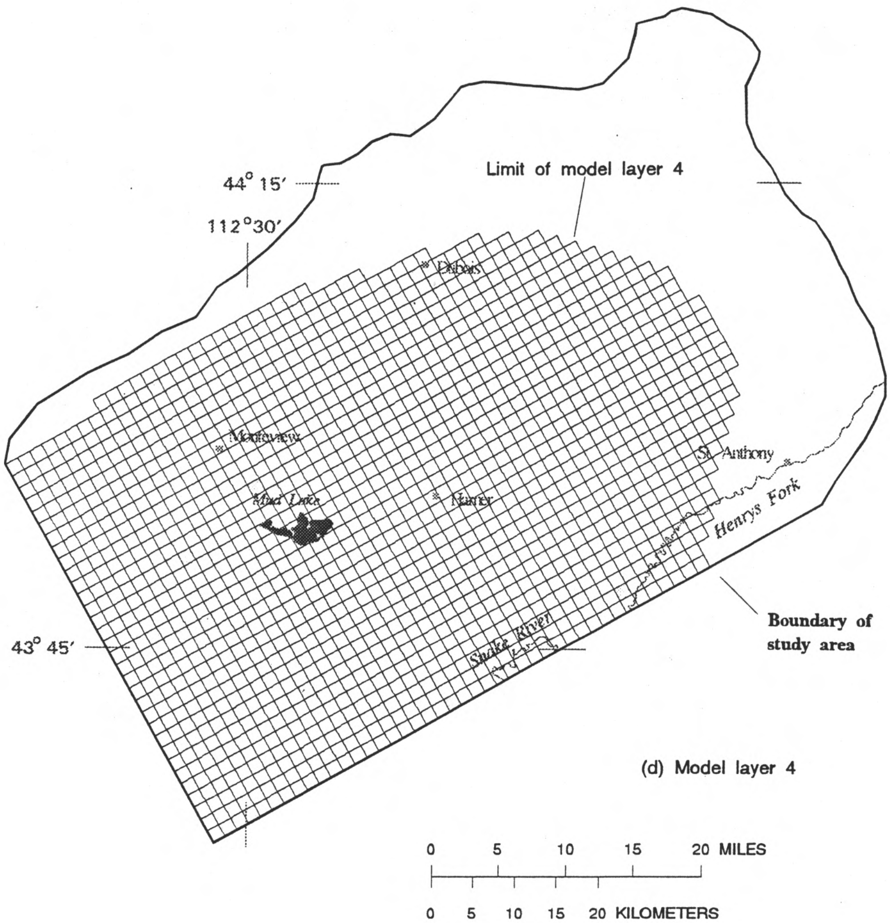


Figure 39. Distributed ground-water withdrawals for model layers 1–4, 1980—Continued.

boundary. An algorithm was developed to automatically reduce the fixed heads in each row relative to the gradient obtained from simulated heads in the two active cells adjacent to the boundary. A computer program that included this algorithm was run externally from the model each time hydraulic conductivity or transmissivity was adjusted along the southwestern boundary to ensure that fixed heads were calculated such that ground water always flowed out of the modeled area from all layers along the southwestern boundary (table 4, back of report). Finalized zone values for horizontal hydraulic conductivity ranged from 0.125 to 5,000 ft/d (fig. 37), which agree reasonably well with values cited in the section "Aquifer properties from field data." In general, lowest hydraulic conductivity values corresponded to areas in the subsurface predominated by fine-grained sediments, and highest values corresponded to areas predominated by basalt (fig. 6). Mid-range values corresponded to areas predominated by coarse-grained sediments or by mixtures of rock types.

Specified-flux boundaries used to represent stream and lake segments were removed, and the stream package of the model was activated to simulate losses from and gains to stream and lake segments after final adjustments were made to values of fixed heads along the southwestern boundary and zone values for hydraulic conductivity and transmissivity. Final values for streambed hydraulic conductivity (table 5, back of report) were obtained by multiplication of initial values for each stream or lake segment by a single constant until simulated losses and (or) gains approximated a target value for each segment (fig. 18). Altitude of the bottom of the streambed was finalized for some reaches in some stream and lake segments where the water table was near stream or lake stage (table 5, back of report). Last, drain conductance for flowing wells (table 6) was obtained by specification of the value required to simulate measured discharge of 10,000 acre-ft in 1980 (fig. 32).

An attempt was made to calibrate the model to transient conditions in monthly increments from January 1981 through December 1990. The 1981–90 period was not one of equilibrium; rather, changes in recharge and discharge resulted in changes in aquifer storage. Ground-water levels measured at the beginning of any one year did not necessarily return to the same level at the end of the year. Minimum and maximum water levels for the entire period of record occurred during 1981–90 (figs. 19, 22, and 27) and indicated that recharge and discharge also approached extremes for the period of record. Streamflow in the area ranged from extremely low to extremely high (figs. 11 and 12).

Intermediate transient simulations produced water levels that were relatively constant compared to levels measured in wells that increased during 1983–86 (figs. 19, 22, and 27). Cumulative effects of uncertainty in one or several components of recharge and (or) discharge were assumed to cause the discrepancies between measured and simulated water levels. A considerable amount of time was spent to verify that the recharge and discharge components, described in the "Recharge" and "Discharge" sections, were reasonable, but model results did not improve.

Several transient simulations were made to determine the sensitivity of simulated water levels to increased rates for individual recharge components and to decreased withdrawals from wells. For a single simulation, values of an individual recharge or discharge component were changed during 1982–86 while all other components retained their original values. Sensitivity results indicated that the best correlation between measured and simulated water levels throughout the modeled area was produced when recharge to cells that represent Egin Bench (fig. 1) and the area south of the Henrys Fork was increased from original amounts by 123,000 to 370,000 acre-ft/yr during 1982–86.

Effects of increased recharge could be realized by increased recharge from irrigation and precipitation, increased streamflow losses, or decreased streamflow gains. Recharge from precipitation and irrigation on Egin Bench, described in the "Recharge" section, was 320,176 acre-ft/yr as reported by King (1987, p. 18) and was within 10 percent of the value of 347,600 acre-ft/yr reported in an earlier study (Garabedian, 1992, p. 15). Streamflow losses and gains were more variable than recharge from precipitation and irrigation. Streamflow gains to the Henrys Fork were estimated to exceed streamflow losses by 83,600 acre-ft for water year 1977 (Wytzes, 1980, p. 29). Calculated streamflow gains for the Henrys Fork from 1982 to 1986 (fig. 18) were as much as three times greater than Wytzes' 1977 estimate. However, available data did not justify the increases in recharge or changes in calculated streamflow losses and gains that the sensitivity simulations indicated were needed. Therefore, the attempt to calibrate the model to transient conditions was discontinued.

Table 6. Drain package data for flowing wells

[Row and column numbers are shown on figure 36]

Layer	Row	Column	Fixed head (feet above sea level)	Conductance (cubic feet per feet squared per foot per day)
1	24	22	4,764	92,233
1	25	23	4,757	134,010

Calibration Results

The model was calibrated by evaluation of the correspondence between values for measured and simulated ground-water levels, for target and simulated losses and gains for stream and lake segments, and for measured and simulated discharge from flowing wells. Correspondence between measured or target and simulated values was a general indication that a suitable representation of the aquifer system had been achieved with the combination of recharge, discharge, and aquifer properties specified in the model. Water budgets provided a summary of the major items of recharge and discharge and identified the quantity of underflow simulated through general-head model boundaries.

The correspondence between measured and simulated water levels was evaluated with maps and graphs. Several considerations must be given when

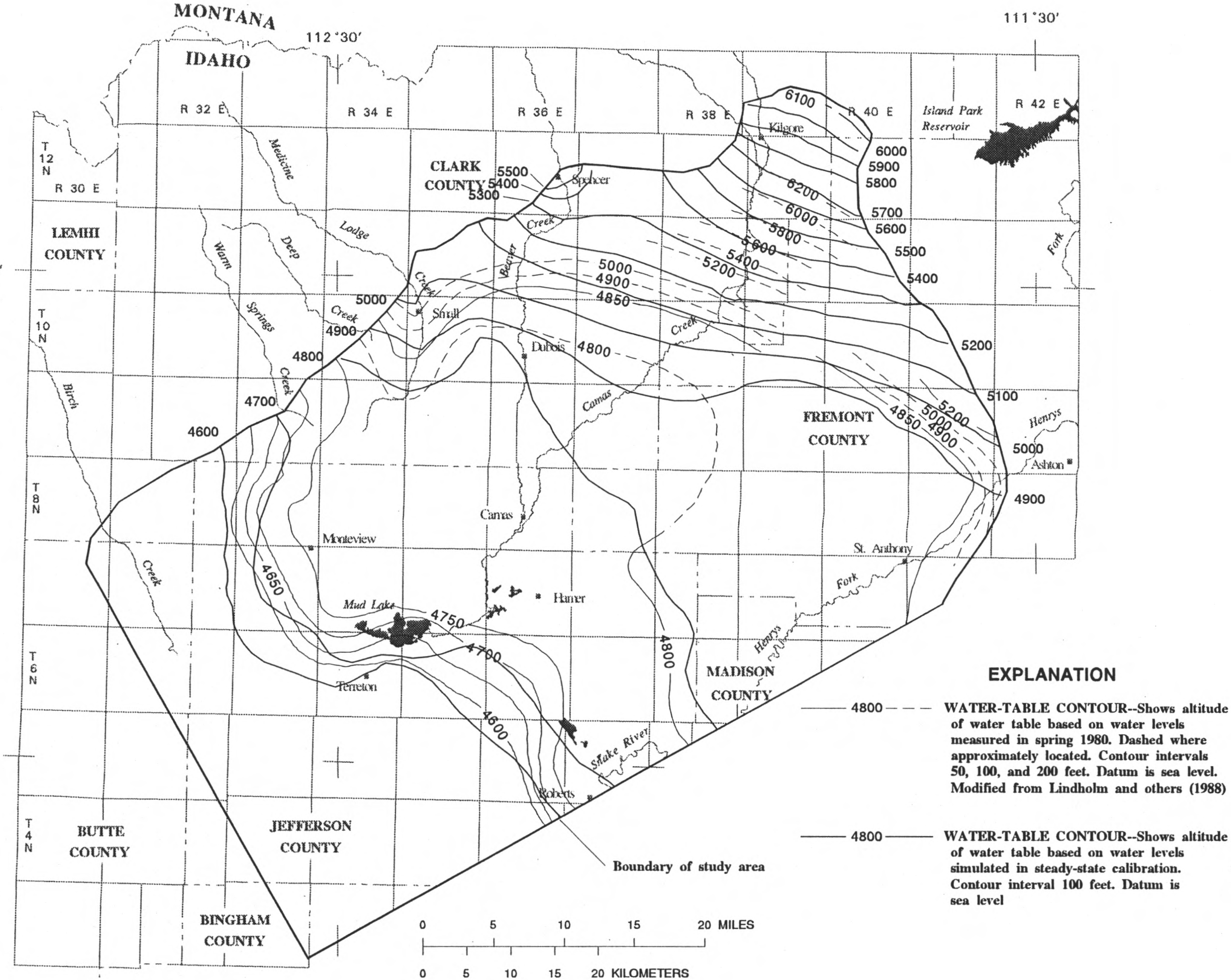


Figure 40. Measured and simulated water tables, 1980.

field-measured water levels are compared with simulated levels. Land-surface altitudes for most wells were estimated in the field from 1:24,000-scale topographic maps. Altitudes so estimated may be as much as 10 ft in error if the well is located properly on a map with a contour interval of 20 ft. If the well is mislocated, the error can be greater. Water-level measurements may not represent equilibrium conditions and could have been affected by pumping. Simulated water levels used to construct a water-table map are representative of the geometric center of each grid cell and were not adjusted to horizontal or vertical positions of measured wells. Also, the effects of pumping are dissipated throughout the volume of a grid cell; therefore, simulated water levels are muted when compared to pumping levels measured in wells. Additional considerations are noted in the following discussion.

Evaluation criteria for comparison of water-table maps generated from measured and simulated water levels included the configuration of the water table and positions of the 4,600-, 4,700-, and 4,800-ft contours. Calibration efforts were directed to obtain correspondence in the area between the 4,600- and 4,800-ft contours because (1) most of the current and potential ground-water use is in that area, and (2) data outside of that area are sparse. The water-table map based on 1980 measured water levels (Lindholm and others, 1988) was compared with the map based on final steady-state calibration simulated water levels for model layer 1 (fig. 40). Similarities between the two maps include the steep hydraulic gradient between the S-shaped bends in the 4,600- and 4,700-ft contours, the width and shape of the low-gradient area between the 4,700- and 4,900-ft contours, and the steep gradient where the water table exceeds 4,900 ft. Water-level measurements greater than 4,900 ft are sparse, and correspondence in the steep gradient area where the water table exceeds 4,900 ft may be better than shown. Discrepancies between measured and simulated 4,800-ft water-level contours might be because (1) location of the 4,800-ft contour based on measured water levels is inferred from sparse data, and (2) the water-table map based on measured water levels depicts horizontal flow in the aquifer system and does not include water levels in shallow wells.

Table 7. Target and simulated losses from and gains to stream and lake segments

[Losses and gains are reported in acre-feet per year to two significant figures for calendar year 1980; segments are identified on figure 9; —, indicates no data]

Segment	Target loss	Target gain	Simulated loss	Simulated gain
Medicine Lodge Creek	45,000	—	45,000	—
Beaver Creek	17,000	—	17,000	—
Camas Creek:				
Upper segment	9,300	—	9,300	—
Middle segment	1,500	—	1,400	—
Rays Lake segment	13,000	—	13,000	—
Mud Lake segment	20	—	17	2
Mud Lake	11,000	—	11,000	—
Wood's diversion	8,400	—	8,400	—
Henrys Fork	—	39,000	7,400	47,000
Snake River	—	68,000	1,400	69,000
Camas National Wildlife Refuge	8,400	—	8,400	160
Market Lake Wildlife Management Area	—	1,000	3	1,100
TOTAL	110,000	110,000	120,000	120,000

Table 8. Water budget, calendar year 1980

[All values reported in acre-feet to two significant figures; —, indicates no data; values for stream and lake losses and gains, discharge to flowing wells, and underflow across model boundaries were simulated with the model; values for other items were derived independently of the model]

Budget item	In	Out
Recharge from precipitation and irrigation:		
Egin Bench	330,000	—
Areas south of Henrys Fork	140,000	—
Remainder of study area	190,000	—
Stream and lake losses	120,000	—
Underflow from tributary basins	450,000	—
Stream and lake gains	—	120,000
Withdrawals from wells	—	240,000
Discharge to flowing wells	—	10,000
Underflow across model boundaries:		
Southeastern boundary	49,000	14,000
Southwestern boundary:		
Layer 1	—	150,000
Layer 2	—	140,000
Layer 3	—	350,000
Layer 4	—	220,000
Layer 5	—	58,000
TOTAL	1,300,000	1,300,000

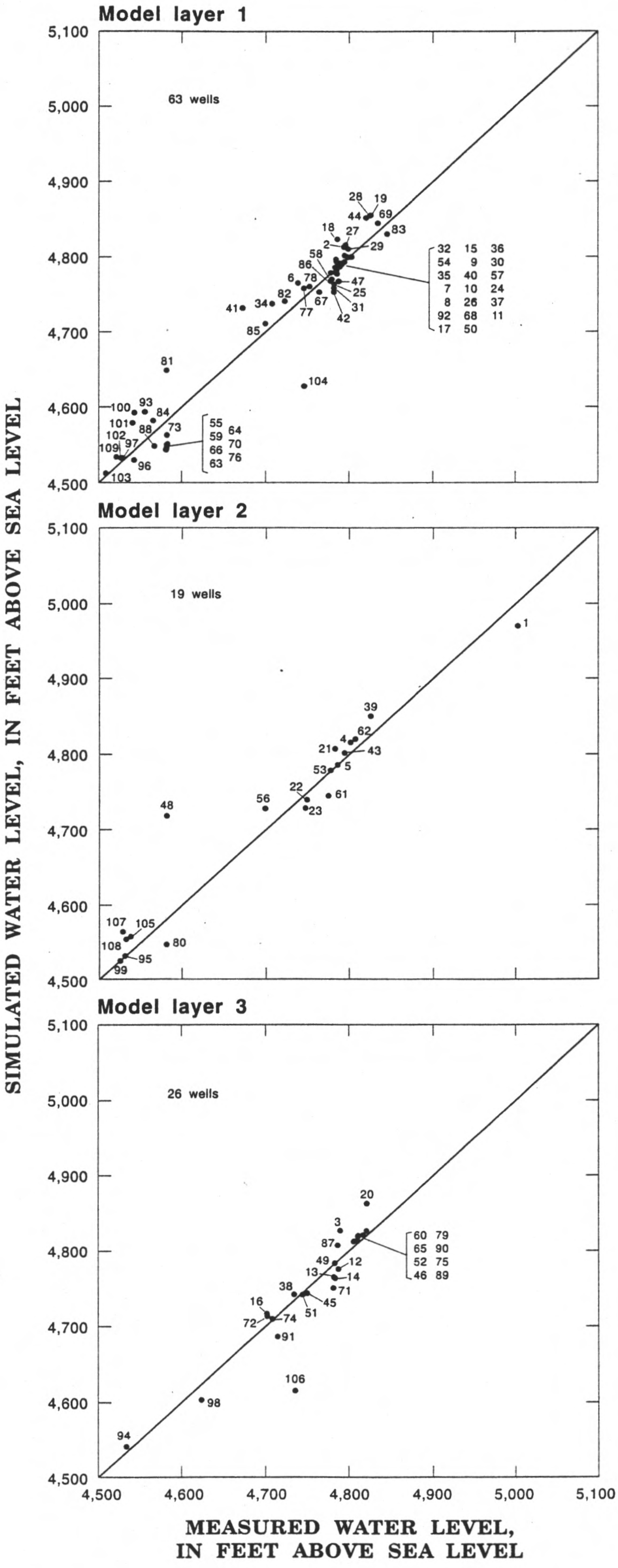


Figure 41. Measured and simulated water levels, 1980. (Well locations shown on figure 21)

Most measured and simulated water levels plot along a line of equality (fig. 41). Index numbers (fig. 41) were assigned to well locations from north to south (fig. 21). Differences are notable for wells 104 (model layer 1), 48 (model layer 2) , and 106 (model layer 3), which are located in areas where the water-table gradient is steep (fig. 40). Although the model simulated a steep gradient in these areas, simulated water levels did not always correlate with their measured counterparts. Simulated levels for wells 86 (model layer 1), 22 (model layer 2), 74 (model layer 3), and , also in the steep water-table area, correlated well with measured levels. Besides reasons cited earlier, differences between measured and simulated water levels could be a result of the model layer that was assigned to each measured level. Wells measured for the 1980 water-table map (fig. 21) were open to depths that corresponded to model layers 1, 2, 3, or a combination of these layers. Each measured water level used to construct the map was assigned to a single model layer that corresponded to the greatest depth of the open or screened interval as reported on drillers' logs on file with the IDWR. If a measured water level assigned to model layers 2 or 3 were open to an interval greater than the thickness represented by a single model layer, the measured level would represent a composite value and could not be directly compared with a water level simulated for a single layer. Drillers' logs on file with the IDWR indicate that wells 48 and 106 are open to an interval greater than the thickness of the layers to which each well was assigned. Simulated water levels were interpolated to well locations with GIS techniques, and errors associated with inter-polated levels are unknown.

Differences in simulated water levels between adjacent model layers indicate areas of vertical flow and the magnitude of water-level differences (fig. 42). Simulation results indicate downward flow in recharge areas along the margin of the plain, below Egin Bench (fig. 1), and in the area between the 4,600- and 4,700-ft water-table contours south of Mud Lake (fig. 21). Upward flow was simulated where interbeds of basalt and sediments occur in the subsurface around Mud Lake (fig. 7).

Simulated losses from and gains to stream and lake segments closely approximated target values (table 7). Target values for stream and lake segments were obtained independently of the model from measurements and estimates as described in the "Losses and gains" section. The 10,000-acre-ft difference between total target and simulated losses and gains is due largely to simulated losses to Henrys Fork and the Snake River. Target values include the sum of losses and gains in one net value. Net simulated losses and gains were within 2 percent of target values for stream or lake segments where differences between the two exceeded 100 acre-ft. Simulated discharge from flowing wells in 1980 (table 8) equaled the measured value of 10,000 acre-ft (fig. 32).

A water budget was developed for the study area for 1980 (table 8). Recharge from precipitation and irrigation (660,000 acre-ft) was the largest inflow item, and total simulated underflow across model boundaries (932,000 acre-ft) was the largest outflow item. Examination of budget values shows that recharge on Egin Bench alone accounted for half of the recharge from precipitation and irrigation; withdrawals from wells (240,000 acre-ft) were slightly greater than half the underflow from tributary basins (450,000 acre-ft); and stream and lake losses were balanced by gains in 1980. The accuracy of individual budget items derived independently of the model was unknown because all items included estimates with unknown degrees of uncertainty. Although uncertainty associated with any one budget item could be large, that uncertainty was compensated by uncertainty in other budget items as indicated by the equality of total inflow and outflow (1,300,000 acre-ft). Simulated underflow across the southwestern model boundary of 918,000 acre-ft/yr was within 0.1 order of magnitude of the 1,070,000 acre-ft/yr simulated through that boundary for 1980 by Garabedian's model of the eastern Snake River Plain aquifer system (D.J. Ackerman, USGS, written commun., 1992). Simulated underflow to the study area across the southeastern model boundary of 49,000 acre-ft for 1980 is plausible in that recharge to areas adjacent to this boundary is larger and ranges from 82,000 to 1,000,000 acre-ft/yr (Garabedian, 1992, p. 15).

Sensitivity analysis was used to assess the relative degree of uncertainty in selected budget items on results from the calibration simulation. In sensitivity analysis, one selected variable was increased or decreased while all other data sets retained their original values, a steady-state simulation was made, and model results were evaluated. Sensitivity analysis was done on data sets for recharge from precipitation and irrigation over the entire model grid, tributary underflow, and withdrawals from wells. Recharge over the entire model grid and tributary underflow were increased by 25 percent, and withdrawals from wells were reduced by 50 percent. The magnitude of these changes was considered to be greater than the uncertainty associated with the budget item. All sensitivity simulations produced higher water levels and increased underflow compared to the calibration simulation. Results from the sensitivity analysis indicated that increased recharge over the entire model grid had the greatest effect on model results, followed by decreased withdrawals from wells and increased tributary underflow. Therefore, errors in recharge would have the greatest effect on model results, and errors in tributary underflow would have the least.

Model Limitations

Assumptions and simplifications made during the development of data sets and in the design and calibration of the numerical model affect solutions obtained with the model. Data sets developed for recharge, discharge, and aquifer properties were assumed to represent the distribution and magnitude of actual field conditions in 1980. This assumption is premised on the condition that these data were measured or estimated with an acceptable degree of accuracy. Model calibration and the results of simulations made with the calibrated model could be improved with better estimates of streamflow, streamflow diversions, irrigation return flows, and withdrawals from wells, supported by regular measurements of each at more locations than are made at present (1993). More accurate estimates of aquifer properties could be obtained from additional aquifer tests distributed regularly throughout the area and designed to test specific depth intervals of the aquifer system. Measurements of vertical hydraulic conductivities of streambeds and lakebeds are needed to better represent interaction between streams, lakes, and the aquifer system.

The dimensions of the model grid were chosen to represent geohydrologic conditions in the aquifer system at a scale commensurate with the level of detail of available information to construct data sets for recharge, discharge, and aquifer properties. Grid dimensions restrict the simulation of water levels to one value per cell and affect interpretation of model results in areas where withdrawals from wells and vertical flow occur. Differences between measured and simulated ground-water levels should be expected due to withdrawals from wells. Although water-level declines due to withdrawals might be large in a single well, the simulated water level in the cell that corresponds with the location of that well would be

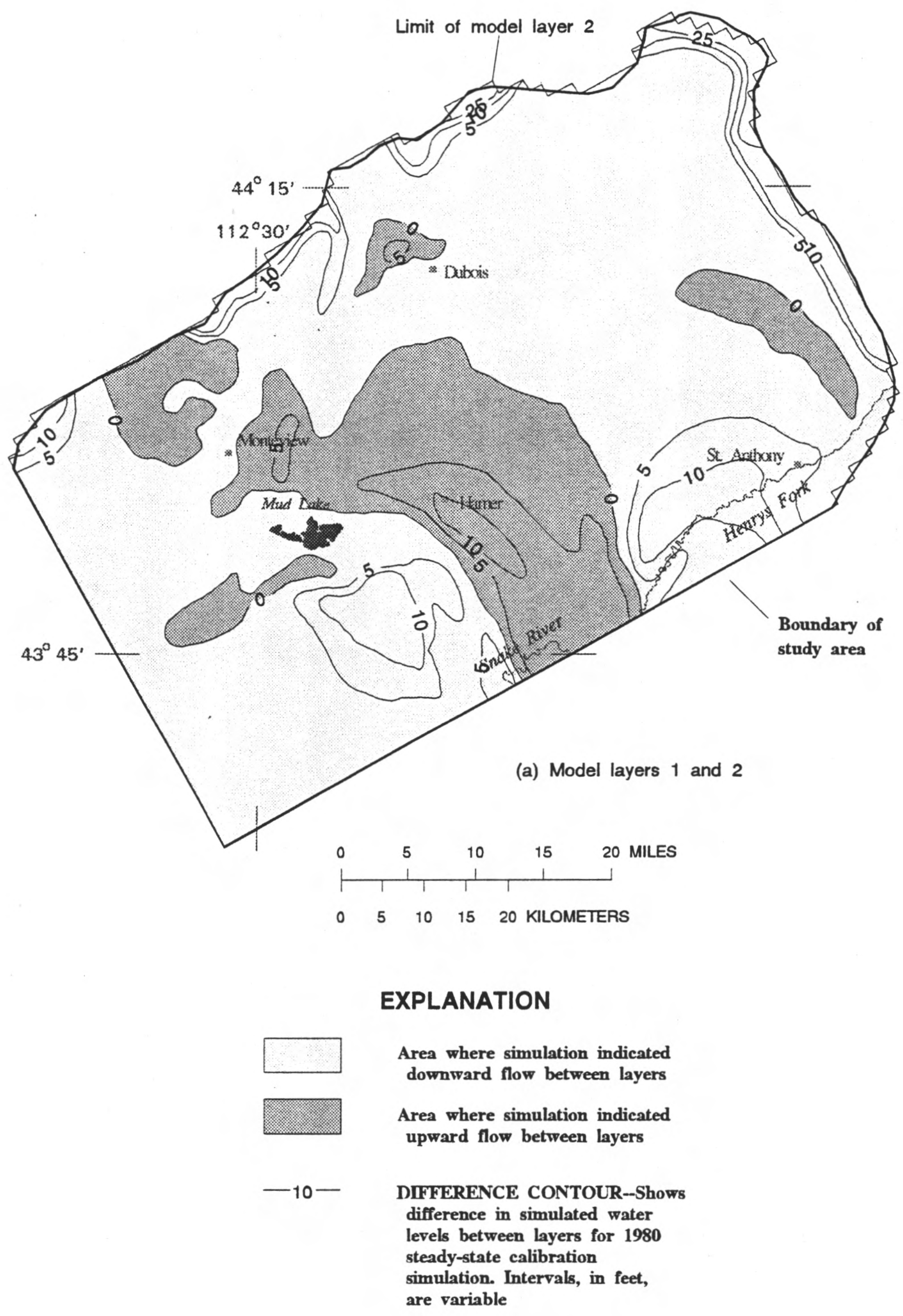


Figure 42. Simulated water-level differences between adjacent model layers.

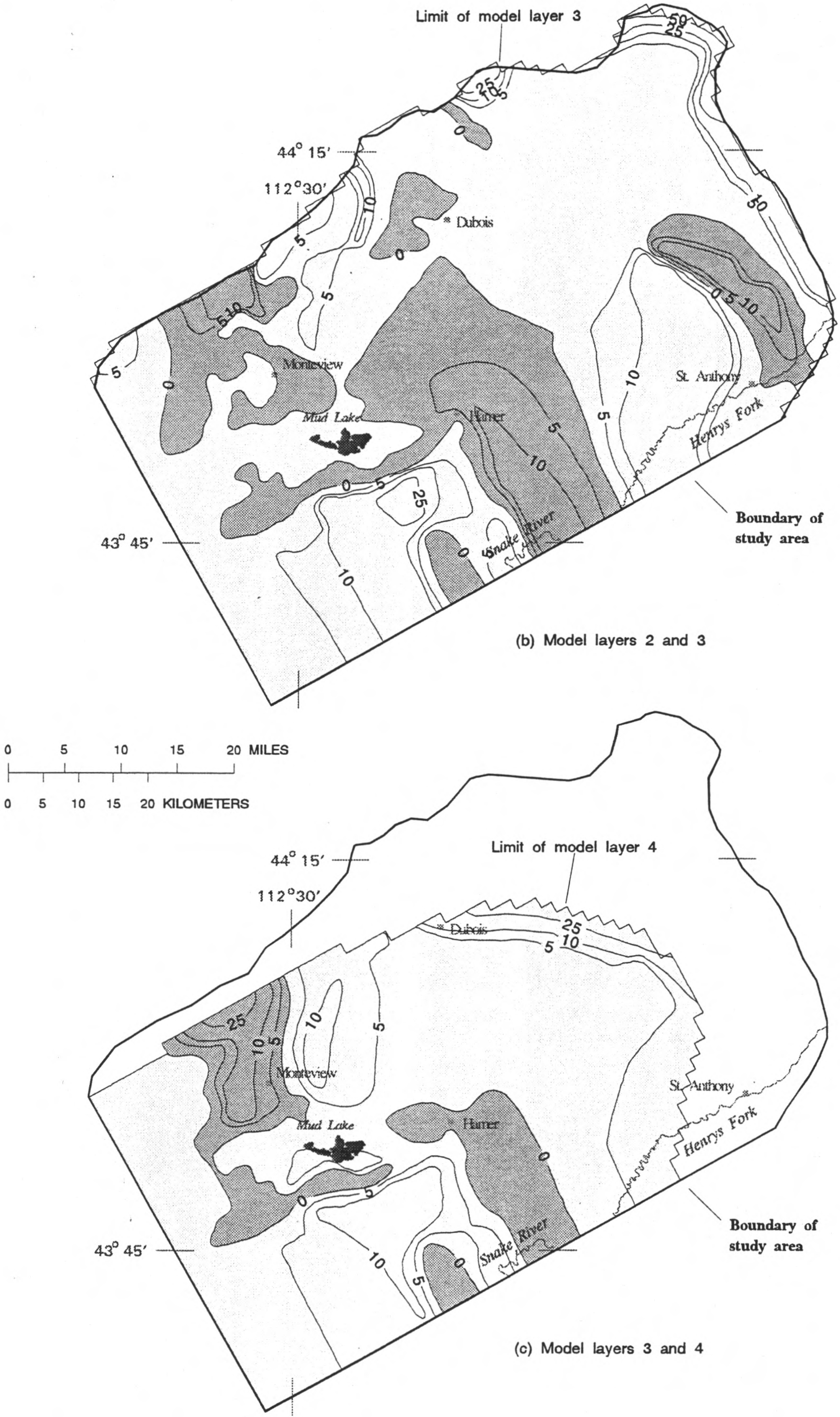


Figure 42. Simulated water-level differences between adjacent model layers—Continued.

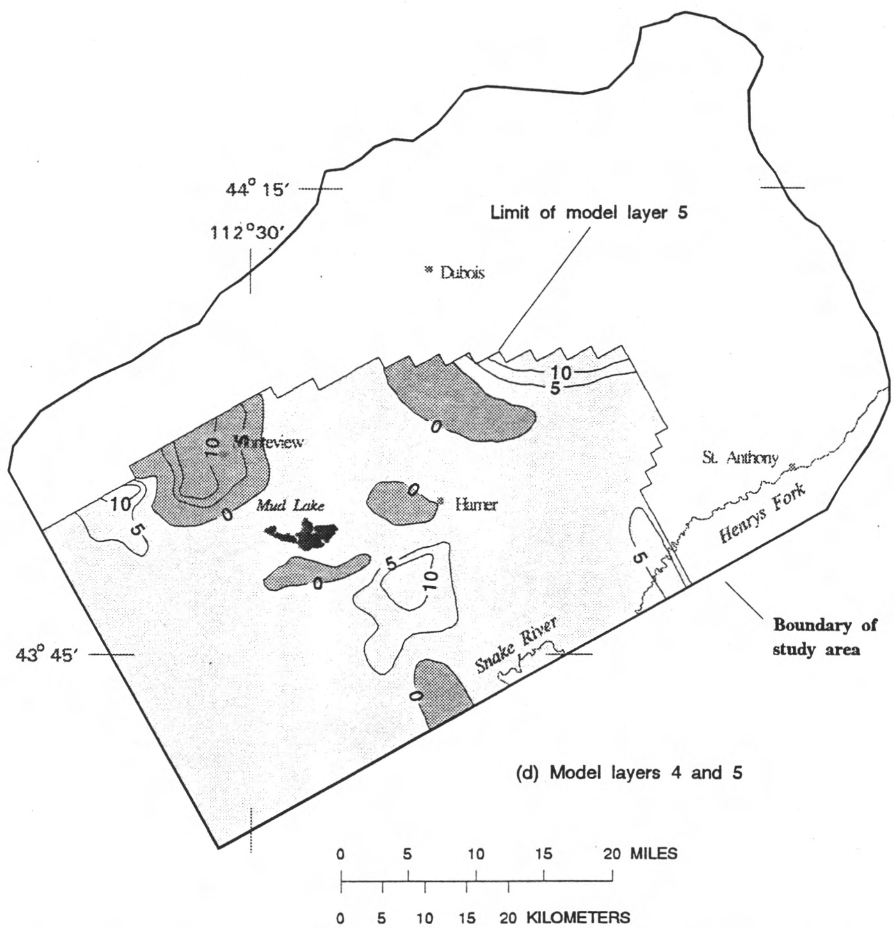


Figure 42. Simulated water-level differences between adjacent model layers—Continued.

smaller because the withdrawal is distributed throughout the cell. A similar result could be expected in areas of vertical flow. Although measurements might indicate that water level changes with depth through an interval that corresponds with the thickness of a cell, only a single water level would be simulated for the cell.

The model simulated water levels, stream and lake losses and gains, and underflow to adjacent areas of the Snake River Plain aquifer system that were comparable with measurements, estimates, and independent results for these values for 1980. Model results need to be interpreted with caution if model inputs are changed to simulate effects for different combinations of recharge and (or) discharge. Simulated results may not be plausible if recharge or discharge is changed beyond the range within which the model was calibrated. Simulated underflow across general-head boundaries should be inspected to ensure that outflow is always simulated through the southwestern boundary and that the amount of underflow simulated into the model area across the southeastern general-head boundary is reasonable relative to recharge outside of the study area.

The specification of vertical conductance between model layers was greatly simplified. Improved estimates of vertical conductance obtained from aquifer tests need to be correlated with rock type to better understand and to enable the model to better represent field conditions.

SUMMARY

Water users in the Mud Lake study area in the northernmost part of the eastern Snake River Plain depend on an adequate supply of ground water for agriculture, wildlife, and other uses. Changes in water use have raised concerns about an adequate future supply of surface and ground water. Water managers need the ability to evaluate the consequences of increased ground-water development throughout the area and 95,000 acre-ft less recharge on Egin Bench on future water levels and water supply. The geohydrology was described and a three-dimensional, finite-difference, numerical ground-water flow model of the aquifer system was calibrated to obtain a better understanding of the geohydrology and provide a tool to evaluate water-use alternatives. Geohydrologic descriptions include surficial and subsurface geology; surface-water supply and use; ground-water occurrence, recharge, and discharge; and aquifer properties. The numerical model is a computer program that generates a distribution of water levels and simulates ground-water flow from a mathematical synthesis of data sets and boundary conditions developed from geohydrologic data.

Precipitation on areas within and adjacent to the study area determines the supply of surface and ground water in the study area. Annual precipitation on the plain (1930–57) ranged from 8 in. to about 35 in. Annual precipitation measured at Dubois from 1980 to 1990 ranged from 9.39 in. during 1988 to 20.6 in. during 1983 and averaged 13.9 in.

The aquifer system that underlies the eastern Snake River Plain is composed of basalt, rhyolite, coarse-grained sediments (sand and gravel) and fine-grained sediments (clay, silt, and sand). Layers of basalt predominate on and under the plain. Total basalt thickness is less than 4,000 ft. Layers of coarse-grained sediments underlie channels of the Henrys Fork and Snake River and are present in alluvial fans that extend southward from the northwestern margin of the plain. Layers of fine-grained sediments are present in lakebeds that underlie the area around Mud Lake. Total sediment thickness is less than 1,000 ft. Basalt and sediment interbeds affect ground-water movement and supply locally.

Medicine Lodge, Beaver, and Camas Creeks, Wood's diversion, Mud Lake, Henrys Fork, Snake River, and lakes on the Camas National Wildlife Refuge and Market Lake WMA are hydraulically connected with the aquifer system. Streamflow, ground-water inflow, diversions, lake ET, and losses from and gains to streams and lakes were quantified monthly from January 1980 through December 1990.

Ground water is both unconfined and confined in the study area. Generally, water nearest land surface is unconfined, and the water table defines the top of the aquifer. Water-table altitudes range from about 4,500 ft above sea level near the southwestern corner of the study area to about 6,200 ft in the northeastern part. Water-table gradients are about 30 ft/mi between the 4,600- and 4,700-ft water-table contours, about 3 ft/mi between the 4,700- and 4,800-ft contours, and about 120 ft/mi where the water table is higher than 4,900 ft. The area between the 4,600- and 4,700-ft contours coincides with a band of sediments that extends into the subsurface. The water table closely resembles the configuration of land surface at altitudes greater than 4,700 ft. Confined conditions are associated with basalt and sediment interbeds in the area around Mud Lake. Water-level measurements indicate downward water movement near the margin of the plain, below Egin Bench, and southwest of Mud Lake, and upward movement in parts of Camas Creek, in Mud Lake, Camas National Wildlife Refuge, and Market Lake WMA.

Recharge to the aquifer system includes infiltration of precipitation and irrigation water in excess of consumptive use by plants, underflow from tributary drainage basins and from the eastern Snake River Plain aquifer system across part of the southeastern boundary of the study area, and stream and lake losses. Discharge from the aquifer system includes underflow across the southwestern and part of the southeastern boundary of the study area to the eastern Snake River Plain aquifer system, stream and lake gains, withdrawals from wells, and flowing wells. Recharge and discharge for most sources were estimated monthly from January 1980 through December 1990. Underflow between the study area and the eastern Snake River Plain aquifer system was calculated as a residual by the numerical model.

Reported transmissivity estimated from aquifer tests on three wells completed in basalt ranged from 480,000 to 2,500,000 ft^2/d; hydraulic conductivity ranged from 1,600 to 22,000 ft/d; and storage coefficient and specific yield ranged from 0.0008 to 0.19. Estimated transmissivity from an aquifer test of two wells completed in sand, gravel, and clay made during this study was 5,300 and 6,300 ft^2/d; hydraulic conductivity was 140 and 330 ft/d; vertical hydraulic conductivity was 1.5 and 4 ft/d; and specific yield was 0.12 and 0.17. Drillers' logs for 73 wells indicate that discharge ranges from 20 to 9,000 gal/min, pumping drawdown from 0.3 to 137 ft, and specific capacity from 3 to 4,490 (gal/min)/ft. Comparisons among values of well discharge, pumping drawdown, and specific capacity indicate that wells completed in sediments yield less water than wells completed in basalt. Median transmissivity estimated from specific-capacity and other data ranged from 43,000 to 200,000 ft^2/d; median hydraulic conductivity estimated from specific-capacity and other data ranged from 780 to 1,500 ft/d. Median transmissivity and hydraulic conductivity estimated from wells completed in basalt were greater than those from wells completed in sediments.

The numerical model is delineated by a grid of cells 40 rows long, 64 columns wide, and 5 layers deep. Cells along rows and columns are 1 mi on a side. Cells in layers 1 and 2 represent a thickness of 100 ft each; cells in layer 3 represent a thickness of 300 ft; and cells in layers 4 and 5 represent thicknesses of 500 ft or less and 1,000 ft or less, respectively. Cells represent a volume of the aquifer system and were assigned values for aquifer properties, boundary conditions, recharge, and discharge.

Hydraulic conductivity values, finalized during model calibration, ranged from 0.125 to 5,000 ft/d. Vertical conductances between adjacent layers, in descending order, were 0.01, 0.005, 0.0025, and 0.0013 (ft/d)/ft. No-flow boundaries were specified to represent the natural extent of the aquifer system and a flowline along part of the southeastern boundary of the study area. Head-dependent flux boundaries were specified along the southwestern boundary of the model grid in layers 1–5 and along the southeastern boundary for some cells in layer 1 to simulate underflow into and out of the eastern Snake River Plain aquifer system adjacent to the study area. Head-dependent boundaries were used to simulate stream and lake losses and gains and discharge from flowing wells. A free-surface boundary represented the water table and the top of layer 1. Recharge from precipitation and irrigation, tributary underflow, and withdrawals from wells were assigned through specified-flux boundaries.

The model was calibrated with the trial-and-error method to steady-state hydrologic conditions for calendar year 1980. A water-table map based on 1980 water-level measurements was compared with a map based on steady-state simulated water levels in model layer 1. Measured and simulated water-table maps show a low hydraulic gradient for much of the area between the 4,700- and 4,900-ft contours, a steeper gradient for the area between the 4,600- and 4,700-ft contours, and the steepest gradient for the area where water-table contours exceed 4,900 ft. Downward flow was simulated between layers along the margin of the plain, below Egin Bench, and south of Mud Lake. Upward flow was simulated where interbeds of basalt and sediments are present in the subsurface around Mud Lake. Simulated losses and gains closely approximated those obtained independently of the model. Net simulated losses or gains were within 2 percent of target values where differences between target and simulated values exceeded 100 acre-ft. Simulated discharge from flowing wells matched the measured discharge of 10,000 acre-ft in 1980. A water budget developed for calendar year 1980 indicated a balance between inflow and outflow; a total of 1,300,000 acre-ft of water moved through the aquifer system. Recharge from precipitation and irrigation, the largest inflow item, was 660,000 acre-ft; total underflow, the largest outflow item, was 932,000 acre-ft. Recharge to Egin Bench alone accounted for half of the recharge from precipitation and irrigation.

An attempt was made to calibrate the numerical model to transient conditions in monthly increments for 1981–90. However, because of the cumulative effects of uncertainty in one or several components of recharge and (or) discharge, significant discrepancies resulted between measured and simulated water levels. Sensitivity analysis indicated that the best correlation between measured and simulated water levels results when recharge to Egin Bench and areas south of the Henrys Fork was increased by 123,000 to 370,000 acre-ft/yr for 1982–86. However, efforts at transient calibration were discontinued because available data did not justify the changes that sensitivity analysis indicated were needed.

The ability of the calibrated model to reproduce measured water levels and simulate ground-water flow is related directly to the accuracy of available geohydrologic data. Model calibration and results could be improved with regular measurements of streamflow on all perennial streams, measured water levels in an expanded network of observation wells, and better estimates of streamflow diversions, irrigation return flows, and withdrawals from wells. Aquifer tests and measurements are needed to obtain better estimates of aquifer properties and stream and lakebed vertical hydraulic conductivities. Improved estimates of vertical conductance are needed to better understand and represent field conditions.

REFERENCES CITED

Ackerman, D.J., 1991, Transmissivity of the Snake River Plain aquifer at the Idaho National Engineering Laboratory, Idaho: U.S. Geological Survey Water-Resources Investigations Report 91–4058, 35 p.

Allen, R.G., and Brockway, C.E., 1983, Estimating consumptive irrigation requirements for crops in Idaho: Moscow, University of Idaho, Idaho Water and Energy Resources Research Institute, Research Technical Completion Report, 183 p.

Anderson, S.R., 1991, Stratigraphy of the unsaturated zone and uppermost part of the Snake River Plain aquifer at the Idaho Chemical Processing Plant and Test Reactors Area, Idaho National Engineering Laboratory, Idaho: U.S. Geological Survey Water-Resources Investigations Report 91–4010 (DOE/ID–22095), 71 p.

Barraclough, J.T., Teasdale, W.E., Robertson, J.B., and Jensen, R.G., 1967, Hydrology of the National Reactor Testing Station, Idaho, 1966: U.S. Geological Survey Open-File Report IDO–22049, 95 p.

Bigelow, B.B., Goodell, S.A., and Newton, G.D., 1987 [1986], Water withdrawn for irrigation in 1980 on the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Hydrologic Investigations Atlas HA–690, 2 sheets, scales 1:500,000 and 1:1,000,000.

Brockway, C.E., and Robison, C.W., 1988, Water resources data and management model for Beaver Creek, Camas Creek and Mud Lake area of eastern Idaho: Moscow, University of Idaho, Idaho Water Resources Research Institute [variously paged].

Cecil, L.D., Orr, B.R., Norton, Teddy, and Anderson, S.R., 1991, Formation of perched ground-water zones and concentrations of selected chemical constituents in water, Idaho National Engineering Laboratory, Idaho, 1986–88: U.S. Geological Survey Water-Resources Investigations Report 91–4166, 53 p.

deSonneville, J.L.J., 1974, Development of a digital ground-water model with application to aquifers in Idaho: Moscow, University of Idaho, unpublished Ph.D. thesis, 228 p.

Franke, O.L., Reilly, T.E., and Bennett, G.D., 1987, Definition of boundary and initial conditions in the analysis of saturated ground-water flow systems—an introduction: U.S. Geological Survey Techniques of Water-Resources Investigations, book 3, chap. B5, 15 p.

Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, N.J., Prentice-Hall, 604 p.

Garabedian, S.P., 1989, Hydrology and digital simulation of the regional aquifer system, eastern Snake River Plain, Idaho: U.S. Geological Survey Open-File Report 87–237, 151 p.

——— 1992, Hydrology and digital simulation of the regional aquifer system, eastern Snake River Plain, Idaho: U.S. Geological Survey Professional Paper 1408–F, 102 p., 10 pls.

Goodell, S.A., 1988, Water use on the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Professional Paper 1408–E, 51 p.

Idaho Agricultural Statistics Service, annual reports from 1980 to 1990, Idaho agricultural statistics: Boise, Idaho Department of Agriculture [variously paged].

Johnson, G.S., Brockway, C.E., and Luttrell, S.P., 1984, Application of a numerical ground-water flow model to the Mud Lake area in southeastern Idaho: Moscow, University of Idaho, Idaho Water Resources Research Institute Technical Completion Report, Contract No. 14–08–0001–A–0016, 60 p.

King, Verl, 1987, Egin Bench sprinkler irrigation impact study report: Boise, Idaho Department of Water Resources, 23 p.

Kjelstrom, L.C., in press, Streamflow gains and losses in the Snake River and ground-water budgets for the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Professional Paper 1408–C.

Lindholm, G.F., Garabedian, S.P., Newton, G.D., and Whitehead, R.L., 1988 [1987], Configuration of the water table and depth to water, spring 1980, water-level fluctuations, and water movement in the Snake River Plain regional aquifer system, Idaho and eastern Oregon: U.S. Geological Survey Hydrologic Investigations Atlas HA–703, 1 sheet, scale 1:500,000.

Lohman, S.W., 1972, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Lohman, S.W., and others, 1972, Definitions of selected ground-water terms—revisions and conceptual refinements: U.S. Geological Survey Water-Supply Paper 1988, 21 p.

Luttrell, S.P., 1982, Ground-water flow characteristics in the Mud Lake area, southeastern Idaho: Moscow, University of Idaho, unpublished M.S. thesis, 69 p.

Mann, L.J., 1986, Hydraulic properties of rock units and chemical quality of water for INEL-1—a 10,365-foot deep test hole drilled at the Idaho National Engineering Laboratory, Idaho: U.S. Geological Survey Water-Resources Investigations Report 86–4020, 23 p.

Mantei, C.L., 1974, Snake Plain aquifer electric analog studies: Denver, Colo., U.S. Bureau of Reclamation, third progress report [variously paged].

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap. A1 [variously paged].

Mundorff, M.J., Crosthwaite, E.G., and Kilburn, Chabot, 1964, Ground water for irrigation in the Snake River Basin in Idaho: U.S. Geological Survey Water-Supply Paper 1654, 224 p., 6 pls.

National Climatic Data Center, monthly reports from 1980 to 1990, Climatological data, Idaho: Asheville, N.C., Department of Commerce, National Oceanic and Atmospheric Administration, National Environmental Satellite, Data and Information Service, v. 83, no. 1 through v. 93, no. 12 [variously paged].

Neuman, S.P., 1975, Analysis of pumping test data from anisotropic unconfined aquifers considering delayed gravity response: Water Resources Research, v. 11, no. 2, p. 326–342.

Newton, G.D., 1978, Application of a simulation model to the Snake Plain aquifer: Moscow, University of Idaho, unpublished M.S. thesis, 82 p.

Norvitch, R.F., Thomas, C.A., and Madison, R.J., 1969, Artificial recharge to the Snake Plain aquifer—an evaluation of potential and effect: Idaho Department of Reclamation Water Information Bulletin 12, 59 p.

Prudic, D.E., 1989, Documentation of a computer program to simulate stream-aquifer relations using a modular, finite-difference, ground-water flow model: U.S. Geological Survey Open-File Report 88–729, 113 p.

Ralston, D.R., and Chapman, S.L., 1969, Water-level changes in the Mud Lake area, Idaho, 1958–68: Idaho Department of Reclamation Water Information Bulletin 7, 29 p.

Robertson, J.B., 1977, Numerical modeling of subsurface radioactive solute transport from waste-seepage ponds at the Idaho National Engineering Laboratory: U.S. Geological Survey Open-File Report 76–717 (IDO–22057), 68 p.

Russell, I.C., 1902, Geology and water resources of the Snake River Plains of Idaho: U.S. Geological Survey Bulletin 199, 192 p.

Skibitzke, H.E., and da Costa, J.A., 1962, The ground-water flow system in the Snake River Plain, Idaho—an idealized analysis: U.S. Geological Survey Water-Supply Paper 1536–D, 67 p.

Skrivan, J.A., and Karlinger, M.R., 1980, Semi-variogram estimation and universal kriging program: U.S. Geological Survey Computer Contribution, National Technical Information Service, PB 81–129560, 98 p.

Spinazola, J.M., 1994, Simulation of changes in water levels and ground-water flow in response to water-use alternatives in the Mud Lake area, eastern Snake River Plain, eastern Idaho: U.S. Geological Survey Water-Resources Investigations Report 93–4228, 29 p.

Spinazola, J.M., Tungate, A.M., and Rogers, T.L., 1992, Geohydrologic and chemical data from wells in the Mud Lake area, eastern Idaho, 1988–91: U.S. Geological Survey Open-File Report 92–133, 92 p.

Stearns, H.T., Bryan, L.L., and Crandall, Lynn, 1939, Geology and water resources of the Mud Lake region, Idaho: U.S. Geological Survey Water-Supply Paper 818, 125 p., 13 pls.

Stearns, H.T., Crandall, Lynn, and Steward, W.G., 1938, Geology and ground-water resources of the Snake River Plain in southeastern Idaho: U.S. Geological Survey Water-Supply Paper 774, 268 p., 31 pls.

Theis, C.V., Brown, R.H., and Meyer, R.R., 1963, Estimating the transmissibility of aquifers from the specific capacity of wells, *in* Bentall, Ray, compiler, Methods of determining permeability, transmissibility and drawdown: U.S. Geological Survey Water-Supply Paper 1536–I, p. 331–341.

Thomas, C.A., Broom, H.C., and Cummans, J.E., 1963, Magnitude and frequency of floods in the United States, Part 13, Snake River basin: U.S. Geological Survey Water-Supply Paper 1688, 244 p.

Whitehead, R.L., 1986, Geohydrologic framework of the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Hydrologic Investigations Atlas HA–681, 3 sheets, scale 1:1,000,000.

Wytzes, Jetze, 1980, Development of a groundwater model for the Henry's Fork and Rigby Fan areas, upper Snake River Basin, Idaho: Moscow, University of Idaho, Idaho Water and Energy Resources Research Institute Research Technical Completion Report, 205 p.

TABLES 4 AND 5

(p. 71 follows)

Table 4. General-head package data for simulated underflow at model boundaries

[Row and column numbers are shown on figure 36]

Layer	Row	Column	Fixed head (feet above sea level)	Conductance (cubic feet per feet squared per foot per day)
1	8	1	4,525.39	150,000
1	9	1	4,531.34	150,000
1	10	1	4,534.17	150,000
1	11	1	4,535.76	150,000
1	12	1	4,536.81	150,000
1	13	1	4,537.49	150,000
1	14	1	4,538.28	150,000
1	15	1	4,538.82	150,000
1	16	1	4,539.11	150,000
1	17	1	4,539.32	100
1	18	1	4,539.65	100
1	19	1	4,539.51	100
1	20	1	4,539.01	100
1	21	1	4,537.98	100
1	22	1	4,536.30	100
1	23	1	4,533.86	100
1	24	1	4,530.59	100
1	25	1	4,526.44	100
1	26	1	4,521.38	100
1	27	1	4,515.43	100
1	28	1	4,508.61	100
1	29	1	4,504.62	250,000
1	30	1	4,504.13	250,000
1	31	1	4,503.27	250,000
1	32	1	4,502.26	250,000
1	33	1	4,501.24	250,000
1	34	1	4,500.33	250,000
1	35	1	4,499.34	250,000
1	36	1	4,498.26	250,000
1	37	1	4,496.17	250,000
1	38	1	4,497.62	250,000
1	39	1	4,497.95	250,000
1	40	1	4,498.12	250,000
2	8	1	4,529.90	150,000
2	9	1	4,532.40	150,000
2	10	1	4,534.49	150,000
2	11	1	4,535.77	150,000
2	12	1	4,536.74	150,000
2	13	1	4,537.49	150,000
2	14	1	4,538.23	150,000
2	15	1	4,538.76	150,000
2	16	1	4,539.05	150,000
2	17	1	4,539.30	100
2	18	1	4,539.63	100
2	19	1	4,539.50	100
2	20	1	4,539.00	100
2	21	1	4,537.97	100
2	22	1	4,536.29	100
2	23	1	4,533.85	100
2	24	1	4,530.58	100
2	25	1	4,526.43	100
2	26	1	4,521.37	100
2	27	1	4,515.42	100
2	28	1	4,508.61	100
2	29	1	4,504.64	250,000
2	30	1	4,504.15	250,000
2	31	1	4,503.28	250,000
2	32	1	4,502.27	250,000
2	33	1	4,501.24	250,000
2	34	1	4,500.33	250,000
2	35	1	4,499.34	250,000
2	36	1	4,498.26	250,000
2	37	1	4,496.17	250,000
2	38	1	4,497.63	250,000
2	39	1	4,497.95	250,000
2	40	1	4,498.13	250,000
3	8	1	4,529.90	450,000
3	9	1	4,532.40	450,000
3	10	1	4,534.14	450,000
3	11	1	4,535.36	450,000
3	12	1	4,536.38	450,000
3	13	1	4,537.25	450,000
3	14	1	4,537.97	450,000
3	15	1	4,538.52	450,000
3	16	1	4,538.81	450,000
3	17	1	4,539.26	300
3	18	1	4,539.59	300
3	19	1	4,539.48	300
3	20	1	4,538.98	300
3	21	1	4,537.95	300
3	22	1	4,536.27	300
3	23	1	4,533.83	300
3	24	1	4,530.56	300
3	25	1	4,526.41	300
3	26	1	4,521.35	300
3	27	1	4,515.41	300
3	28	1	4,508.59	300
3	29	1	4,504.70	750,000
3	30	1	4,504.20	750,000
3	31	1	4,503.33	750,000
3	32	1	4,502.29	750,000
3	33	1	4,501.24	750,000
3	34	1	4,500.37	750,000
3	35	1	4,499.38	750,000
3	36	1	4,498.26	750,000
3	37	1	4,496.17	750,000
3	38	1	4,497.67	750,000
3	39	1	4,498.02	750,000
3	40	1	4,498.19	750,000
4	9	1	4,532.40	500,000

Table 4. General-head package data for simulated underflow at model boundaries—Continued

Layer	Row	Column	Fixed head (feet above sea level)	Conductance (cubic feet per feet squared per foot per day)
4	10	1	4,534.31	500,000
4	11	1	4,535.35	500,000
4	12	1	4,536.30	500,000
4	13	1	4,537.15	500,000
4	14	1	4,537.88	500,000
4	15	1	4,538.42	500,000
4	16	1	4,538.72	500,000
4	17	1	4,539.19	3,333
4	18	1	4,539.53	3,333
4	19	1	4,539.44	3,333
4	20	1	4,538.94	3,333
4	21	1	4,537.91	3,333
4	22	1	4,536.23	3,333
4	23	1	4,533.79	3,333
4	24	1	4,530.53	3,333
4	25	1	4,526.38	3,333
4	26	1	4,521.32	3,333
4	27	1	4,515.37	3,333
4	28	1	4,508.58	3,333
4	29	1	4,504.82	555,556
4	30	1	4,504.28	555,556
4	31	1	4,503.38	555,556
4	32	1	4,502.30	555,556
4	33	1	4,501.24	555,556
4	34	1	4,500.39	555,556
4	35	1	4,499.40	555,556
4	36	1	4,498.26	555,556
4	37	1	4,496.17	555,556
4	38	1	4,497.68	555,556
4	39	1	4,498.03	555,556
4	40	1	4,498.22	555,556
5	14	1	4,538.41	500,000
5	15	1	4,539.05	500,000
5	16	1	4,539.25	500,000
5	17	1	4,539.37	3,333
5	18	1	4,539.54	3,333
5	19	1	4,539.43	3,333
5	20	1	4,538.93	3,333
5	21	1	4,537.89	3,333
5	22	1	4,536.20	3,333
5	23	1	4,533.76	3,333
5	24	1	4,530.50	3,333
5	25	1	4,526.35	3,333
5	26	1	4,521.29	3,333
5	27	1	4,515.36	3,333
5	28	1	4,508.62	3,333
5	29	1	4,504.90	277,778
5	30	1	4,504.33	277,778
5	31	1	4,503.40	277,778
5	32	1	4,502.30	277,778
5	33	1	4,501.24	277,778
5	34	1	4,500.38	277,778
5	35	1	4,499.40	277,778
5	36	1	4,498.26	277,778
5	37	1	4,496.17	277,778
5	38	1	4,497.68	277,778
5	39	1	4,498.03	277,778
5	40	1	4,498.21	277,778
1	40	2	4,513.00	8,620
1	40	3	4,515.00	8,620
1	40	4	4,520.00	8,928
1	40	5	4,523.00	8,928
1	40	6	4,527.00	8,928
1	40	7	4,528.00	9,259
1	40	8	4,530.00	9,615
1	40	9	4,535.00	9,615
1	40	10	4,538.00	9,615
1	40	11	4,543.00	10,000
1	40	12	4,547.00	10,000
1	40	13	4,550.00	10,000
1	40	14	4,553.00	10,417
1	40	15	4,557.00	10,870
1	40	16	4,560.00	10,870
1	40	17	4,564.00	10,870
1	40	18	4,569.00	10,870
1	40	19	4,573.00	11,364
1	40	20	4,577.00	11,905
1	40	21	4,590.00	2,500
1	40	22	4,650.00	25,000
1	40	23	4,700.00	25,000
1	40	24	4,725.00	25,000
1	40	25	4,765.00	400,000
1	40	26	4,770.00	400,000
1	40	27	4,775.00	400,000
1	40	28	4,780.00	400,000
1	40	29	4,781.00	400,000
1	40	30	4,782.00	400,000
1	40	31	4,783.00	30,769
1	40	32	4,785.00	30,769
1	40	33	4,790.00	30,769
1	40	34	4,795.00	30,769
1	40	35	4,830.00	30,769
1	40	36	4,850.00	3,846
1	40	37	4,860.00	3,571
1	40	38	4,860.00	3,333
1	40	39	4,860.00	3,333

Table 5. Stream package data for stream and lake segments

[Relations among stream and lake segments, ground-water inflow, diversions, and lake evapotranspiration are shown on figure 9; reaches are numbered in a downstream order; row and column numbers are shown on figure 36; values for streamflow, ground-water inflow, diversions, and lake evapotranspiration are listed only for the first reach in each segment; —, indicates that streamflow to the first reach of the segment is calculated by the numerical model; datum for stage and streambed bottom is sea level]

Name of stream or lake segment, surface inflow or surface outflow	Reach number	Row	Column	Streamflow, ground-water inflow, diversion, or lake evapotranspiration (acre-feet per year)	Stage (feet)	Streambed bottom (feet)	Vertical hydraulic conductivity (feet per day)	Reach length (feet)	Reach width (feet)
Medicine Lodge Creek	1	3	30	44,916	5,347	5,020	0.0182	5,754	15
	2	4	30		5,287	4,996	.0205	5,955	15
	3	5	30		5,231	4,968	.0228	6,133	15
	4	6	29		5,166	4,930	.0254	6,801	15
	5	7	29		5,115	4,888	.0263	2,287	15
	6	7	28		5,085	4,884	.0298	5,057	15
	7	8	28		5,052	4,850	.0295	2,674	15
	8	8	27		5,015	4,843	.0347	5,646	15
	9	9	27		4,979	4,804	.0300	2,095	15
	10	9	26		4,952	4,792	.0389	6,074	15
	11	9	25		4,932	4,790	.0420	5,175	15
	12	8	25		4,926	4,789	.0438	1,802	15
	13	8	24		4,917	4,786	.0456	5,373	15
	14	9	24		4,907	4,787	.0509	1,014	15
	15	9	23		4,894	4,785	.0546	6,554	15
	16	10	23		4,875	4,786	.0668	1,961	15
	17	10	22		4,867	4,783	.0706	1,351	15
Beaver Creek (upstream from diversion)	1	1	43	29,419	5,881	5,537	.0023	5,465	15
	2	2	43		5,816	5,414	.0020	6,452	15
	3	3	43		5,716	5,346	.0022	6,086	15
	4	3	42		5,650	5,315	.0024	920	15
	5	4	42		5,603	5,272	.0024	5,204	15
	6	4	41		5,558	5,240	.0025	5,561	15
	7	4	40		5,510	5,198	.0026	4,978	15
	8	3	40		5,497	5,224	.0029	1,495	15
	9	3	39		5,479	5,175	.0026	1,684	15
	10	4	39		5,442	5,153	.0028	5,546	15
	11	4	38		5,424	5,108	.0025	1,872	15
	12	5	38		5,405	5,081	.0025	5,213	15
	13	5	37		5,400	5,054	.0023	3,112	15
	14	6	37		5,377	5,031	.0023	6,875	15
	15	7	37		5,306	5,005	.0028	6,251	15
	16	8	37		5,279	4,975	.0023	2,159	15
Beaver Creek (downstream from diversion)	1	8	36	—	5,251	4,945	.0026	5,265	15
	2	9	36		5,208	4,908	.0027	5,071	15
	3	9	35		5,184	4,847	.0024	3,856	15
	4	10	35		5,160	4,828	.0024	6,191	15
	5	11	35		5,124	4,820	.0026	5,681	15
	6	11	34		5,102	4,795	.0025	625	15
	7	12	34		5,082	4,794	.0028	6,933	15
	8	13	34		5,035	4,800	.0034	2,785	15
	9	13	33		5,012	4,792	.0036	6,227	15
	10	14	33		5,002	4,792	.0037	1,893	15
	11	14	32		4,976	4,791	.0044	4,191	15
	12	15	32		4,950	4,790	.0050	6,088	15
	13	16	32		4,925	4,790	.0060	256	15

Table 5. Stream package data for stream and lake segments—Continued

Name of stream or lake segment, surface inflow or surface outflow	Reach number	Row	Column	Streamflow, ground-water inflow, diversion, or lake evapotranspiration (acre-feet per year)	Stage (feet)	Streambed bottom (feet)	Vertical hydraulic conductivity (feet per day)	Reach length (feet)	Reach width (feet)
Beaver Creek (downstream from diversion) —Continued	14	16	31		4,900	4,789	0.0072	5,837	15
	15	17	30		4,890	4,788	.0084	124	15
	16	17	31		4,875	4,789	.0093	6,204	15
	17	18	31		4,871	4,789	.0098	371	15
	18	18	30		4,861	4,788	.0110	7,282	15
	19	19	30		4,850	4,788	.0129	7,940	15
	20	20	30		4,836	4,788	.0161	2,833	15
	21	20	29		4,831	4,787	.0185	4,857	15
	22	21	29		4,821	4,786	.0233	7,615	15
	23	21	28		4,800	4,785	.0544	3,270	15
Beaver Creek diversion	1	8	37	3,528	5,279	4,975	.0023	2,159	15
Camas Creek (upper segment)	1	3	58	80,318	6,439	6,076	.0005	6,498	20
	2	4	56		6,391	5,842	.0003	6,723	20
	3	4	58		6,387	5,964	.0006	4,977	20
	4	4	57		6,375	5,918	.0005	1,618	20
	5	5	56		6,365	5,772	.0004	4,996	20
	6	5	57		6,361	5,843	.0005	7,197	20
	7	6	56		6,342	5,728	.0004	6,504	20
	8	6	57		6,337	5,792	.0005	7,839	20
	9	7	57		6,311	5,750	.0004	5,349	20
	10	7	56		6,305	5,692	.0004	11,057	20
	11	8	56		6,293	5,657	.0004	4,448	20
	12	8	55		6,275	5,602	.0004	8,171	20
	13	9	55		6,252	5,571	.0004	1,541	20
	14	9	54		6,251	5,522	.0003	10,853	20
	15	10	54		6,244	5,491	.0003	863	20
	16	10	53		6,244	5,447	.0003	7,244	20
	17	11	53		6,242	5,417	.0003	6,580	20
	18	12	53		6,219	5,385	.0003	7,456	20
	19	12	52		6,199	5,346	.0003	7,949	20
	20	13	51		6,199	5,278	.0003	1,397	20
	21	13	52		6,197	5,313	.0003	4,723	20
	22	14	51		6,187	5,252	.0003	6,520	20
Camas Creek (middle segment)	1	14	50	—	6,097	5,224	.0003	4,602	20
	2	13	50		6,089	5,243	.0003	1,849	20
	3	14	49		5,978	5,177	.0003	7,519	20
	4	15	49		5,890	5,141	.0003	298	20
	5	15	48		5,818	5,106	.0004	9,269	20
	6	15	47		5,722	5,073	.0004	6,430	20
	7	16	46		5,681	5,011	.0003	4,482	20
	8	15	46		5,677	5,043	.0005	2,658	20
	9	15	45		5,631	5,016	.0004	7,041	20
	10	15	44		5,558	4,993	.0004	7,192	20
	11	16	44		5,542	4,969	.0004	427	20
	12	16	43		5,513	4,952	.0004	7,983	20
	13	16	42		5,463	4,935	.0005	7,886	20
	14	16	41		5,416	4,918	.0005	3,371	20
	15	17	41		5,400	4,904	.0005	3,087	20

Table 5. Stream package data for stream and lake segments—Continued

Name of stream or lake segment, surface inflow or surface outflow	Reach number	Row	Column	Streamflow, ground-water inflow, diversion, or lake evapotranspiration (acre-feet per year)	Stage (feet)	Streambed bottom (feet)	Vertical hydraulic conductivity (feet per day)	Reach length (feet)	Reach width (feet)
Camas Creek (middle segment) —Continued	16	17	40		5,340	4,890	0.0006	3,635	20
	17	16	40		5,296	4,902	.0006	2,886	20
	18	16	39		5,229	4,886	.0007	2,209	20
	19	17	39		5,208	4,876	.0007	5,165	20
	20	17	38		5,172	4,862	.0008	7,195	20
Wood's diversion	1	14	51	8,366	6,168	5,252	.0043	2,402	20
	2	15	50		6,141	5,188	.0042	5,749	20
	3	14	50		6,133	5,224	.0044	4,411	20
Camas Creek diversion	1	17	38	6,680	5,172	4,862	.0008	7,195	20
Camas Creek (middle segment)	1	17	37	—	5,111	4,848	.0005	6,936	20
	2	17	36		5,070	4,834	.0006	6,629	20
	3	17	35		5,029	4,819	.0006	6,485	20
	4	18	35		5,005	4,814	.0007	1,040	20
	5	18	34		4,988	4,799	.0007	5,718	20
	6	18	33		4,945	4,790	.0009	415	20
	7	17	33		4,914	4,790	.0011	6,222	20
	8	17	32		4,899	4,790	.0012	3,024	20
	9	18	32		4,894	4,790	.0013	7,380	20
	10	19	32		4,885	4,790	.0014	1,272	20
	11	19	31		4,874	4,789	.0016	5,576	20
	12	20	31		4,856	4,789	.0020	5,336	20
	13	20	30		4,838	4,788	.0027	4,439	20
	14	21	30		4,830	4,788	.0032	3,741	20
Camas Creek diversion (Rays Lake segment upstream from refuge)	1	21	29	—	4,822	4,786	.0243	6,839	30
	2	21	28		4,802	4,785	.0513	4,867	30
	3	22	28		4,798	4,784	.0642	1,672	30
Camas Creek diversion (Rays Lake segment downstream from refuge)	1	22	27	—	4,798	4,782	.0557	6,851	30
	2	22	26		4,797	4,781	.0545	6,723	30
	3	22	25		4,797	4,779	.0441	6,211	30
	4	23	25		4,797	4,777	.0434	8,698	30
	5	24	25		4,796	4,773	.0384	1,313	30
	6	24	24		4,790	4,771	.0458	7,132	30
	7	25	24		4,786	4,766	.0434	10,145	30
Camas Creek diversions from Rays Lake segment to lakes in Camas National Wildlife Refuge	1	22	28	11,051	4,785	4,784	.0078	230	230
Lakes in Camas National Wildlife Refuge	1	23	27	—	4,785	4,781	.0017	1,394	1,394
	2	24	27		4,785	4,779	.0013	3,129	3,129
	3	24	28		4,785	4,781	.0021	133	133
	4	25	27		4,785	4,776	.0008	2,293	2,293

Table 5. Stream package data for stream and lake segments—Continued

Name of stream or lake segment, surface inflow or surface outflow	Reach number	Row	Column	Streamflow, ground-water inflow, diversion, or lake evapotranspiration (acre-feet per year)	Stage (feet)	Streambed bottom (feet)	Vertical hydraulic conductivity (feet per day)	Reach length (feet)	Reach width (feet)
Lakes in Camas National Wildlife Refuge —Continued	5	25	26		4,785	4,773	0.0006	3,476	3,476
	6	26	25		4,785	4,764	.0004	1,356	1,356
	7	25	25		4,785	4,769	.0004	3,230	3,230
	8	26	24		4,785	4,757	.0003	2,276	2,276
	9	25	24		4,785	4,766	.0004	4,067	4,067
	10	26	23		4,785	4,747	.0001	745	745
	11	25	23		4,785	4,761	.0003	1,152	1,152
	12	24	24		4,785	4,771	.0006	3,153	3,153
	13	24	25		4,785	4,773	.0007	3,790	3,790
	14	24	26		4,785	4,776	.0008	3,397	3,397
	15	23	25		4,785	4,777	.0010	3,802	3,802
	16	23	24		4,785	4,775	.0007	823	823
	17	22	24		4,785	4,778	.0010	660	660
	18	22	25		4,785	4,779	.0013	2,604	2,604
	19	23	26		4,785	4,779	.0011	2,690	2,690
	20	22	27		4,785	4,782	.0029	586	586
	21	21	27		4,785	4,784	.0055	1,488	1,488
	22	21	28		4,785	4,780	.0897	2,054	2,054
Lake evapotranspiration minus ground-water inflow for lakes on Camas National Wildlife Refuge	1	22	28	2,649	4,785	4,784	.0056	230	230
Ground-water inflow to Mud Lake segment of Camas Creek	1	25	23	72,952	4,785	4,761	.0000	0	0
Camas Creek (Mud Lake segment)	1	25	23	—	4,785	4,761	.0001	5,135	30
	2	25	22		4,783	4,756	.0001	6,220	30
	3	25	21		4,783	4,752	.0001	6,746	30
	4	25	20		4,781	4,755	.0001	6,221	30
	5	25	19		4,781	4,782	.0150	3,362	30
	6	24	19		4,782	4,799	.0002	1,453	30
Lake evapotranspiration and diversions from Mud Lake	1	24	17	75,095	4,780	4,774	.0000	0	0
Mud Lake	1	25	19	—	4,780	4,775	.1436	625	625
	2	24	19		4,780	4,775	.0002	4,791	4,791
	3	24	18		4,780	4,775	.0002	4,936	4,936
	4	25	18		4,780	4,772	.0051	1,523	1,523
	5	24	17		4,780	4,774	.0007	4,586	4,586
	6	23	16		4,780	4,775	.0005	3,597	3,597
	7	23	15		4,780	4,775	1.7076	71	71
	8	22	15		4,780	4,775	.0012	2,871	2,871
	9	22	16		4,780	4,775	.0037	2,285	2,285
	10	23	17		4,780	4,775	.0003	3,590	3,590

Table 5. Stream package data for stream and lake segments—Continued

Name of stream or lake segment, surface inflow or surface outflow	Reach number	Row	Column	Streamflow, ground-water inflow, diversion, or lake evapotranspiration (acre-feet per year)	Stage (feet)	Streambed bottom (feet)	Vertical hydraulic conductivity (feet per day)	Reach length (feet)	Reach width (feet)
Mud Lake —Continued	11	23	18		4,780	4,775	0.0002	4,295	4,295
	12	22	18		4,780	4,775	.0166	1,623	1,623
	13	23	19		4,780	4,775	.0092	1,704	1,704
	14	24	20		4,780	4,775	.0318	1,355	1,355
Henrys Fork	1	35	60	1,182,283	5,082	5,077	.0006	5,457	200
	2	35	59		5,063	5,058	.0007	6,081	200
	3	35	58		5,048	5,043	.0007	1,865	200
	4	36	58		5,044	5,039	.0007	4,125	200
	5	37	56		5,038	5,033	.0006	2,356	200
	6	36	57		5,037	5,032	.0006	6,471	200
	7	37	55		5,028	5,023	.0006	5,509	200
	8	36	56		5,026	5,021	.0007	3,819	200
	9	37	54		5,001	4,996	.0008	5,756	200
	10	37	53		4,995	4,990	.0008	5,673	200
	11	37	52		4,989	4,984	.0008	5,858	200
	12	37	51		4,982	4,977	.0009	6,030	200
	13	37	50		4,962	4,957	.0010	6,004	200
	14	37	49		4,908	4,903	.0021	5,316	200
	15	36	49		4,896	4,891	.0027	2,392	200
	16	37	48		4,885	4,880	.0034	8,074	200
	17	37	47		4,882	4,877	.0036	12,832	200
	18	37	46		4,864	4,859	.0084	5,943	200
	19	37	45		4,852	4,847	.0344	6,295	200
	20	36	45		4,848	4,843	.0596	1,390	200
	21	37	44		4,840	4,835	.0158	6,505	200
	22	37	43		4,837	4,832	.0141	1,650	200
	23	36	43		4,835	4,830	.0083	5,499	200
	24	36	42		4,831	4,826	.0063	1,800	200
	25	37	42		4,831	4,826	.0090	8,075	200
	26	37	41		4,831	4,826	.0089	11,062	200
	27	38	41		4,832	4,827	.0145	2,283	200
	28	37	40		4,827	4,822	.0072	3,479	200
	29	38	40		4,826	4,821	.0094	9,383	200
	30	39	39		4,823	4,818	.0120	1,971	200
	31	38	39		4,822	4,817	.0081	12,190	200
	32	39	38		4,821	4,816	.0164	9,626	200
	33	39	37		4,821	4,816	.1718	4,497	200
	34	40	34		4,819	4,814	.0061	8,343	200
	35	40	37		4,818	4,813	.0325	3,127	200
	36	40	36		4,814	4,809	.1143	7,753	200
	37	40	35		4,810	4,805	.0306	2,797	200
Snake River	1	40	30	2,800,585	4,780	4,775	.4252	4,559	100
	2	40	29		4,777	4,772	.4257	4,399	100
	3	39	29		4,774	4,769	.2415	6,909	100
	4	39	28		4,772	4,767	.3583	9,141	100
	5	39	27		4,767	4,762	.2951	10,200	100
	6	39	26		4,764	4,759	.4921	1,089	100
	7	40	26		4,763	4,758	.3124	8,641	100
	8	39	25		4,763	4,758	.9071	3,754	100
	9	40	25		4,762	4,757	3.0589	8,954	100

Table 5. Stream package data for stream and lake segments—Continued

Name of stream or lake segment, surface inflow or surface outflow	Reach number	Row	Column	Streamflow, ground-water inflow, diversion, or lake evapotranspiration (acre-feet per year)	Stage (feet)	Streambed bottom (feet)	Vertical hydraulic conductivity (feet per day)	Reach length (feet)	Reach width (feet)
Lakes in Market Lake Wildlife Management Area	1	38	26	0	4,765	4,764	0.0020	605	605
	2	38	25		4,765	4,764	.0001	951	951
	3	37	25		4,765	4,764	.0001	1,573	1,573
	4	36	25		4,765	4,764	.0001	2,083	2,083
	5	35	25		4,765	4,764	.0001	3,018	3,018
	6	34	25		4,765	4,764	.0001	76	76
	7	35	24		4,765	4,764	.0001	2,493	2,493
	8	36	24		4,765	4,764	.0001	1,713	1,713
	9	37	24		4,765	4,764	.0001	284	284