STREAM-AQUIFER SYSTEM IN THE UPPER BEAR RIVER VALLEY, WYOMING

By Kent C. Glover

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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
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per day per foot (ft/d)/ft	0.3048	meter per day per meter
gallon per minute (gal/min)	0.06309	liter per second
gallon per minute per foot (gal/min)/ft	0.2070	liter per second per meter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
foot squared per day (ft^2/d)	0.09290	meter squared per day

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

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ABSTRACT

Pumping from aquifers that are hyraulically connected to the Bear River in western Wyoming is likely to reduce flow in the river. The principal aquifer in the Cokeville, Wyoming area is in alluvial deposits adjacent to the streams; aquifers in the bedrock are isolated hydraulically from the alluvium and are not part of the stream-aquifer system. In the Evanston, Wyoming area the stream-aquifer system includes the alluvial aquifer and the Wasatch aquifer, which generally are connected hydraulically.

A finite-element model of the stream-aquifer system near Cokeville was applied to both steady-state and transient flow. Analysis using the model for a year of less-than-average streamflow shows that approximately 84 percent of water pumped from existing wells will be derived from water that otherwise would have seeped into the Bear River, and 16 percent from water that otherwise would have been lost to phreatophytes. The simulation also shows that the largest reduction in streamflow is likely to occur during August, which correlates with the period of maximum pumping, July and August. The amount of ground-water pumpage is small, in comparison with the total amount of ground-water discharge.

Limited hydrologic data have precluded the construction of a reliable model of the flow system near Evanston. An analytical stream-depletion method, applied to pumpage data from Evanston municipal wells, shows that the largest reduction in streamflow occurs during the pumping season, and streamflow is affected for an extended period after pumping stops. The largest changes in streamflow are the result of pumping from wells completed in the alluvial aquifer. The effect of pumping from wells completed in the Wasatch aquifer occurs over a longer time.

INTRODUCTION

The Bear River originates in the Uinta Mountains of Utah and flows northward along the Utah-Wyoming border before flowing westward into Idaho (fig. 1). The river turns abruptly to the southwest near Soda Springs, Idaho, re-enters Utah, and flows into the Great Salt Lake. The upper Bear River valley is the part of the river valley between the Uinta Mountains and the Wyoming-Idaho State line.

Water in the upper Bear River valley is used for irrigation and for municipal, industrial, and other needs. Numerous diversions from the river have been constructed to irrigate alfalfa, hay, and pasture along the river.

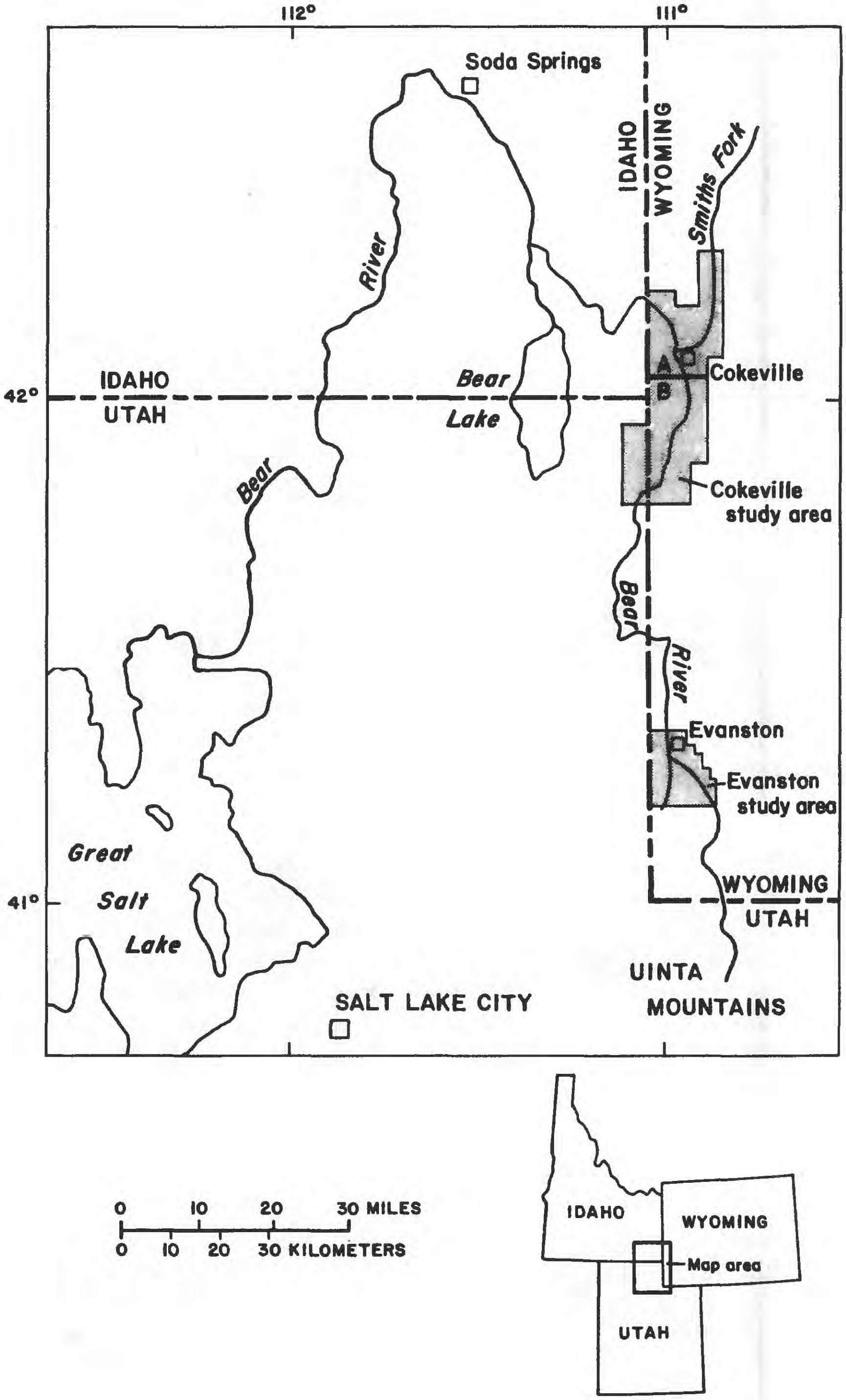


Figure 1.--Location of the Cokeville and Evanston study areas.

Large-capacity wells also are used for irrigation. The City of Evanston obtains water from the Bear River and from ground water. Unincorporated communities rely on ground water for municipal supplies. Recently, exploration for oil and gas in deeply buried rocks of the Overthrust Belt has increased demand for water.

Principal areas of ground-water development are near Cokeville and Evanston, Wyoming (fig. 1). Ground water near Cokeville is used for irrigation; irrigation wells have been drilled in the Bear River alluvium from the Utah-Wyoming border north to the Idaho-Wyoming border. Evanston and nearby unincorporated communities are the principal users of ground water near Evanston. Wells near Evanston have been completed in several formations.

Water from the Bear River drainage is allocated among Idaho, Utah, and Wyoming by interstate compact. Streamflow-gaging stations near State borders are used to identify time periods when water is in short supply. Regulation of some diversions is needed in all but the most wet years.

Water managers in the compact States need additional hydrogeologic information in order to quantify the relation of ground-water movement to streamflow in the Bear River system. Aquifers that are in hydraulic connection with the Bear River have not been identified. The amount of ground water discharging to the Bear River has not been described, and the effects of ground-water withdrawals on streamflow have not been quantified.

In an effort to provide this information, the U.S. Geological Survey and the Wyoming State Engineer began a cooperative study of the stream-aquifer system of the upper Bear River valley in 1982. The two areas selected for detailed study, the Cokeville and Evanston areas, include nearly all large-capacity wells in the upper Bear River valley.

Purpose and Scope

This report describes the interaction between the stream and ground-water systems of the upper Bear River valley. Topics discussed include: (1) The hydrology of the stream-aquifer systems, (2) methods used to evaluate the effects of ground-water withdrawals on streamflow, and (3) results of applications of the methods.

System for Numbering Wells and Surface-Water Stations

Wells cited in this report are numbered according to the Federal system of land subdivision in Wyoming. The first number indicates the township north of the 40th Parallel Base Line, the second the range west of the Sixth Principal Meridian, and the third the section in which the well is located. Lower case letters following the section numbers indicate the position of the well in the section. The first letter denotes the quarter section (160 acres), the second letter the quarter-quarter section (40 acres), and the third letter the quarter-quarter-quarter section (10-acre tract). Subdivisions of a section are lettered a, b, c, and d in a counterclockwise direction, starting in the northeast quarter. If more than one well is listed in a 10-acre tract, consecutive numbers starting with 1 follow the lower case letter of the well number. If a section does not measure 1 mi^2, it is treated as a full section with the southeast section corner serving as the reference point for subdivision of the section. An example is illustrated in figure 2.

An eight-digit station identification number is used by the U.S. Geological Survey to designate surface-water stations in a downstream order. The first two digits identify the major drainage in which the station is located--in this case, 10 (Great Basin). The remaining six digits identify the relative location of the station, with numbers increasing progressively in the downstream direction.

STREAM SYSTEM IN THE UPPER BEAR RIVER VALLEY

Cokeville Study Area

Streams in the Cokeville study area include the Bear River, Smiths Fork, and several smaller tributaries (fig. 3). Sublette Creek and Twin Creek are tributaries with perennial flow. Other tributaries flow into the Bear River only during periods of high runoff. During periods of low flow all water in these tributaries is either diverted for irrigation or lost as ground-water recharge on alluvial fans.

Diversions for irrigation are common along the Bear River and Smiths Fork. Major diversions from the Bear River include, in a downstream order, the B.Q. West Slough, McFarland Ditch, B.Q. Eastside Ditch, Pixley Ditch, and the Cook Canal. The major diversion from Smiths Fork is the Covey Canal. Numerous small diversions from the Bear River and Smiths Fork also have been constructed.

Streamflow of the Bear River and major tributaries is monitored at a series of continuously recording streamflow-gaging stations (fig. 3). Flow in diversions is measured frequently during the irrigation season. Streamflow data are published annually by the U.S. Geological Survey (1981; 1982). Streamflow-diversion data are published biennially by the Bear River Commission (1981; 1983).

Evanston Study Area

Streams in the Evanston study area include the Bear River, Yellow Creek, and several smaller tributaries (fig. 4). Tributaries include Duncomb Hollow, Pleasant Valley Hollow, and Wasatch Creek. Compton Reservoir provides off-channel storage for water from the Bear River.

Diversions for irrigation and municipal use are common along the Bear River (fig. 4). Major diversions are the Rocky Mountain Ditch, used for irrigation, and the Evanston Water Ditch, used for municipal water supply. Numerous small diversions are not mapped on figure 4.

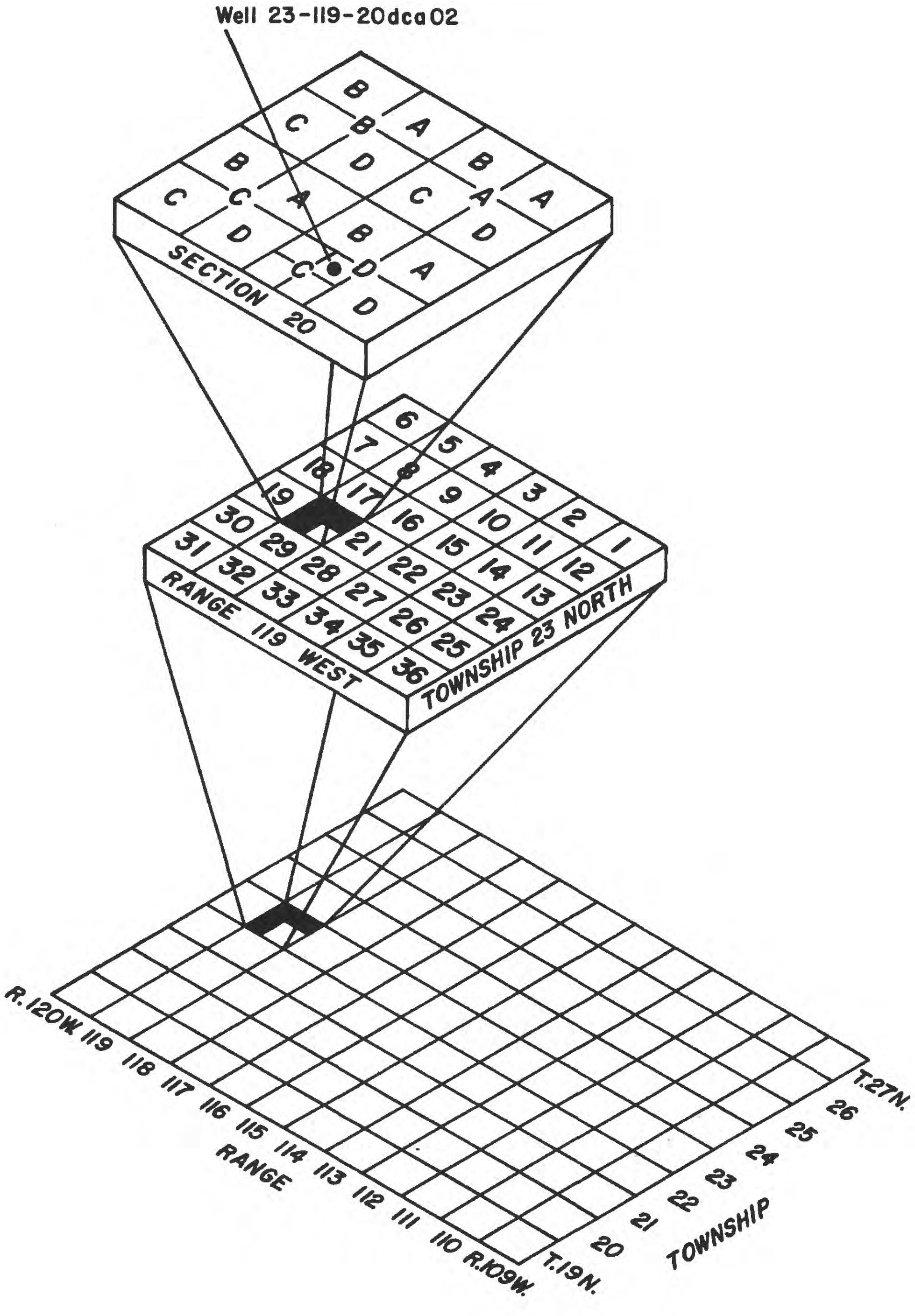


Figure 2.--System of numbering wells.

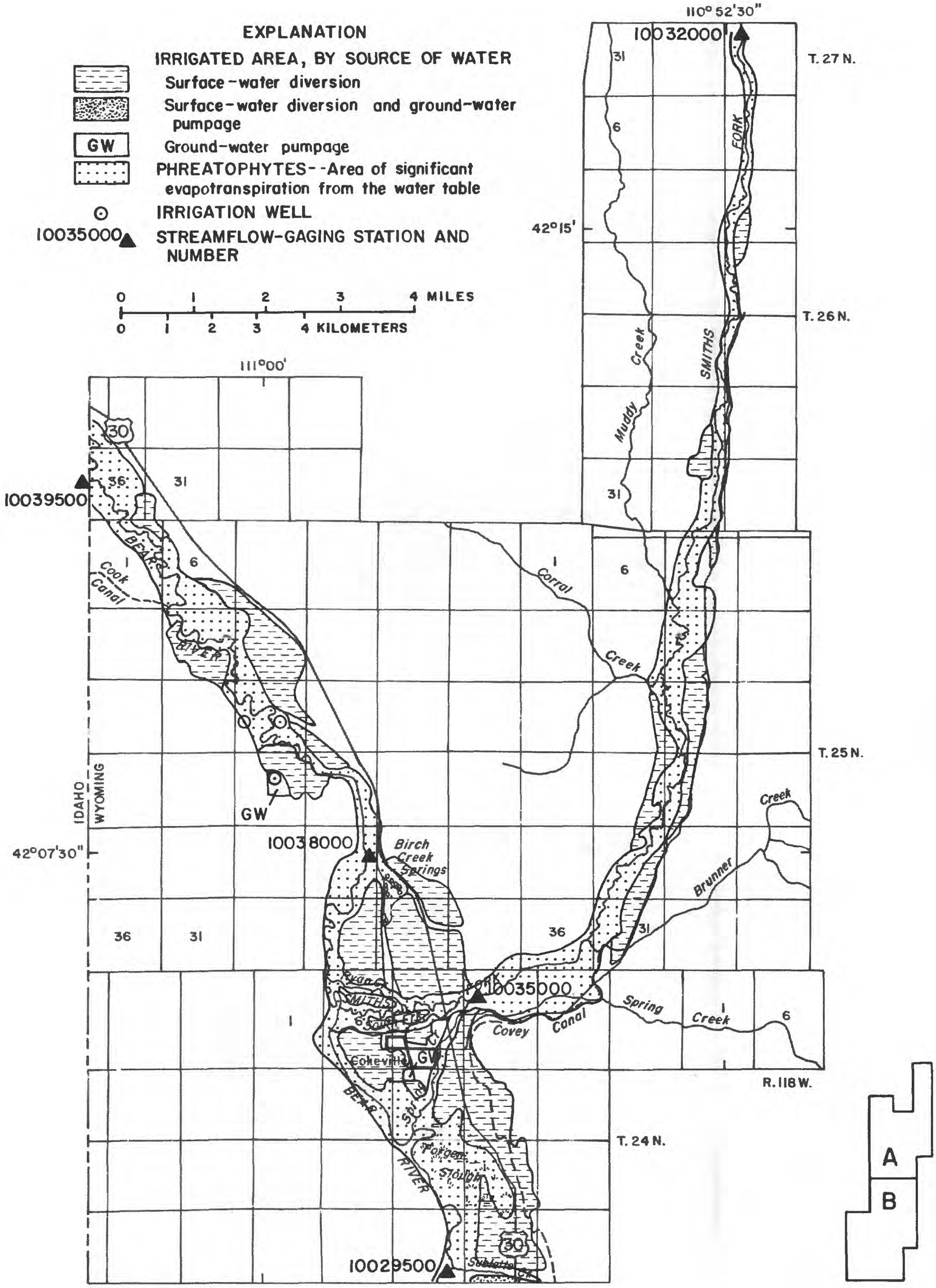
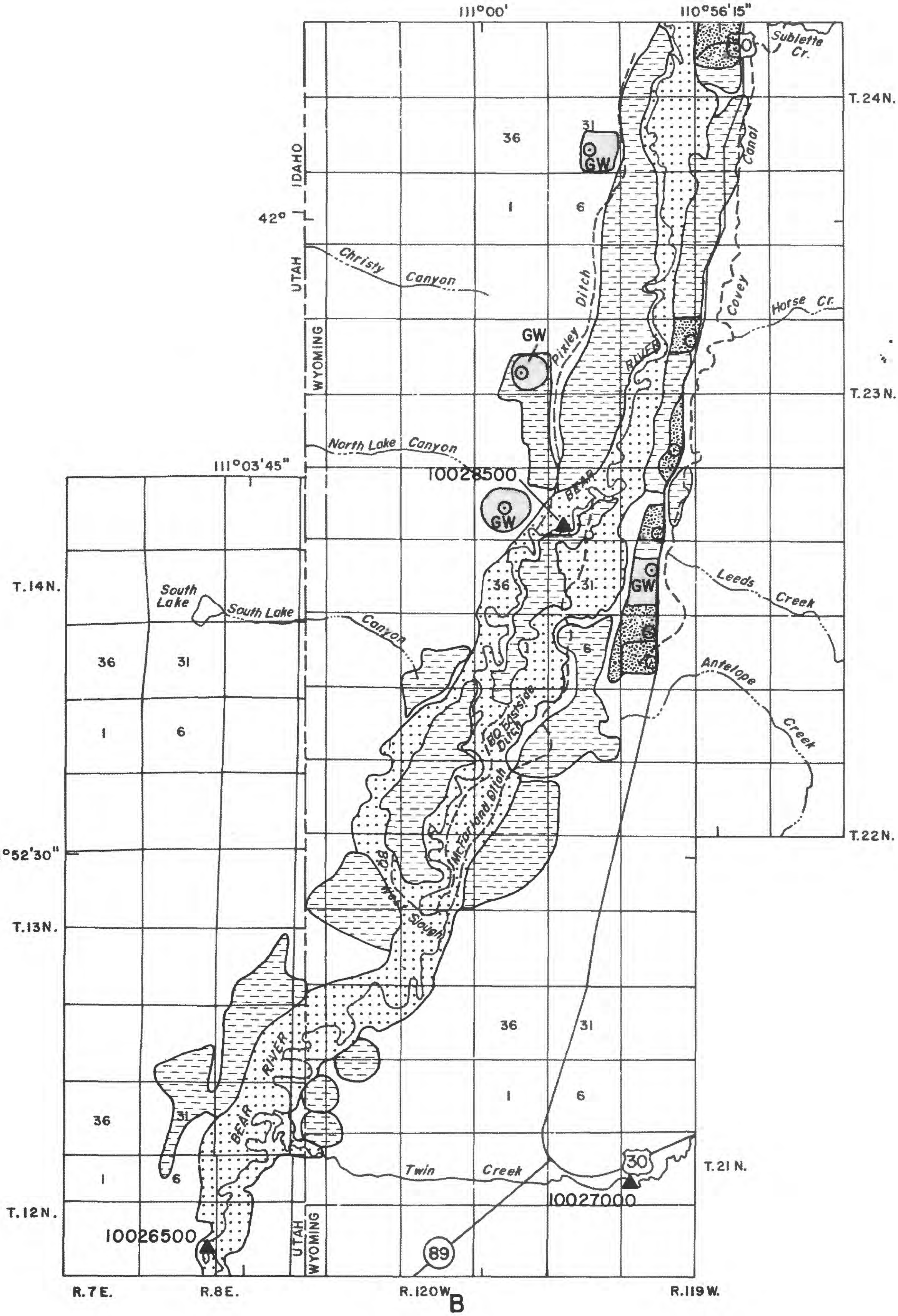


Figure 3.--Irrigated areas and location of irrigation wells and streamflow-gaging stations in the Cokeville study area.


111°00'
110°56'15"
Sublette Cr.
T.24N.
36
31
GW
IDAHO
UTAH
42°
1
6
Christy Canyon
Ditch
Canal
Covey
Horse Cr.
WYOMING
GW
Pixley
RIVER
T.23N.
North Lake Canyon
111°03'45"
10028500
BEAR
GW
GW
T.14N.
South Lake
South Lake
Canyon
36
31
Leeds Creek
36
31
1
6
Antelope Creek
1
6
I.B.O. Eastside Ditch
McFarland Ditch
T.22N.
41°52'30"
West Slough
T.13N.
36
31
BEAR RIVER
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6
36
31
1
6
30
Twin Creek
T.21N.
10027000
T.12N.
10026500
UTAH
WYOMING
89
R.7E.
R.8E.
R.120W.
R.119W.

B

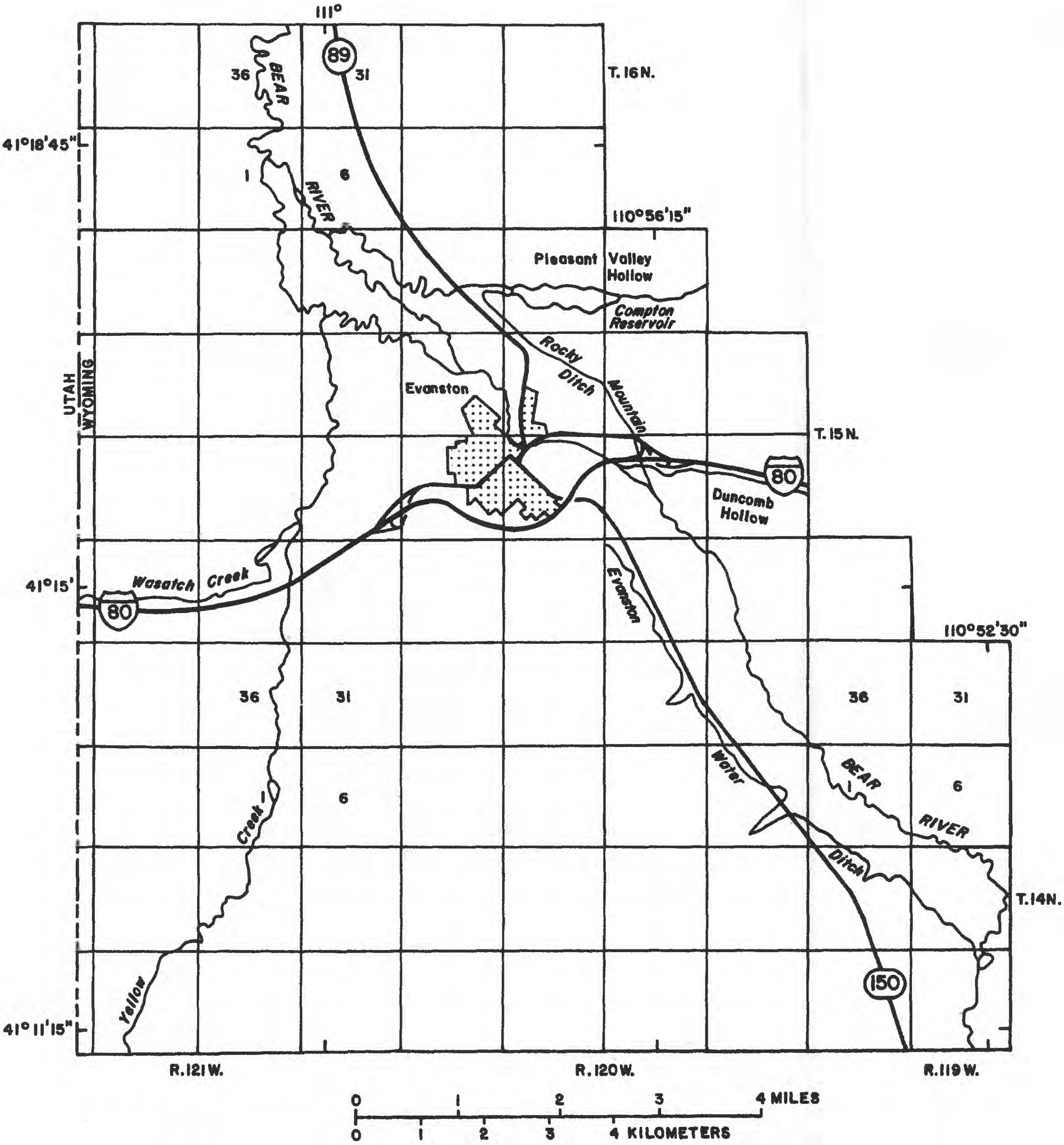


Figure 4.--Streams and surface-water diversions in the Evanston study area.

Although diversions are monitored frequently during the irrigation season (Bear River Commission, 1981; 1983), no continuously recording streamflow-gaging stations are operated in the Evanston study area. The nearest streamflow-gaging station (10020100) is located approximately 20 miles downstream from Evanston near the Utah-Wyoming border (outside the area shown in figure 4). Miscellaneous streamflow measurements have been made at a number of locations within the Evanston study area.

AQUIFERS IN THE UPPER BEAR RIVER VALLEY

Cokeville Study Area

Exposed rocks in the Cokeville study area range in age from Pennsylvanian to Quaternary (fig. 5). Generally, the consolidated rocks of Tertiary age and older either are isolated topographically from the unconsolidated valley fill (Quaternary age), or where not isolated, are shales and siltstones of low permeability.

Unconsolidated sediments of Quaternary age (fig. 5) form a water-table aquifer in the Cokeville study area. The aquifer, called the alluvial aquifer in this report, includes terrace deposits, flood-plain deposits, alluvial-fan deposits, and slope-wash deposits. The deposits consist of rock fragments ranging in size from silt to large cobbles. The degree of sorting of these materials can change greatly over short distances.

Thickness of the alluvial aquifer in the Cokeville study area is unknown, because water wells completed in the aquifer have not been drilled to bedrock. Well depths of 200 ft are common, although well 23-119-20dca02 is 400 ft, and well 23-119-29dcc is 450 ft deep.

Minor aquifers in the Cokeville study area are present in the Wasatch Formation of Tertiary age, the Nugget Sandstone of Jurassic and Triassic age, and the Wells Formation of Permian and Pennsylvanian age (fig. 5; and Lines and Glass, 1975). The Wasatch Formation and Nugget Sandstone crop out in the study area primarily west of the Bear River, whereas the Wells Formation crops out east of the Bear River near the Utah-Wyoming border. Few wells have been drilled in these aquifers, and aquifer characteristics generally are unknown.

Aquifers in the Wasatch Formation, Nugget Sandstone, and Wells Formation are not hydraulically connected to the alluvial aquifer and therefore are not part of the stream-aquifer system in the Cokeville study area. With the exception of a small area of the Wells Formation in the southern part of the study area, the bedrock aquifers are separated from the alluvial aquifer by low-permeability shale and siltstone.

Evanston Study Area

Exposed rocks in the Evanston study area range in age from Jurassic to Quaternary (fig. 6). In contrast to the Cokeville study area, there is a hydraulic connection between consolidated rocks of Tertiary age and the alluvium of Quaternary age.

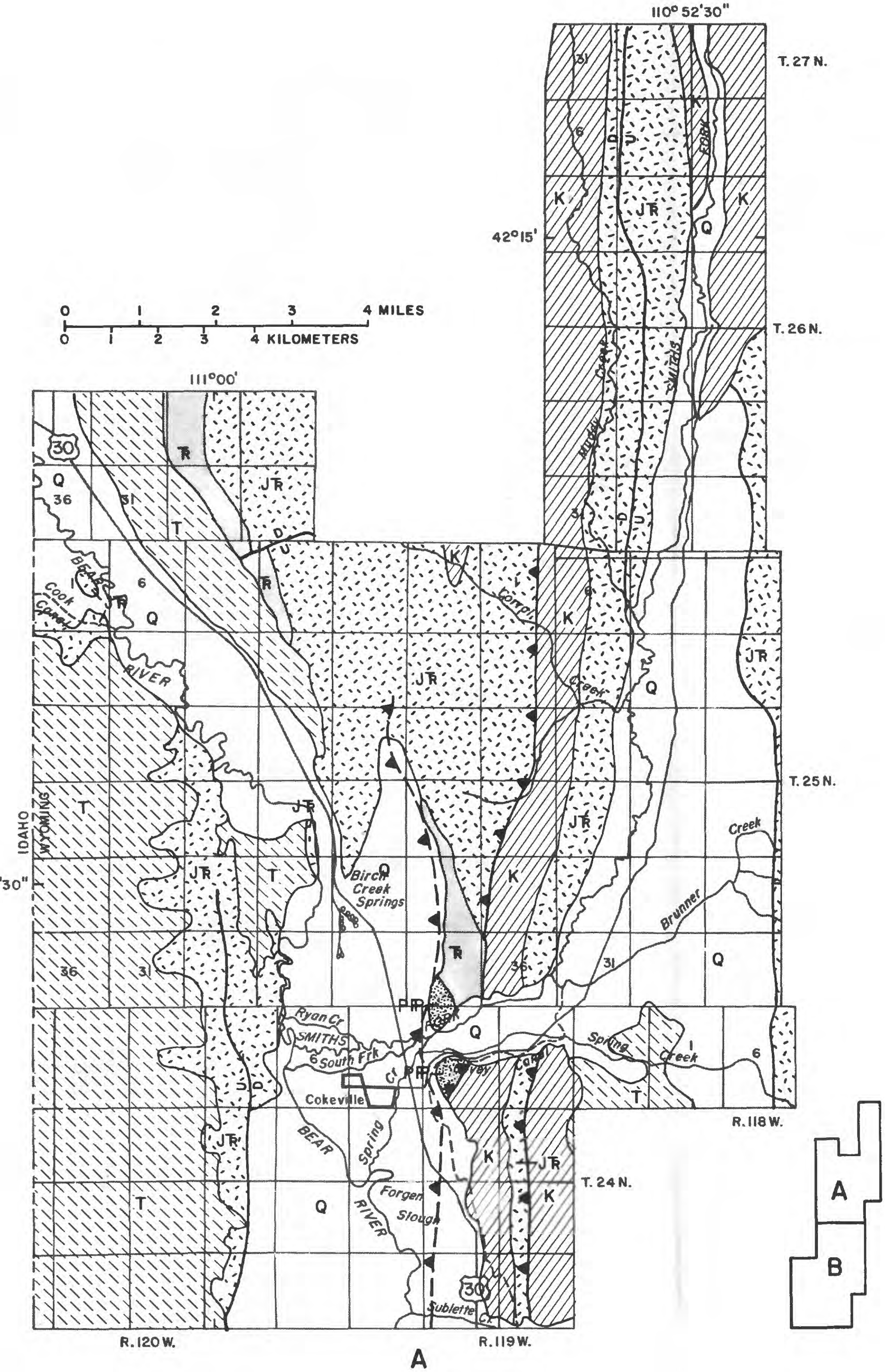
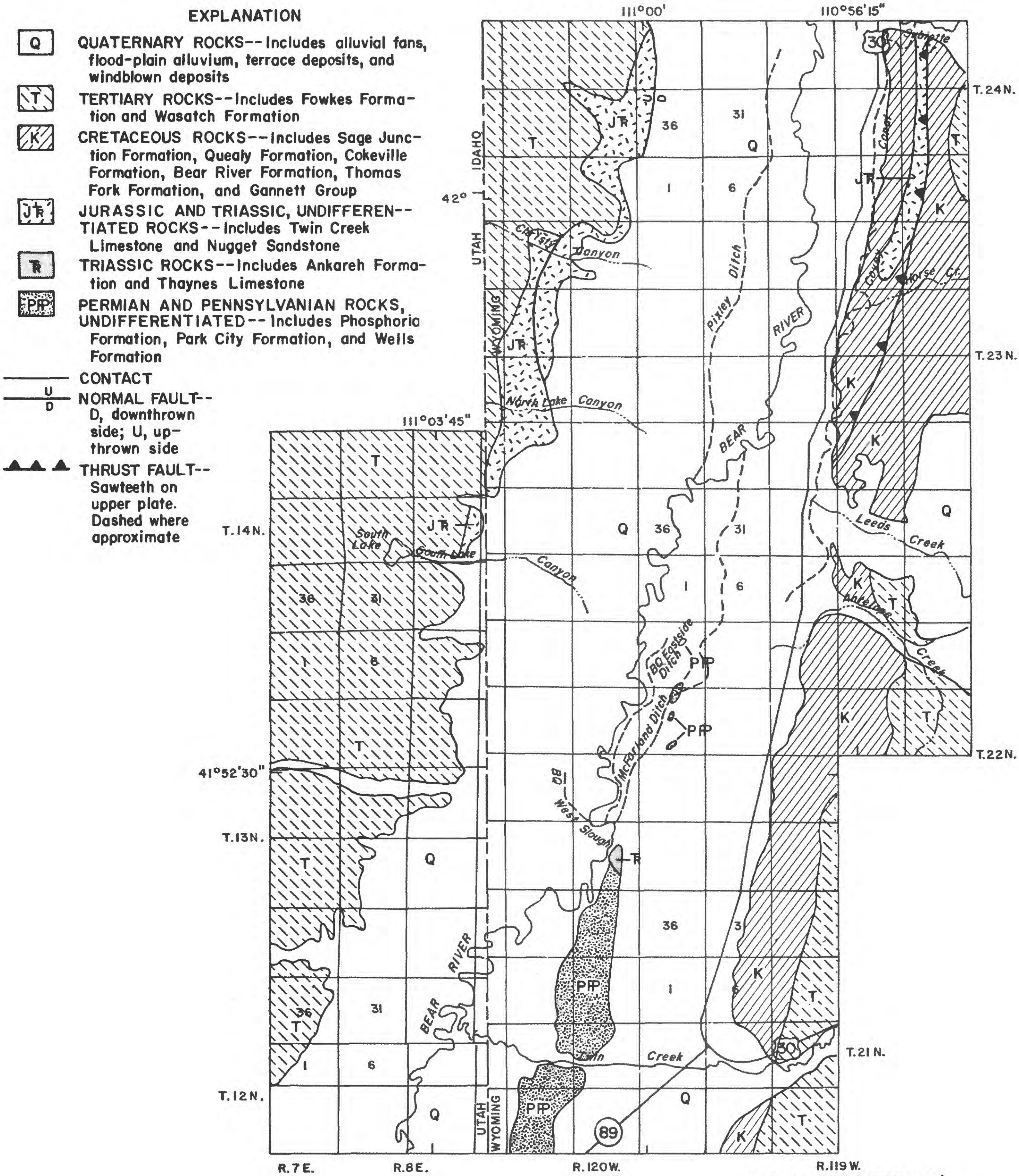


Figure 5.--Generalized surface geology in the Cokeville study area.



B

Geology modified from Lines and Gloss (1975), and Robinove and Berry (1963)

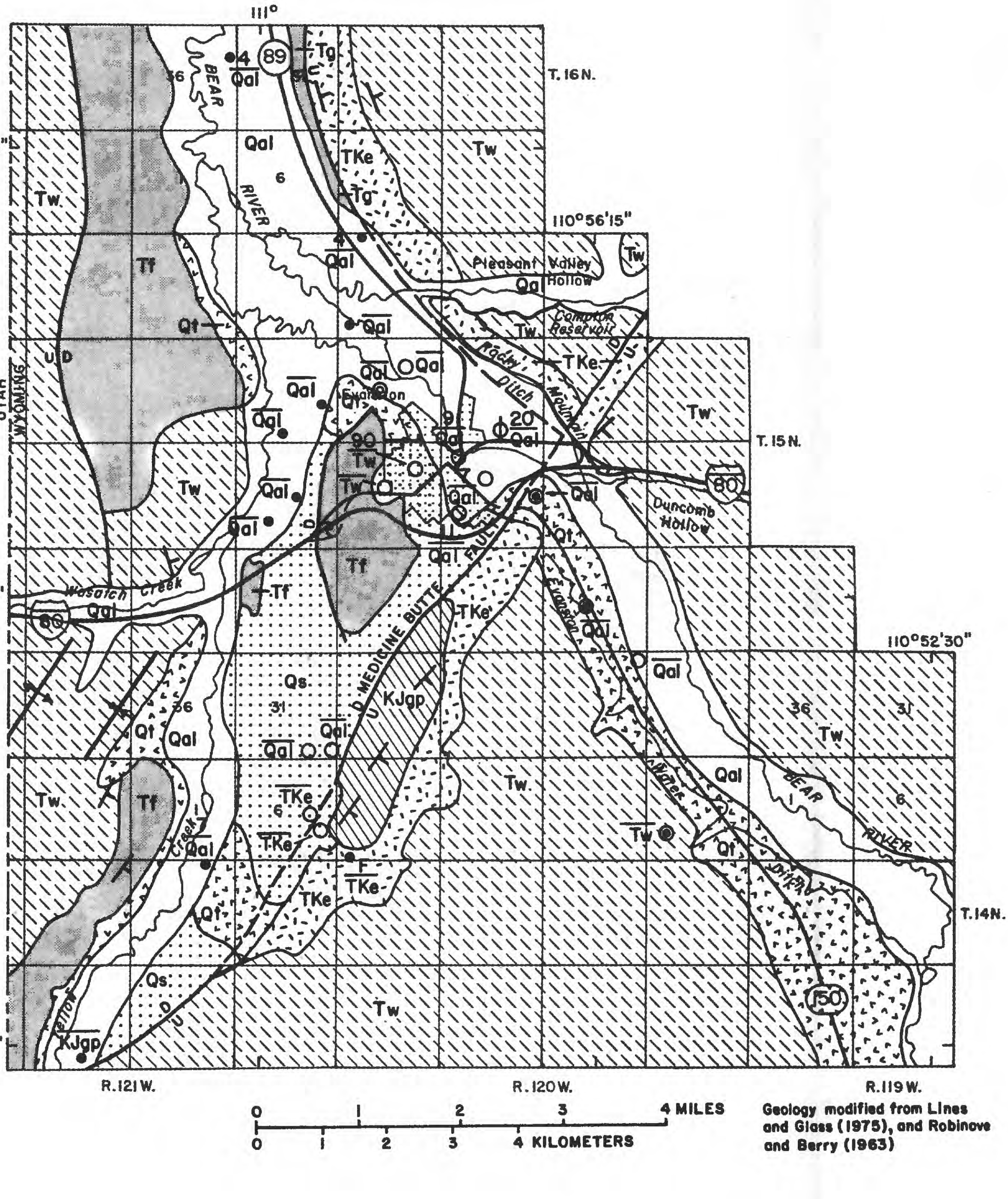


Figure 6.--Surface geology and well locations in the Evanston study area.

EXPLANATION

QUATERNARY	Qal	ALLUVIUM
QUATERNARY	Qt	TERRACE DEPOSITS
QUATERNARY	Qs	SLOPE WASH
TERTIARY	Tf	FOWKES FORMATION
TERTIARY	Tw	WASATCH FORMATION
TERTIARY / CRETACEOUS	TKe	EVANSTON FORMATION
CRETACEOUS / JURASSIC	KJgp	GANNETT GROUP, STUMP FORMATION, AND PREUSS RED BEDS, UNDIVIDED

CONTACT

U / D — NORMAL FAULT--D, downthrown side; U, upthrown side. Dashed where approximate

ANTICLINE--Showing trace of crestal plane

SYNCLINE--Showing trace of trough plane

STRIKE AND DIRECTION OF DIP OF BEDS

WELLS--Upper symbol is depth to water, in feet below land surface: blank, not measured; F, flowing. Lower symbol is geologic source of water

$\frac{9}{\text{Qal}}$ ○ Public supply

$\frac{\quad}{\text{Qal}}$ ● Industrial supply

$\frac{\quad}{\text{Qal}}$ Irrigation supply

$\frac{\text{F}}{\text{TKe}}$ • Stock, domestic, and other supply--Yields are less than 25 gallons per minute

$\frac{20}{\text{Qal}}$ ϕ Unused

Quaternary sediments form an alluvial aquifer within the Evanston study area (fig. 4). The aquifer is characterized as unconsolidated sand and gravel interbedded with silt and clay. Total thickness of alluvium is not known. No water wells completely penetrate the saturated thickness of alluvium in the Evanston study area. The alluvium is a water-table aquifer.

The Wasatch Formation of Tertiary age is a thick aquifer within the Evanston study area (fig. 6). The lithology of the aquifer is highly variable but, in general, consists of sandy clay and mudstone with irregularly bedded sandstone (Oriel and Tracey, 1970). Conglomeratic beds are common near Evanston. The thickness of the Wasatch aquifer is at least 2,000 ft within the Evanston study area (Oriel and Tracey, 1970, p. 19). The Wasatch is a water-table aquifer in most areas.

Minor aquifers in the Evanston study area are located in the Hams Fork Conglomerate Member of the Evanston Formation of Late Cretaceous and early Tertiary (Paleocene) age and in conglomerate near the base of the Gannett Group of Early Cretaceous age. The aquifers are at or near land surface in few areas (fig. 6). The Hams Fork Conglomerate Member, consisting of poorly sorted cobbles and boulders in a matrix of crossbedded sand, crops out along the upthrown side of normal faults both north and south of Evanston. The Gannett Group crops out southwest of Evanston on the upthrown side of the Medicine Butte Fault.

The alluvial aquifer and Wasatch aquifer are part of the stream-aquifer system in the Evanston study area, but other aquifers are hydraulically isolated from the Bear River. The Wasatch aquifer is in direct hydraulic connection with the alluvial aquifer. The Hams Fork Conglomerate Member of the Evanston Formation is separated vertically from the Wasatch aquifer by as much as 1,400 ft of low-permeability, gray, carbonaceous siltstone within the main body of the Evanston Formation. Aquifers in the Gannett Group are separated vertically from the Evanston Formation by relatively impermeable Lower Cretaceous rocks. Water-level changes in either the Hams Fork Conglomerate Member or the Gannett Group have no measurable effect on water levels in the Wasatch or alluvial aquifers.

HYDROLOGY OF THE COKEVILLE STREAM-AQUIFER SYSTEM

Streamflow

Discharge of the Bear River varies seasonally over several orders of magnitude (fig. 7). The large flows from April or May through June or July are the result of snowmelt runoff. The very small flows from August through September reflect the large amount of water stored or diverted upstream of the Cokeville study area. Flows from October through March primarily are the result of ground-water discharge to the river.

The amount of water diverted from the Bear River and tributaries is greatest during June and July (fig. 8), but the need to meet compact commitments typically limits the amount diverted during late July and August. During most years, compact commitments are met primarily by flow from Smiths

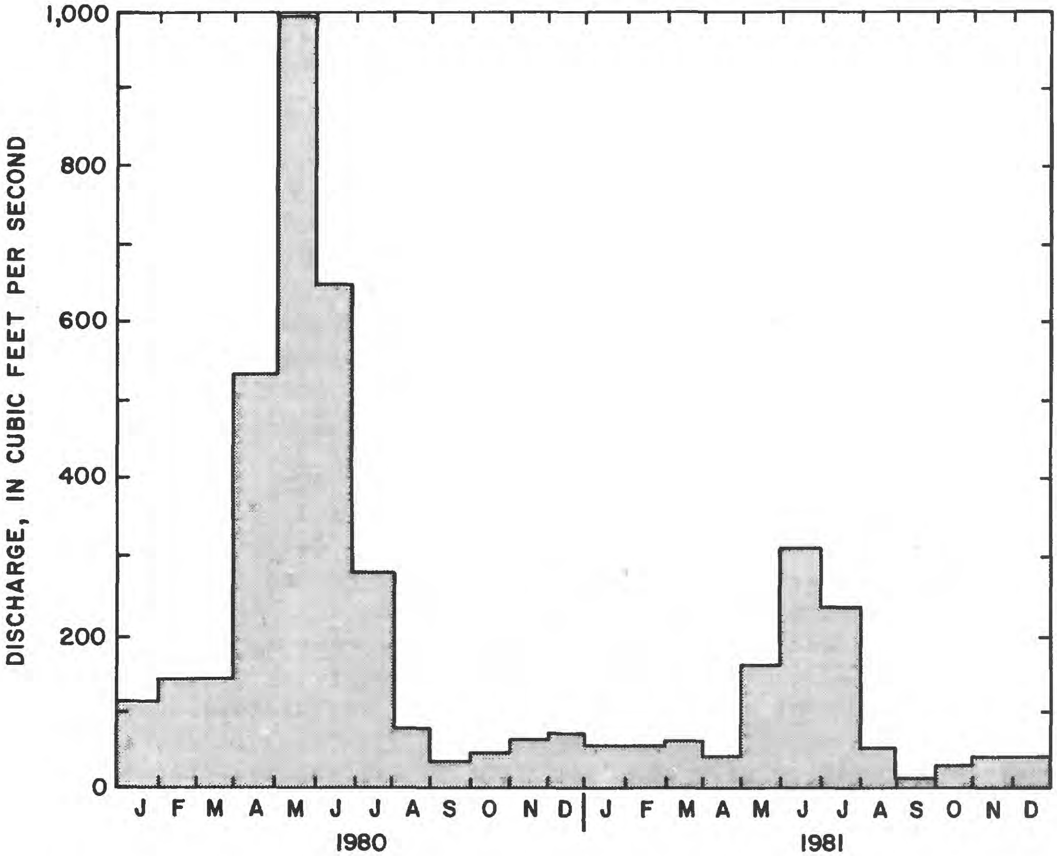


Figure 7.--Monthly mean discharge of the Bear River near Randolph, Utah, 1980 and 1981. (Streamflow-gaging station 10026500)

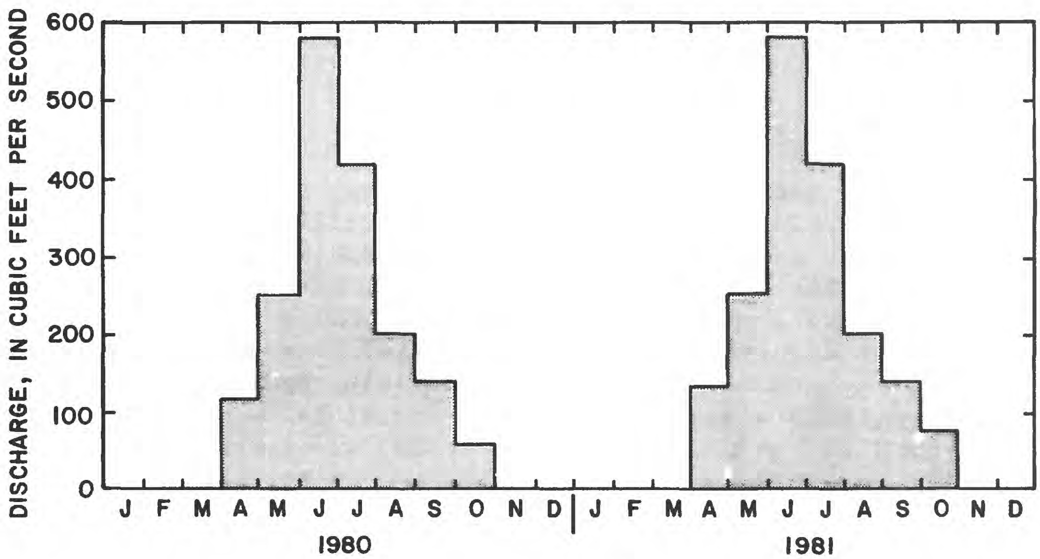


Figure 8.--Monthly mean discharge diverted in the Cokeville study area, 1980 and 1981.

Fork. During years when streamflow of Smiths Fork is below average, diversions to B.Q. West Slough, B.Q. Eastside Ditch, and Pixley Ditch are greatly limited.

Streamflow gain-and-loss studies were conducted using November and December monthly mean discharge at paired streamflow-gaging stations (table 1). Streamflow data collected during these months were used because diversions were not in operation, and sources of possible error were minimal. Stream-aquifer relations during November and December correspond to low-flow conditions. Possible errors in estimating stream-aquifer relations during other months include storage of streamflow as ice during January and February, unmeasured streamflow in small tributaries and runoff from March through June, and unmeasured irrigation return flow from July through October.

Results of the streamflow gain-and-loss study show that the Bear River gains a total of 36 ft^3/s within the Cokeville study area during low-flow periods, while Smiths Fork loses 19.4 ft^3/s. The water gained by the Bear River includes that gained by Birch Creek springs and Forgen Slough. The results were obtained by comparing monthly mean streamflow at streamflow-gaging stations and accounting for inflow from tributaries. The flow of tributaries, as well as flow from Birch Creek springs and Forgen Slough, was estimated from a series of miscellaneous discharge measurements.

To analyze streamflow gain-and-loss data it was assumed that monthly mean discharge for November and December at streamflow-gaging stations operated during different years could be compared. Data for the stations Bear River near Randolph, Utah (10026500) and Bear River at Border, Wyo. (10039500) were used to test this assumption. From 1948 through 1982 monthly mean discharge (November and December) at these two stations varied by no more than 11 percent of the long-term November and December average. When compared to streamflow-measurement errors that may be as large as 10 percent, the assumption is reasonable.

Hydraulic Characteristics of the Alluvial Aquifer

Aquifer Boundaries

The alluvial aquifer in the Cokeville study area is bounded laterally and vertically by relatively impermeable shale. The relatively impermeable shale effectively prevents ground-water movement between the alluvial aquifer and other formations. The upstream and downstream alluvial boundaries do not represent barriers to ground-water movement. These boundaries were arbitrarily selected at distances sufficiently removed from existing ground-water pumping so that any errors that occur in describing hydrologic conditions of the boundaries will have minimal effect in the vicinity of pumping. A thin layer of unsaturated rock debris (slope wash) northeast of Birch Creek springs, mapped as Quaternary rocks (fig. 5), is not part of the alluvial aquifer.

Table 1.--*Estimated steady-state ground-water leakage to streams in the Cokeville study area*

[+, increase in streamflow; -, decrease in streamflow]

Stream reach and streamflow-gaging station numbers	Ground-water leakage (cubic feet per second)
Bear River upstream boundary to Pixley Ditch (Between stations 10026500 and 10028500)	+5.5
Bear River from Pixley Ditch to Sublette Creek (Between stations 10028500 and 10029500)	+12.5
Bear River from Sublette Creek to Birch Creek springs (Between stations 10029500 and 10038000)	+7.0
Forgen Slough	+6.0
Bear River from Birch Creek springs to downstream boundary (Between stations 10038000 and 10039500)	+3.0
Birch Creek springs, 2 miles north of Cokeville	+2.0
Total, Bear River	+36.0
Smiths Fork (all branches) (Between stations 10032000 and 10035000)	-19.4

Transmissivity

Transmissivity estimates from specific-capacity data (table 2) range over an order of magnitude; the geometric mean is 11,600 ft^2/d. The geometric mean is used in this report because in many cases the geometric mean of transmissivity data is a better measure of central tendency than the arithmetic average (Freeze, 1975). No aquifer tests with observation wells have been conducted in the Cokeville study area.

Because very few transmissivity estimates from field data are available in the Cokeville study area, a flow model (described later in this report) was used to estimate the distribution of transmissivity in the alluvial aquifer. Transmissivity, estimated during model development, ranged from 2,760 to 184,000 ft^2/d.

Table 2.--*Transmissivity estimates based on specific-capacity tests*

Study area	Well number	Aquifer	Specific capacity (gallons per minute per foot)	Transmissivity (feet squared per day)
Evanston	15-120-18cd	Alluvium	150	71,500
	15-120-20ab	Wasatch	14	1,740
	15-120-21bba	Alluvium	30	14,300
	15-120-21bdd	Alluvium	30	14,300
Cokeville	23-119-18bb	Alluvium	19	7,620
	23-119-18bdb	Alluvium	31	21,400
	23-119-29cdd	Alluvium	56	26,700
	23-119-32bda	Alluvium	18	5,830
	24-119-05cc	Alluvium	56	27,200
	24-119-32cad	Alluvium	43	21,400
	25-119-20acd	Alluvium	38	18,100
	25-119-20dbc	Alluvium	5	2,400
	26-120-01ccd	Alluvium	34	6,100

Specific Yield

No field data are available to estimate specific yield of the alluvial aquifer within the Cokeville study area. In studies of alluvial aquifers in other parts of Wyoming, Crist (1975, p. 14) used an estimated specific yield of 0.23, and Glover (1983, p. 33-34) used 0.22. In this study the flow model, described later in this report, was used to estimate a value of 0.15 for specific yield of the alluvial aquifer in the Cokeville study area.

Distribution of Hydraulic Head in the Alluvial Aquifer

The steady-state potentiometric surface of the alluvial aquifer (fig. 9) shows water entering the aquifer as underflow at the upstream end of the study area and discharging to the Bear River. A second source of recharge water is leakage from tributary streams. In areas where no tributaries recharge the alluvial aquifer, potentiometric contours are nearly perpendicular to lateral bedrock boundaries, indicating no-flow barriers.

Data on differences in head with depth in the alluvium are limited; however, data from two wells indicate that vertical gradients are small. Well 23-119-20dca01 is 200 ft deep and well 23-119-20dca02 is 400 ft deep. These wells are within 30 ft of each other and have essentially the same static water level.

Hydraulic head in the alluvial aquifer varies seasonally in response to changing patterns of recharge. Water levels in wells (fig. 10) typically rise during the spring and early summer months when surface water is available for diversion and irrigation recharge is large. During the late summer months the quantity of surface water diverted is less, ground-water recharge from irrigation is small, and water levels in wells decline. Typically, water-level fluctuations from October through March are relatively small, indicating the aquifer probably is at near-steady-state condition.

A comparison of water-level measurements in wells during 1971 with measurements made in the same wells during 1982 show no long-term changes in hydraulic head of the alluvial aquifer. The differences in water-level measurements generally were less than 3 ft and showed no consistent patterns of decline or rise. Measurements during 1971 were made before most large-capacity wells were drilled in the Cokeville study area.

Ground-Water Recharge

Stream Leakage

The streamflow gain-and-loss study and potentiometric-surface map, described previously, identify tributaries, particularly Smiths Fork, as points of ground-water recharge. Recharge along Smiths Fork totals 19.4 ft^3/s. Estimates of leakage from Sublette Creek cannot be obtained from field data. The flow model, described later in this report, was used to estimate steady-state leakage of 0.7 ft^3/s from Sublette Creek.

The streamflow gain-and-loss study identifies long-term steady-state recharge to the alluvial aquifer but does not identify seasonal changes in recharge from streams. The gain-and-loss study used streamflow data for October through December when ground-water levels are at steady state. Changes in stream stage that occur during the spring and summer months result in rates of stream leakage that differ from those given by the gain-and-loss study.

Irrigation

Water diverted from the Bear River for irrigation from May through September usually exceeds the consumptive-use requirements of crops, and some of this water enters the alluvial aquifer as recharge. The amount of irrigation recharge depends on the amount of water diverted from streams, the consumptive-use requirements of the irrigated crops, total irrigated acreage, moisture-storage characteristics of the soil, and the amount of water that returns to the stream as surface runoff (called return flow in this report). Data are available to estimate the amount of water diverted (Bear River Commission, 1981; 1983), the consumptive-use requirement of crops (Trelease and others, 1970), and irrigated acreage (fig. 3). Because soil-characteristics and return-flow data are unavailable, irrigation recharge was estimated during flow-model development. Although varying throughout the irrigation season, irrigation recharge frequently exceeds 200 ft^3/s. Details of model-estimated recharge from irrigation are given later in this report.

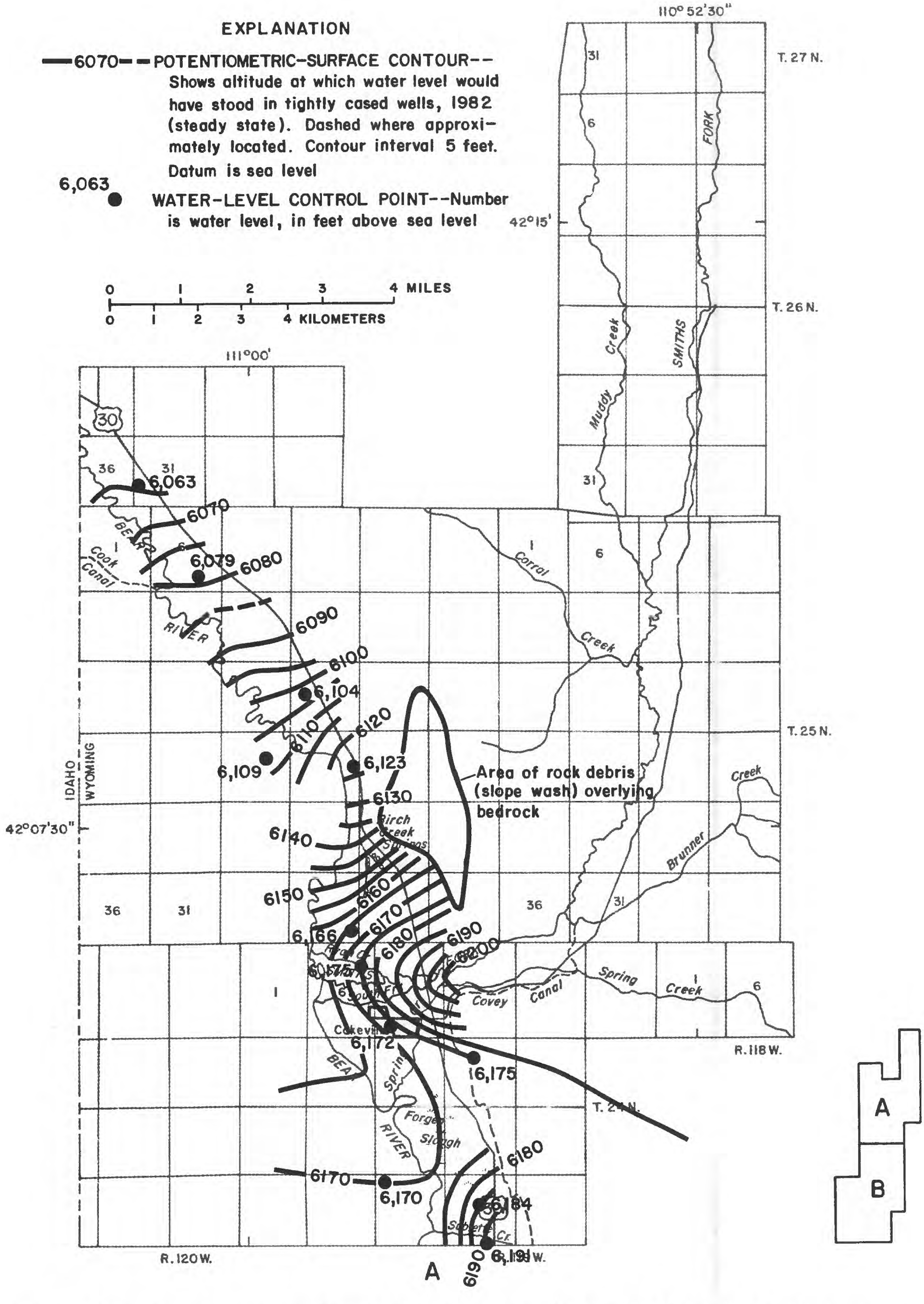
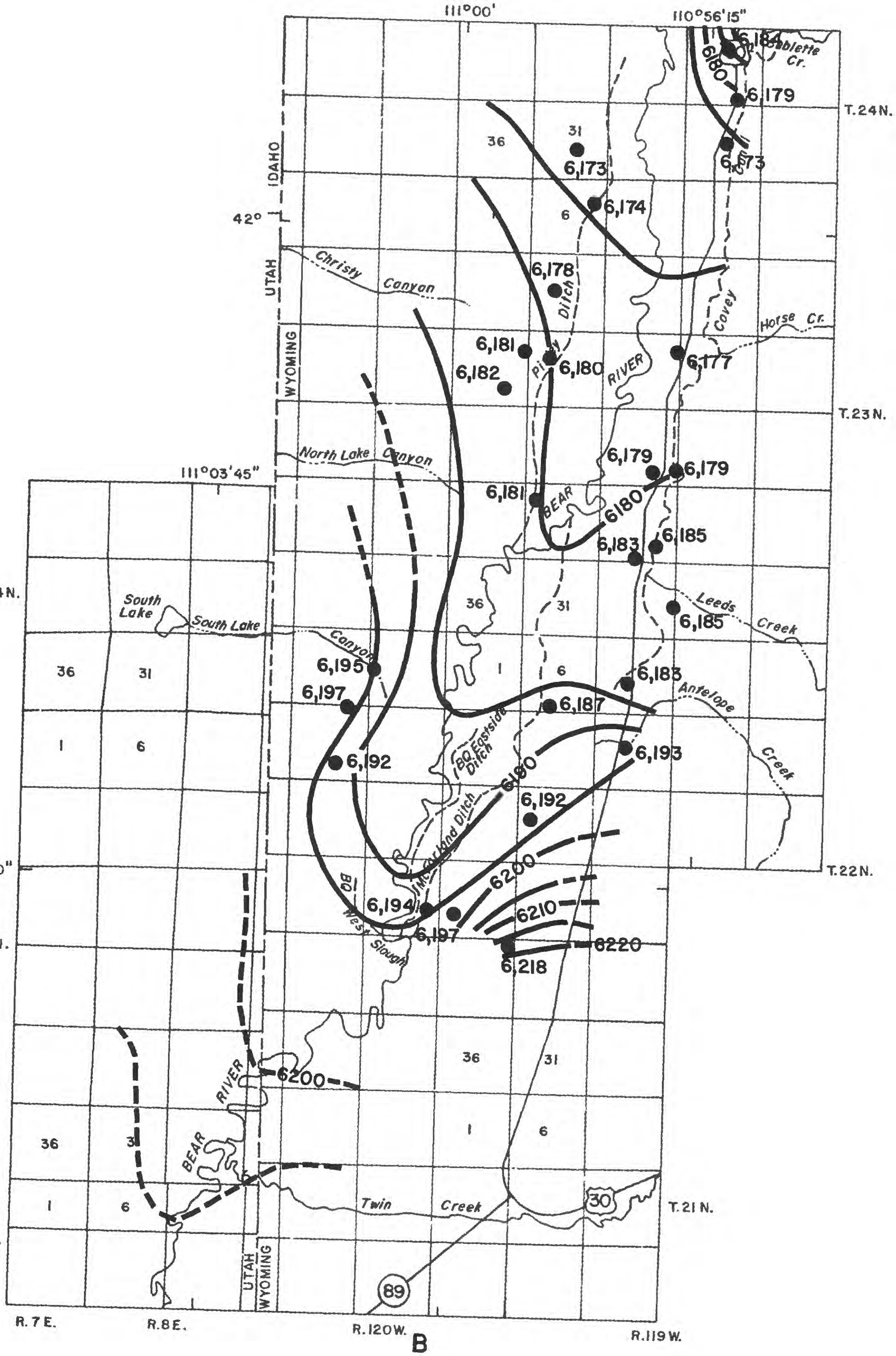


Figure 9.--Potentiometric surface of the alluvial aquifer in the Cokeville study area.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110°56'15"
111°03'45"
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IDAHO
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T.14N.
T.13N.
T.12N.
R.7E.
R.8E.
R.120W.
R.119W.
Sublette Cr.
Christy Canyon
Piney Ditch
Covey
Horse Cr.
BEAR RIVER
North Lake Canyon
South Lake
South Lake Canyon
Leeds Creek
Antelope Creek
BQ Eastside Ditch
McCorkland Ditch
BQ West Slough
Twin Creek
30
89
6180
6200
6210
6220
6,184
6,179
6,173
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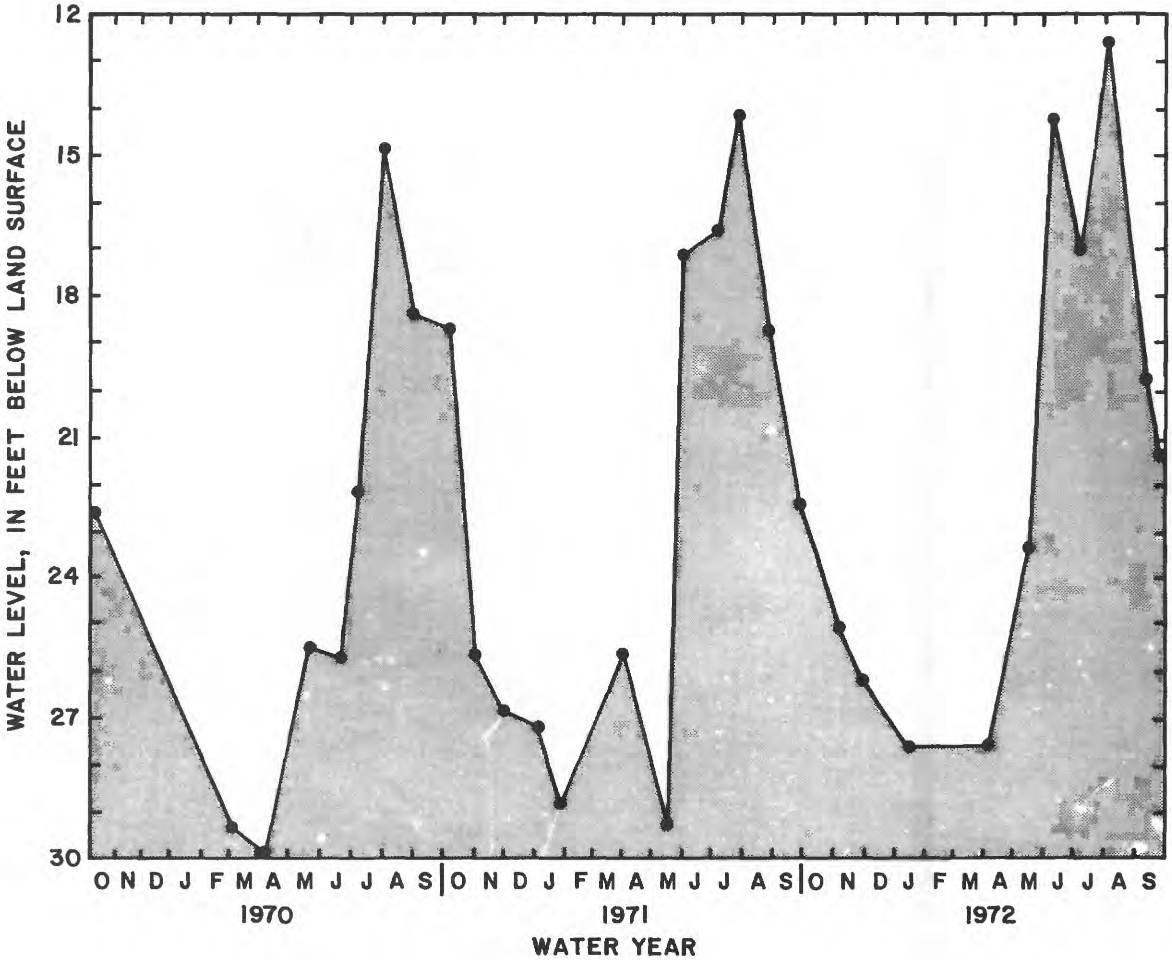


Figure 10.--Water levels in well 23-119-32bda during water years 1970-72. Timing of water-level rises corresponds to timing of recharge from irrigation; however, the magnitudes of the fluctuations are affected by recharge from water in a nearby ditch and by water-level declines due to pumping of this well.

Underflow

The potentiometric-surface map (fig. 9) shows water entering the alluvial aquifer as underflow at the upstream study boundary and where alluvium along tributaries meets the alluvium of the Bear River. The flow model, described later in this report, was used to estimate the amount of underflow. Approximately 18.5 ft^3/s enters the study as underflow along the upstream model boundary while 14.9 ft^3/s enters as underflow from alluvium along tributaries.

Ground-Water Discharge

Streams and Springs

The streamflow gain-and-loss study, described previously, identifies the Bear River and springs near the Bear River as points of ground-water discharge. Results of the study, which describe steady-state conditions, show the Bear River gaining 18 ft^3/s while Forgen Slough gains 6 ft^3/s and Birch Creek springs gains 2 ft^3/s.

Evapotranspiration

Evapotranspiration, primarily by willows and grasses that obtain water directly from the water table, is a significant type of ground-water discharge during the summer. The amount of water that discharges as evapotranspiration depends upon the consumptive-use requirements of the plants and the depth to water. Evapotranspiration is maximum when the water table is at land surface and decreases as depth to water increases. Lenfest (1987) suggests that there is essentially no evapotranspiration in Wyoming at depths to water greater than 10 ft.

Evapotranspiration during the summer intercepts some of the ground water that would otherwise discharge as stream leakage. Evapotranspiration by willows and other phreatophytes occurs in the Cokeville study area adjacent to the Bear River; areas of significant evapotranspiration are shown in figure 3. Streamflow gain-and-loss studies show that the Bear River is a point of ground-water discharge.

The amount of ground water that discharges as evapotranspiration was estimated during development of the flow model. The flow model simulates evapotranspiration as a linear function of depth to water. For the areas delineated in figure 3, the total amount of ground water calculated to be discharging as evapotranspiration varied throughout the simulation (for 1980 and 1981), but averaged 18.7 ft^3/s.

Pumpage

Large-capacity irrigation wells are points of seasonal ground-water discharge, but pumpage does not affect the long-term distribution of hydraulic head in the alluvial aquifer. Although well yields as large as 1,000 gal/min are common, few wells have been drilled in the alluvial aquifer. The comparison of water-level measurement in 1971 and 1982, discussed previously, identified no long-term declines in water levels due to pumping.

Direct measurement of pumpage was not possible, but the amount of water withdrawn by wells was estimated by determining consumptive-use requirements of crops (Trelease and others, 1970) and acreage irrigated by well water (fig. 3) from 1980 to 1982. When a well was used as a supplemental source of water for land irrigated primarily by surface-water diversions, part of the consumptive-use requirements of the crop was met by surface water. The product of consumptive-use requirement and irrigated acreage is estimated pumpage (table 3).

Underflow

Ground water leaves the Cokeville study area by underflow at the downstream end of the study area. Discharge by underflow was estimated during flow-model development to be 17.8 ft^3/s.

HYDROLOGY OF THE EVANSTON STREAM-AQUIFER SYSTEM

Streamflow

Although streamflow-gaging stations are not operated in the Evanston study area, streamflow characteristics of the Bear River probably are similar to those described in the Cokeville study area. Large flows from April through June are the result of snowmelt runoff, and small flows during the rest of the year reflect stream diversions and ground-water discharge.

Streamflow gain-and-loss studies using seepage-run data collected on April 6, 1981, give a quantitative description of stream-aquifer relations (table 4). The seepage run was conducted during the low-flow period before significant snowmelt had occurred. No ice was noted along the stream channels. Unusually large streamflow during late 1981 and 1982 prevented additional seepage runs. Results of the seepage run indicate that the Bear River gained 29.4 ft^3/s and Yellow Creek lost 16.3 ft^3/s within the Evanston study area.

Hydraulic Characteristics of Aquifers

Aquifer Boundaries

The surface-geology map (fig. 6) shows the alluvial aquifer and Wasatch aquifer in direct hydraulic connection in several places. The largest area of the Wasatch aquifer in hydraulic connection with the alluvial aquifer is southeast of the Medicine Butte Fault; other areas are near the mouth of Wasatch Creek and near Evanston, where the Wasatch aquifer is overlain by a thin section of the Fowkes Formation of Eocene age, or unsaturated Quaternary slope wash.

Table 3.--*Estimated total pumpage from irrigation wells in the Cokeville study area*

Well number	Total pumpage (acre-feet)				
	May	June	July	August	September
23-119-32bd	25.8	52.8	73.1	55.2	29.4
23-120-13db	19.5	39.9	55.2	43.0	22.2
23-120-25ca	34.2	70.1	97.0	75.7	39.0
24-119-05cc	5.0	10.2	14.1	11.0	5.6
24-119-31db	22.8	46.7	64.7	50.4	26.0
25-119-17cb	23.8	48.8	67.5	52.6	27.1
25-119-17db	7.7	16.8	22.5	17.0	9.0

Table 4.--*Estimated ground-water leakage to streams in the Evanston study area*

[+, increase in streamflow; -, decrease in streamflow]

Stream reach	Ground-water leakage (cubic feet per second)
Bear River from Evanston Water Ditch to Duncomb Hollow	+19.0
Bear River from Duncomb Hollow to sec. 1, T. 15 N., R. 121 W.	+10.4
Yellow Creek from sec. 1, T. 14 N., R. 121 W. to mouth of creek	-16.3

Relatively impermeable rocks at land surface, usually on the upthrown sides of normal faults, form lateral barriers to ground-water movement between the alluvial and Wasatch aquifers. Cretaceous rocks along the Medicine Butte Fault form one lateral barrier. Outcrops of the main body of the Evanston Formation, associated with a fault along the alluvium-bedrock contact, effectively prevent flow between the alluvial and Wasatch aquifers in the northeastern part of the study area. The Wasatch aquifer west of Yellow Creek is separated from the alluvial aquifer by lake deposits of the Fowkes Formation; the lake deposits have small hydraulic conductivity.

Transmissivity

Transmissivity estimates from specific-capacity data (table 2) show the alluvial aquifer to be similar in transmissivity to alluvium in the Cokeville study area, while transmissivity of the Wasatch aquifer is at least one order of magnitude less than that of the alluvium. No pumping tests with observation wells have been conducted in either the alluvial or Wasatch aquifer. Therefore, the accuracy of the estimates given in table 2 is not known, nor are the estimates necessarily representative of the entire study area. The range of estimates probably is reasonable for transmissivity of the two aquifers.

Specific Yield

No field data are available to estimate specific yield, but values for alluvium probably are similar to that given previously for the Cokeville study area (0.15). Values for the Wasatch aquifer probably are smaller. Based on greater shale content for the Wasatch aquifer, specific yield may range from 0.03 to 0.1.

Distribution of Hydraulic Head in Aquifers

Insufficient hydraulic-head data are available to map reliable potentiometric surfaces for aquifers in the Evanston study area. Water levels in fewer than 20 wells in the alluvial aquifer and 5 wells in the Wasatch aquifer have been reported (fig. 6). With the complex distribution of aquifer boundaries in the study area, accurate potentiometric-surface mapping is not possible. All measurements show water-table conditions in the alluvial and Wasatch aquifers.

A comparison of water-level measurements in wells during 1971 with measurements in the same wells during 1982 show no long-term changes in hydraulic head of the alluvial aquifer. Water-level data are not available to identify long-term trends in the Wasatch aquifer.

Water-level measurements from wells show that water in deep parts of the Wasatch aquifer is in direct hydraulic connection with water in the alluvial aquifer, while water at shallow depth often is part of local, perched flow systems. Discharge from perched systems usually occurs as seepage along hillsides. Wells 14-120-2cbd and 15-120-20ab are examples of wells that pump water in hydraulic connection with the Bear River. Both are deep wells open to sandstone conglomerate of the Wasatch aquifer. Both have static water levels that are approximately the same as stream level. Shallow wells open to perched flow systems in the Wasatch aquifer are not shown in figure 6.

Ground-Water Recharge

The streamflow gain-and-loss study, described previously, identifies Yellow Creek as a source of ground-water recharge. Ground-water recharge from Yellow Creek within the Evanston study area was estimated to be 16.3 ft^3/s during low-flow periods. During the spring and early summer months streamflow is large; because flows in the Bear River and in Yellow Creek and other tributaries are deeper (greater head), ground-water recharge probably is greater than at other times of the year. No reliable measurements of recharge have been made during periods of large streamflow.

Water diverted from the Bear River from May through September usually exceeds the consumptive-use requirements of irrigated plants, and part of the excess water enters the alluvial aquifer as recharge. Data are available to estimate the amount of water diverted (Bear River Commission, 1981; 1983), but reliable estimates of irrigation recharge cannot be made. Flow-model development, used in the Cokeville study area to estimate recharge, is not possible in the Evanston study area because of the limited amount of hydraulic-head data and other data available for model calibration.

Underflow at upgradient study boundaries is an important source of ground-water recharge, but quantitative estimates of underflow cannot be made with existing data. Estimates of underflow could be made if the potentiometric surface of both alluvial and Wasatch aquifers could be mapped and if transmissivity of both aquifers were known. As discussed previously, neither potentiometric-surface maps nor reliable transmissivity estimates are available in the Evanston study area. Underflow in the alluvial aquifer occurs at the southern study boundary along the Bear River and Yellow Creek. Areas of underflow in the Wasatch aquifer cannot be identified.

Recharge from precipitation is an additional source of recharge in upland areas of the Wasatch aquifer. Recharge probably is greatest in the spring when the snowpack melts. While recharge from snowmelt also occurs in the alluvial aquifer, the increased snow depth at higher altitudes increases the likelihood of significant recharge. No quantitative estimates of recharge from precipitation are possible with existing data.

Ground-Water Discharge

The streamflow gain-and-loss study, described previously, identifies the Bear River as a major point of ground-water discharge. A total of 29.4 ft^3/s discharges from the alluvial aquifer into the Bear River. During summer months when recharge from irrigation is large, discharge to the Bear River probably increases.

Evapotranspiration, primarily from willows and grasses that grow where the water table is at or near land surface, is a significant type of ground-water discharge during the summer. While quantitative estimates of discharge by evapotranspiration were made in the Cokeville study area during flow-model development, the inability to develop a reliable flow model of the Evanston study area has prevented quantitative estimation of evapotranspiration. Nevertheless, the large area where ground water is at or near the land surface, as indicated by the numerous ponds and drains in the study area, shows evapotranspiration from the water table to be an important form of ground-water discharge.

The City of Evanston and nearby unincorporated communities obtain ground water by pumping large-capacity wells that have been drilled in the alluvial and Wasatch aquifers. Accurate pumpage records are available for wells maintained by the City of Evanston (table 5), but pumpage records were not available for wells maintained by unincorporated communities. The amount of water pumped varies seasonally. The municipal wells are rarely used during the winter months. Pumpage from other public-supply wells continues throughout the year, but total pumpage is small.

Ground water in the alluvial aquifer discharges by underflow at the northern boundary of the study area. Quantitative estimates of underflow cannot be made because reliable potentiometric-surface maps and transmissivity estimates are not available.

SIMULATION OF THE STREAM-AQUIFER SYSTEM IN THE COKEVILLE STUDY AREA

Movement of water within the stream-aquifer system of the Cokeville study area was simulated mathematically for two reasons:

1. To improve estimates of aquifer properties such as transmissivity, and flow parameters such as ground-water recharge, that are poorly described by field data; and

2. to evaluate the hydrologic effects of existing ground-water development on streamflow.

Model Theory

The computer model used to simulate flow in the stream-aquifer system of the Cokeville study area is based on differential equations that describe one-dimensional kinematic flow in streams and two-dimensional Darcian flow in aquifers. Leakage of water between streams and aquifers is described by Darcy's equation. Because these differential equations cannot be solved directly, finite-element approximations of the equations are used. Physical properties, such as hydraulic head and transmissivity, that may vary in two dimensions, are approximated with an assemblage of flat triangular surfaces. In areas where the slope of the real surface changes rapidly, smaller triangles are used to maintain an accurate approximation. Triangular planes in

Table 5.--*Pumpage from Evanston municipal wells, 1981-83*

Month	Amount of water pumped (acre-feet) 1981	1982	1983
May	0.0	2.7	16.3
June	67.1	156	97.6
July	134	160	161
August	124	95.0	55.9
September	12.0	5.0	.0
October	1.0	.0	.0
Total	338	419	331

the finite-element method are defined by the position of their corners, called nodes. Each triangle is called an element. The computer program that solves the finite-element equations is described in detail by Glover (1988).

All hydrologic characteristics of the stream-aquifer system are simulated within the model using a finite-element grid (fig. 11). Streamflow characteristics are mapped along the sides of elements. Characteristics that are areal in nature, such as transmissivity and ground-water recharge from irrigation, are treated is uniform within each element. Factors such as well pumpage that apply only to a point are identified by values at nodes. The distribution of hydraulic head is described by values at nodes. The value of hydraulic head within an element is simulated(by a linear approximation of nodal head values.

Application of the Model to Steady-State Flow

Hydraulic heads within the alluvial aquifer remain essentially constant during the winter months; although seasonal changes in hydraulic head are common, the water level in wells returns to similar values each winter. Therefore, ground-water movement within the aquifer system occurs under steady-state conditions during the winter months, with ground-water recharge equalling ground-water discharge. There is no change in the amount of ground water stored in the aquifer.

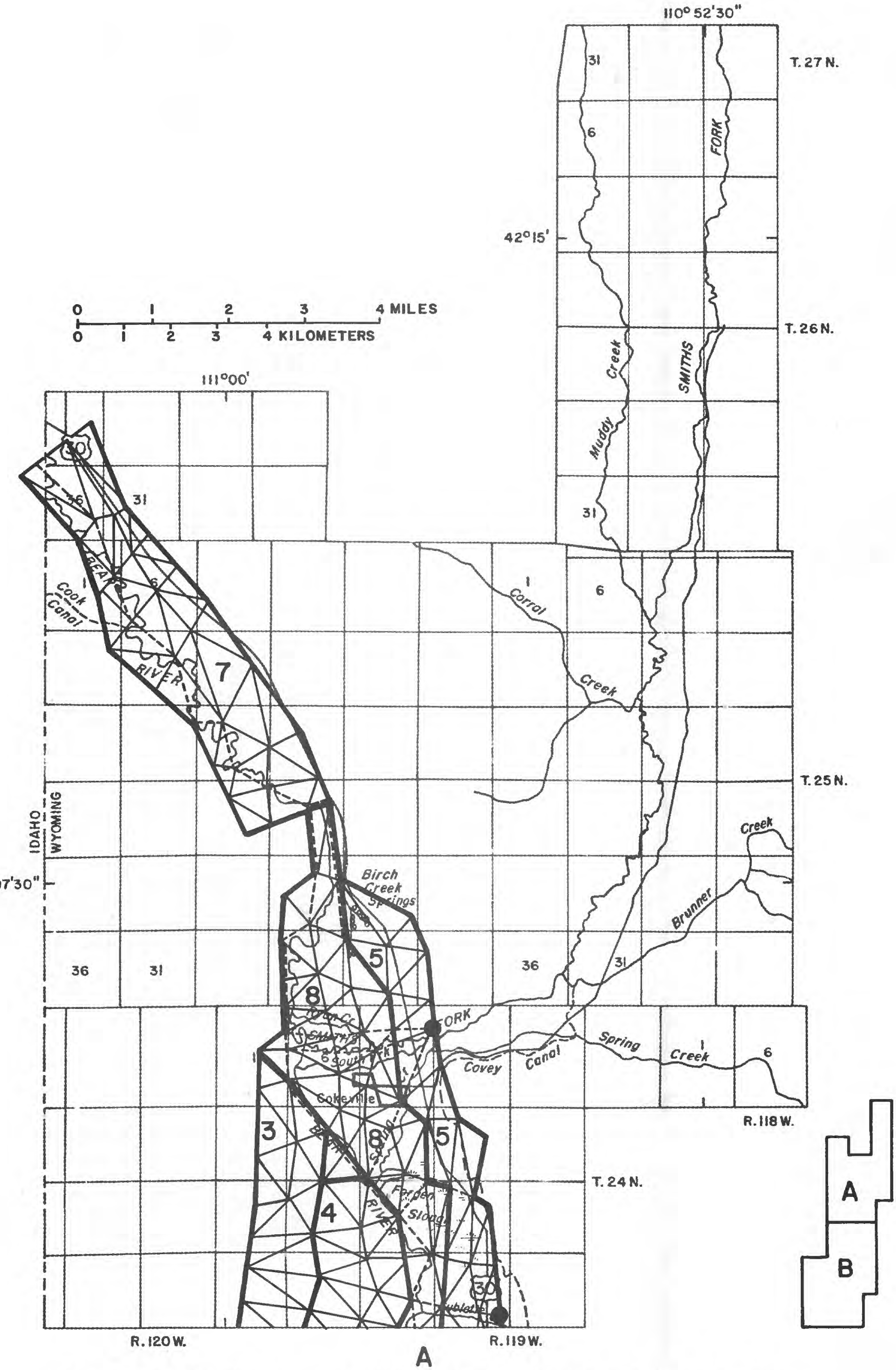


Figure 11.--Finite-element grid used to simulate the alluvial aquifer in the Cokeville study area.

EXPLANATION

AQUIFER ZONE AND NUMBER--Transmissivity is simulated by a single value within each zone

FINITE ELEMENT--Two-dimensional aquifer element used to model the alluvium

-------- STREAM FINITE ELEMENT--One-dimensional stream element located along the side of an aquifer finite element

BOUNDARY CONDITIONS

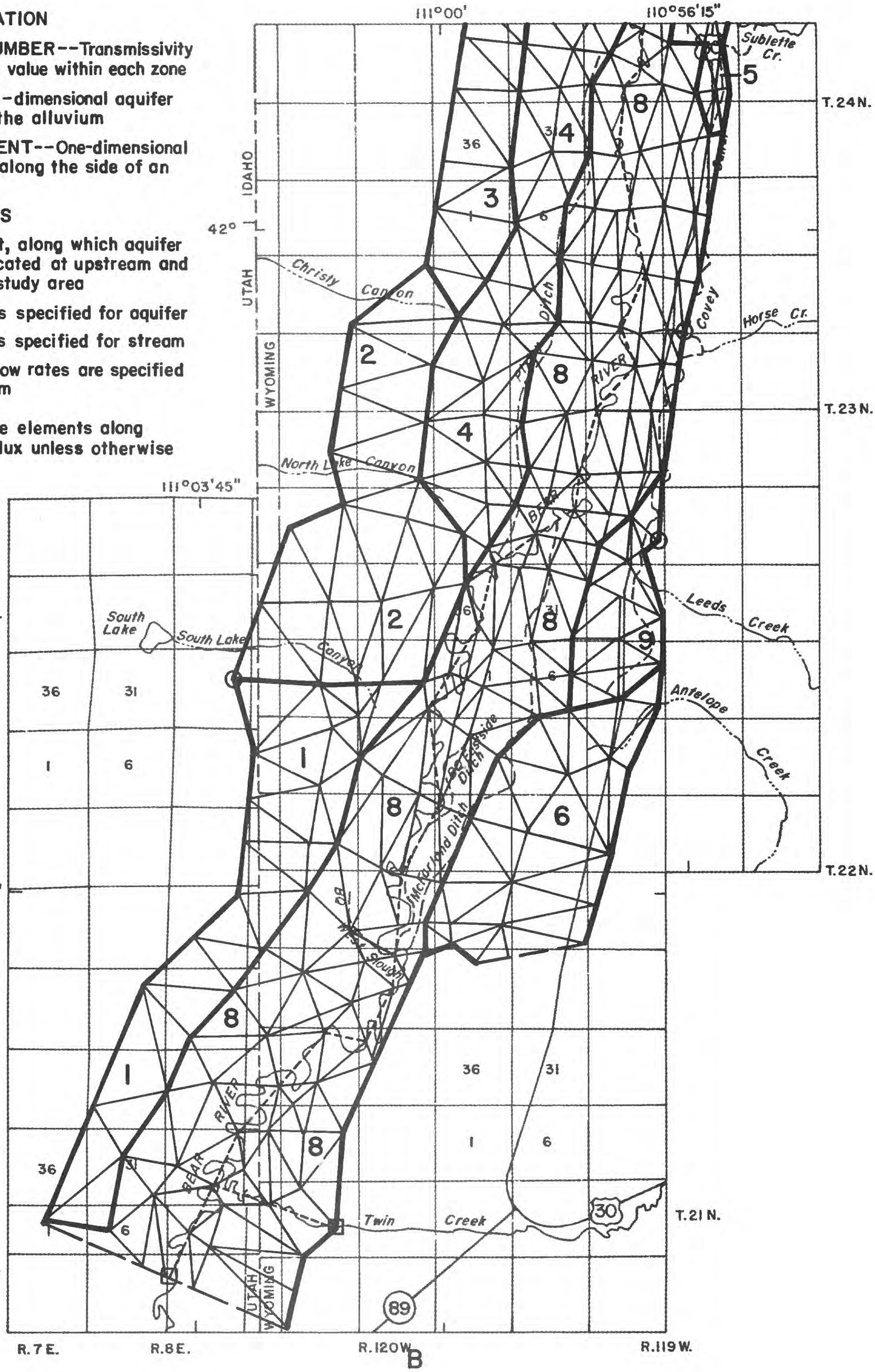
— — — Side of aquifer element, along which aquifer head is specified. Located at upstream and downstream ends of study area

○ Node where flow rate is specified for aquifer

□ Node where flow rate is specified for stream

● Node where separate flow rates are specified for aquifer and stream

(Note.--Sides of finite elements along boundaries are zero-flux unless otherwise designated.)



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The steady-state characteristic of the alluvial aquifer is used as a beginning point for simulating flow in the stream-aquifer system because steady-state flow is less complex to simulate than transient flow. Time derivatives, and model parameters associated with the time derivatives are not a part of the differential equations that describe steady-state flow. As a result, estimates of specific yield, seasonal ground-water recharge and discharge, and stream-channel characteristics that describe kinematic streamflow are not considered in the steady-state simulation.

Only those forms of ground-water recharge and discharge that influence the steady-state distribution of hydraulic head are considered in the steady-state simulation. Steady-state simulation includes ground-water recharge and discharge in the forms of stream leakage and underflow. Seasonal recharge and discharge, such as irrigation recharge and evapotranspiration, cause changes in the distribution of head and are not included.

Initial Estimates of Model Parameters

Initial estimates of model parameters are summarized in this section of the report. In some cases, field data are available to estimate model parameters accurately. In other cases, very little or no field data are available, and initial estimates are unreliable. Model calibration was used to revise unreliable estimates.

External aquifer boundaries (fig. 11) were simulated as having either a specified hydraulic head or zero flux. Specified-head boundaries were used to simulate underflow at upstream and downstream model boundaries. Boundary values of hydraulic head were obtained from the potentiometric-surface map of the alluvial aquifer (fig. 9). Specified-head boundaries were located sufficiently distant from anticipated areas of well pumpage to minimize the chance of drawdown cones intercepting the boundaries. Zero-flux boundaries were located where the alluvial aquifer is in contact with bedrock formations.

The potentiometric-surface map of the alluvial aquifer (fig. 9) shows water entering the simulated aquifer as underflow where alluvium along tributaries of the Bear River is in contact with alluvium of the Bear River; however, the rates of underflow cannot be estimated from field data. Tributaries where underflow occurs are Leeds, South Lake, and Sublette Creeks, and Smiths Fork. Underflow from tributary alluvium was simulated as ground-water recharge. Each point of recharge was initially estimated to be 1.0 ft^3/s and was treated as a calibration parameter.

Very few field data are available to estimate transmissivity within the alluvial aquifer, but field mapping of lithofacies within the aquifer was used to identify zones of similar transmissivity (fig. 11). Transmissivity within each zone was simulated by a single value. Although the distribution of transmissivity zones is reliable, the value within each zone is very poorly described by field data. Therefore, a uniform initial estimate of transmissivity equal to the geometric mean of values given in table 2 (11,600 ft^2/d) was used in the simulation. Transmissivity within each zone was treated as a calibration parameter.

Values for hydraulic head of the aquifer were obtained from figure 9 for specified-head boundaries and finite-element nodes that correspond to observation wells. The solution algorithm used to solve the finite-element equations (Glover, 1988) required initial estimates of head only along specified-head boundaries. The procedure used to calibrate the steady-state simulation, discussed in the following section of this report, used measurements of hydraulic head at observation wells. Initial estimates of hydraulic head at other finite-element nodes were not needed.

Hydraulic heads of water in the Bear River, Smiths Fork, and Sublette Creek during low-flow periods were treated as known boundary conditions for steady-state simulation. Head at each stream node was obtained from topographic maps at a scale of 1:24,000. The contour interval for topographic maps of the Cokeville study area is 5 ft.

Steady-state rates of flow between the alluvial aquifer and streams in the Cokeville study area were used to estimate streambed-leakance coefficients for stream reaches identical to those used in the streamflow gain-and-loss study (table 1). By applying Darcy's equation for steady-state flow across a semipermeable streambed of known length, and assuming a head difference of 1 ft between stream and aquifer, initial estimates of streambed-leakance coefficient were obtained that are consistent with reliable streambed-leakage data. However, the assumption of a head difference of 1 ft is based on water-level measurements at very few locations, and the streambed-leakage data, due to probable error in measuring stream discharge during low-flow periods, are accurate only within 20 percent. Therefore, streambed-leakance coefficients were treated as calibration parameters.

Calibration of Model

Initial estimates of model parameters did not result in a hydrologically reasonable description of steady-state flow in the Cokeville stream-aquifer system. Specifically, hydraulic head at observation wells and rates of stream leakage calculated during the initial steady-state simulation did not correspond well to field data. This initial result was expected because initial estimates of transmissivity and ground-water recharge were not based on reliable field data.

Because initial estimates of some model parameters were unreliable, the estimates were varied until model-calculated hydraulic head at observation wells and rates of stream leakage correctly simulated measured water levels and stream leakage. This process of varying model parameters to more accurately simulate reliable data is called calibration. Model parameters were varied within physically reasonable limits. Methods of calibration include trial-and-error variations of model parameters and statistical optimization.

A statistical algorithm, based on nonlinear regression techniques (Cooley, 1982), was used to calibrate the steady-state simulation of the Cokeville stream-aquifer system. The algorithm was used to compute estimates of model parameters that minimized the squared difference between calculated and measured hydraulic head. Model parameters are treated within the algorithm as coefficients of a regression equation that, upon solution, is equivalent to the finite-element equation of ground-water flow. Estimates of model parameters may be constrained within the algorithm if adequate information is available prior to simulation. The constraint is included within the algorithm by providing both an initial estimate of the parameter value and a value for the error variance of the initial estimate.

Results of the statistical calibration algorithm include estimates of model parameters and a measure of reliability, called the standard error, for each parameter. Each standard error has units equal to the units of the corresponding parameter. For example, the standard error of transmissivity has units of feet squared per day. Qualitatively, a small standard error, relative to the parameter estimate, indicates that the estimate is reliable. A small standard error also usually indicates that the simulation results are relatively sensitive to changes in the parameter estimate.

Aquifer transmissivity and ground-water recharge were treated as unconstrained calibration parameters within the steady-state simulation of the Cokeville stream-aquifer system, while streambed-leakance coefficient was treated as a constrained calibration parameter. Transmissivity and recharge were treated as unconstrained because initial estimates were unreliable, and sufficient field data were unavailable to determine an error of variance for the initial estimates. Streambed-leakance coefficients were treated as constrained parameters because initial estimates were reasonable and field data were available to determine an error variance. Sources of error in estimating streambed-leakance coefficients included errors in measured rates of stream leakage (20 percent of values given in table 1) and errors in measured head difference between stream and aquifer (1 ft). The error variance of streambed-leakance coefficient, expressed as a coefficient of variation, is 0.4234. This value greatly limits the range of reasonable values that can be used for streambed-leakance coefficient during calibration.

Results of Simulation

Estimates of transmissivity, streambed-leakance coefficient, and ground-water recharge are accurately described by model calibration (table 6). In general, standard errors of model parameters are small in comparison to calibrated estimates. However, model parameters that correspond to areas of no hydraulic-head data or areas of contradictory head data were estimated with larger standard errors than areas with adequate and consistent data.

The distribution of hydraulic head and rates of stream leakage calculated during the simulation compare favorably with field data (fig. 12, and tables 1 and 7). The difference between calculated and measured hydraulic heads did not exceed 4.5 ft at any well location and, in most cases, was less than 3.0 ft. No trends were apparent in map plots of these differences. The standard error of estimated heads, a measure of the overall fit of the model, was 2.4 ft. This error is approximately the same as the errors in land-surface altitudes that were obtained from topographic maps with contour intervals of 5 ft. The differences between rates of stream leakage estimated from field

Table 6.--*Results of steady-state model calibration in the Cokeville study area*

Aquifer zone, stream reach, or stream	Calibrated value	Standard error	95-percent confidence limits
	Transmissivity, in feet squared per day		
Zone 1	2,760	793	1,660
Zone 2	2,760	793	1,660
Zone 3	184,000	12,800	26,700
Zone 4	184,000	12,800	26,700
Zone 5	6,310	1,340	2,820
Zone 6	1,460	1,080	2,260
Zone 7	133,000	4,260	8,930
Zone 8	184,000	12,800	26,700
Zone 9	184,000	12,800	26,700
	Streambed-leakance coefficient, in feet per day per foot		
Bear River between upstream boundary of study area and Pixley Ditch	0.141	0.0183	0.0383
Bear River between Pixley Ditch and Sublette Creek	.521	.0154	.0323
Bear River between Sublette Creek and Birch Creek spring	.223	.0066	.0137
Bear River between Birch Creek spring and downstream boundary of study area	.358	.0408	.0853
Sublette Creek	2.18	.1319	.2760
Smiths Fork	2.02	.0321	.0673
	Ground-water flux boundary, in cubic feet per second		
South Lake Creek	1.29	0.252	0.527
Leeds Creek	5.88	.374	.783
Sublette Creek	3.55	.215	.450
Smiths Fork	4.19	.066	.139

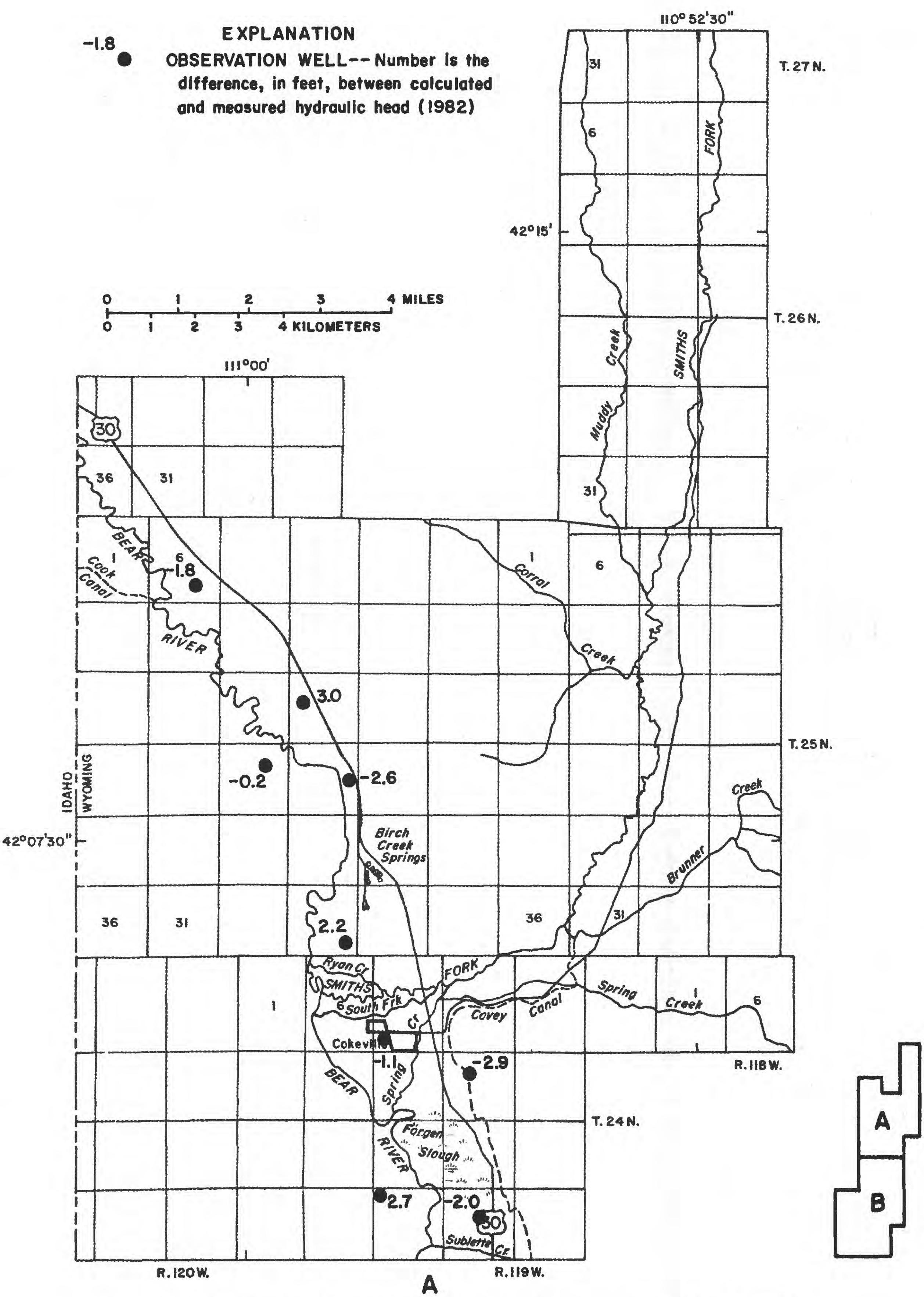

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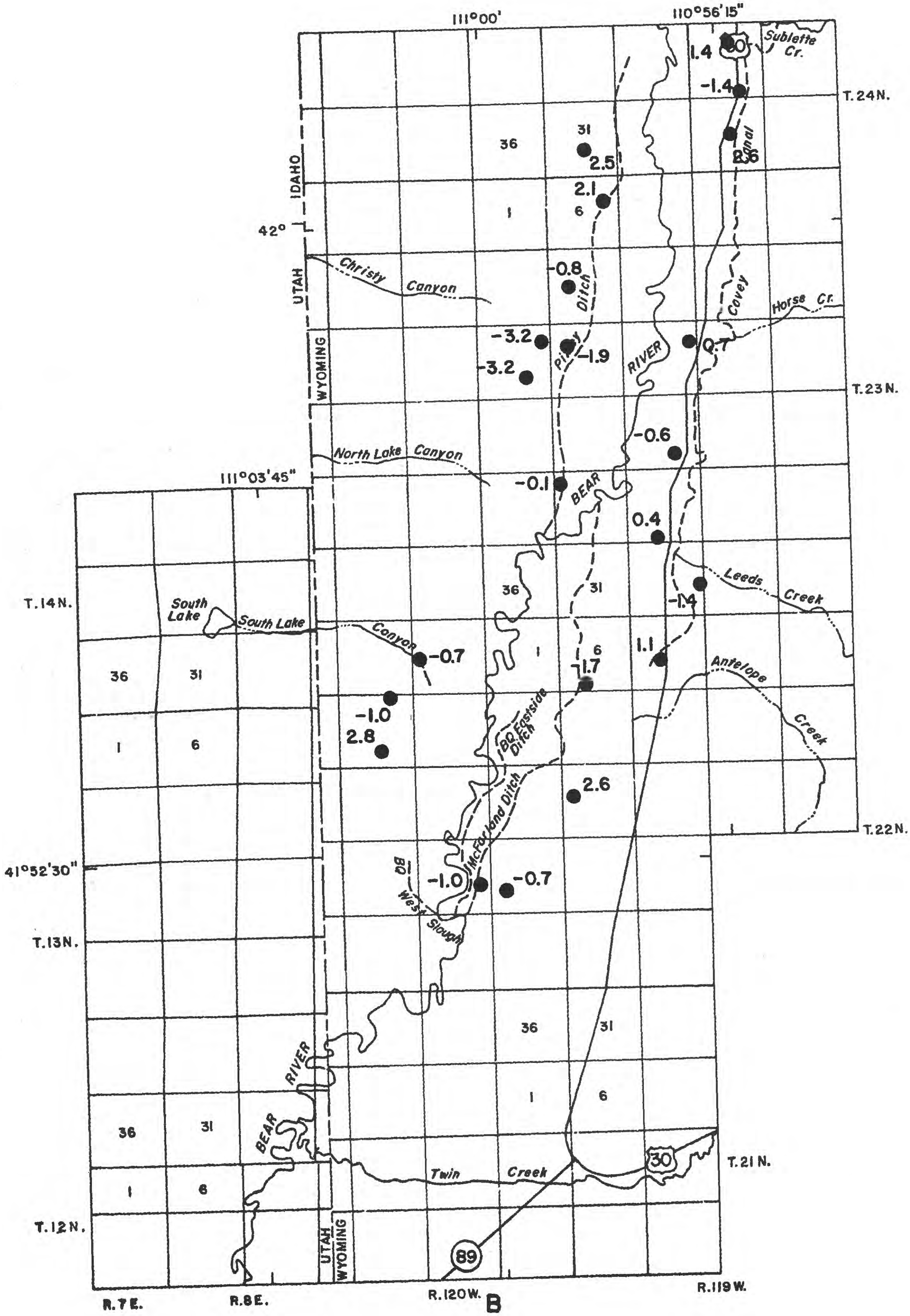
Figure 12.--Difference between calculated and measured steady-state hydraulic head in the Cokeville study area.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McFarland Ditch
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-1.0
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RIVER
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BEAR
Twin Creek
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T.12N.
UTAH
WYOMING
89
R.7E.
R.8E.
R.120W.
R.119W.
B

Table 7.--*Calculated steady-state water budget for the Cokeville study area*

Component of water budget	Ground-water recharge (+) or discharge (-) (cubic feet per second)
Stream leakage	
Bear River from upstream boundary to Pixley Ditch	-5.7
Bear River from Pixley Ditch to Sublette Creek	-12.7
Bear River from Sublette Creek to Birch Creek springs	-7.1
Bear River from Birch Creek springs to downstream boundary	-3.2
Forgen Slough	-6.1
Birch Creek springs	-2.0
Sublette Creek	+.7
Smiths Fork	+20.5
Underflow	
Upstream model boundary	+18.5
Downstream model boundary	-17.8
Tributaries	+14.9

data and rates calculated during steady-state simulation were less than the probable errors of the initial estimates. The small differences between measured and calculated stream-leakage rates indicated that stream-aquifer interaction was accurately simulated.

The ratios of standard error to calibrated value (table 6) provide relative measures of model sensitivity to changes in the parameters. In general, the steady-state simulation is very sensitive to variations in streambed-leakance coefficient and less sensitive to variation in ground-water recharge and transmissivity.

Application of the Model to Transient Flow

Simulation of flow in the Cokeville stream-aquifer system under transient conditions was used to improve estimates of specific yield and seasonal ground-water recharge and discharge. Other model parameters, including stream-channel characteristics, rates of streamflow at model boundaries, well pumpage, and maximum rates of evapotranspiration, were adequately determined from field data and were not varied during transient calibration.

The 1980 and 1981 irrigation seasons were selected for transient calibration, because 1980 represented a year of above-average streamflow, while 1981 represented a year of below-average streamflow. Successful simulation of flow for these 2 years, therefore, represents an appropriate test of the stream-aquifer model.

Initial Estimates of Model Parameters

The steady-state distribution of hydraulic head in the stream and aquifer was used as initial condition for transient simulation. A value of hydraulic head in the aquifer was determined from figure 9 for each finite-element node. Hydraulic head in the stream was determined for each stream node from topographic maps with contour intervals of 5 ft.

In addition to estimates of transmissivity, streambed-leakance coefficient, and recharge and discharge as underflow used in the steady-state simulation, the transient simulation included estimates of stream-channel characteristics, changes in rates of streamflow at model boundaries, specific yield, irrigation recharge, pumping by wells, and water loss to phreatophytes. Initial estimates of the transient parameters are given in this section of the report.

Stream-channel characteristics, including stream width and steady-state depth of water, channel slope, and Manning's coefficient of friction, were estimated from field measurement, topographic maps, and Manning's equation. Stream widths and depths of water were measured at representative locations throughout the study area. Seldom does the width of the Bear River vary significantly from 60 ft when flow is bankfull or less. The corresponding width of tributaries ranges from 10 ft at Twin Creek to 45 ft at Smiths Fork.

Channel slope was determined from topographic maps with a contour interval of 5 ft. Manning's coefficient of friction was estimated from steady-state streamflow using the Manning equation. Steady-state streamflow was routed by considering continuity of mass, including rates of streambed-aquifer leakage. The Manning equation is:

$$Q = \frac{1.486}{n} S^{1/2} wd^{5/3}$$

where Q is stream discharge, in cubic feet per second;

n is Manning's coefficient of friction, dimensionless;

S is channel slope, dimensionless;

w is stream width, in feet; and

d is depth of water, in feet.

Repeated application of the Manning equation resulted in a unique Manning's coefficient at each stream node. The average Manning's coefficient was 0.025 for flows of bankfull size or less. Because depth of water was not accurately known at all stream nodes, and because a uniform value of Manning's coefficient was considered reasonable, the depth of water was varied slightly from initial estimates to insure a uniform value of 0.025 for all stream nodes. This computed coefficient is for simulation of the stream-aquifer system; it should not be used for other purposes.

Streamflow at model boundaries was included in the transient simulation as monthly mean discharge from the Bear River, Smiths Fork, Sublette Creek, and Twin Creek. Streamflow-gaging stations are located near the model boundary on the Bear River (station 10026500) and Twin Creek (station 10027000) (U.S. Geological Survey, 1981; 1982). Streamflow in Sublette Creek was estimated from frequent measurements made during the irrigation season. Streamflow during 1980 and 1981 in Smiths Fork at Cokeville (station 10035000) was estimated from the streamflow record at Smiths Fork near Border, Wyoming (station 10032000). The downstream station was not operated during 1980 and 1981.

Monthly mean flow at Smiths Fork near Cokeville was estimated from the regression equation:

$$\begin{aligned}\log Q_c = \; & -0.9628 - 0.3990 \sin (\tfrac{\pi}{6}t) - 0.2141 \cos (\tfrac{\pi}{6}t) \\ & + 1.484 \log Q_b + 0.3175 \sin (\tfrac{\pi}{6}t) \log Q_b \\ & + 0.01845 \cos (\tfrac{\pi}{6}t) \log Q_b\end{aligned}$$

where Q_c is the monthly mean flow at Smiths Fork near Cokeville; Q_b is the monthly mean flow at Smiths Fork near Border; and t is the month. October is month 1, and September is month 12. The equation was obtained by regression analysis of monthly mean flow when both gages were in operation (1943-52). The correlation coefficient of the analysis was 0.974, and the standard error of estimate for most months was less than 10 percent of the discharge at Smiths Fork near Border.

An initial estimate of 0.20 was used for specific yield of the alluvial aquifer. Specific yield was a calibration parameter in the transient simulation. A uniform value was used throughout the Cokeville study area.

The amount of water diverted from streams for irrigation is monitored closely, but estimates of the amount of ground-water recharge in areas cannot be estimated accurately. The stream-aquifer model (Glover, 1988) simulates surface-water irrigation by distributing diverted water uniformly over an irrigated field, subtracting the consumptive-use requirement of the crop from the distributed water, and treating the remainder as recharge to the aquifer. The model does not automatically simulate soil infiltration and surface runoff, nor does it simulate consumptive use by plants. Infiltration and runoff were simulated implicitly in the Cokeville study area by reducing the rate of diversion. The amount of diversion water returning to streams by surface runoff is not known in the Cokeville study area. Consequently, an appropriate factor for reducing the rate of surface diversions was estimated during model calibration. This factor was initially estimated to be 30 percent of measured diversion rates.

Ground-water pumpage estimates given in table 3 were used in the transient simulation. Two types of irrigation wells exist in the model area: wells that are the sole source of irrigation water for a field and wells that are used to supplement surface-water supplies. The stream-aquifer model (Glover, 1988) automatically simulates pumpage from wells that supplement surface-water supplies for irrigation. The model calculates the difference between diverted surface water and consumptive-use requirements. If the diverted water is not sufficient to meet the consumptive-use requirement, the difference is supplied by ground water. In most cases, the entire area irrigated by a diversion is not irrigated by the supplemental ground water (fig. 3).

Ground-water loss to phreatophytes was simulated as a linear function of the depth to water. When the water table is at the land surface, evapotranspiration occurs at the maximum rates shown in table 8. The maximum rates were estimated by the Blaney-Criddle method as modified by Rantz (1968). Air temperature and precipitation data, which were used in the Blaney-Criddle method, are available at two sites within the study area--Border and Sage, Wyo. Evapotranspiration decreases to a value of zero as the depth to water increases to 10 ft.

Table 8.--*Calculated rate of evapotranspiration by phreatophytes when the water table is at land surface in the Cokeville study area*

Month	Evapotranspiration (feet per day)
April	1.694×10^{-3}
May	8.198×10^{-3}
June	1.353×10^{-2}
July	1.599×10^{-2}
August	1.320×10^{-2}
September	7.917×10^{-3}
October	2.984×10^{-3}

Calibration of Model

Trial-and-error techniques were used to calibrate the transient simulation because no statistical calibration algorithm was available for transient flow. During trial-and-error calibration, estimates of model parameters were varied within reasonable ranges. After each perturbation of parameter estimates, results of a simulation were compared to measured hydraulic-head and streamflow-gain-and-loss data. An attempt was made to improve estimates of model parameters, and a new simulation was undertaken in the hope that the new simulation results would more closely reproduce measured data. The process of revising parameter estimates and comparing simulation results to measured data continued until a satisfactory comparison was obtained.

Two model parameters, specific yield and irrigation recharge, were varied during calibration. Estimates of stream-channel characteristics, well pumpage, and evapotranspiration parameters were considered to be relatively reliable in comparison to specific yield and irrigation recharge. During calibration, specific yield and irrigation recharge were varied uniformly throughout the study area.

The primary calibration criterion for transient simulation was the ability to reproduce the measured streamflow at the downstream model boundary (station 10039500). Other criteria, such as the ability to reproduce measured changes in water levels in wells, were considered to be relatively insensitive measures of calibration success. While seasonal fluctuations in stream leakage, as measured by streamflow at the downstream model boundary, are large, water-level changes generally are less than 10 ft, except at pumping-well bores.

The closest correspondence of measured and simulated streamflow at the downstream model boundary (fig. 13) was obtained using a value of 0.15 for specific yield and the values of irrigation recharge given in table 9. The measured and simulated streamflow hydrographs (fig. 13) correspond quite well for the low-flow periods, when water is in short supply. The match for the high flows of May, June, and July is not as good. Errors in the simulated hydrograph during high flows probably are the result of unmeasured streamflow from tributaries. Simulated water levels throughout the model area changed less than 7 ft and returned to approximately steady-state values during the winter; this agrees well with observed data.

The calculated ground-water budget for the 1980 and 1981 simulation (table 9) shows that the major source of ground-water recharge, during most of the irrigation season, is flood-irrigated fields while the major area of ground-water discharge is the Bear River. The table also shows that the amount of ground-water pumpage is small when compared to the total ground-water discharge. The large values of irrigation recharge probably relate to the extensive practice of flood irrigation in the study area. Approximately 40 percent of measured streamflow diversions was estimated to return to streams as surface runoff.

Results of Sensitivity Analysis

Results of sensitivity analysis, as measured by the root-mean-square deviation between simulated and measured streamflow, gave a qualitative description of reliability in estimates of specific yield and irrigation recharge. The May-July period of high flow was not used to calculate the root-mean-square deviation; errors in the simulation, caused by unmeasured tributary inflow, would mask the effect of varying specific yield and irrigation recharge. The root-mean-square deviation was calculated for each combination of specific yield and irrigation recharge by:

$$rms = \left[\sum_{i=1}^{N} (Q_o - Q_c)^2 \right]^{1/2} \frac{1}{N}$$

where rms is the root-mean-square deviation, in cubic feet per second;

N is the number of months simulated;

Q_o is monthly mean discharge, in cubic feet per second, measured at streamflow-gaging station 10039500; and

Q_c is monthly mean discharge, in cubic feet per second, simulated by the model at the downstream boundary (at station 10039500).

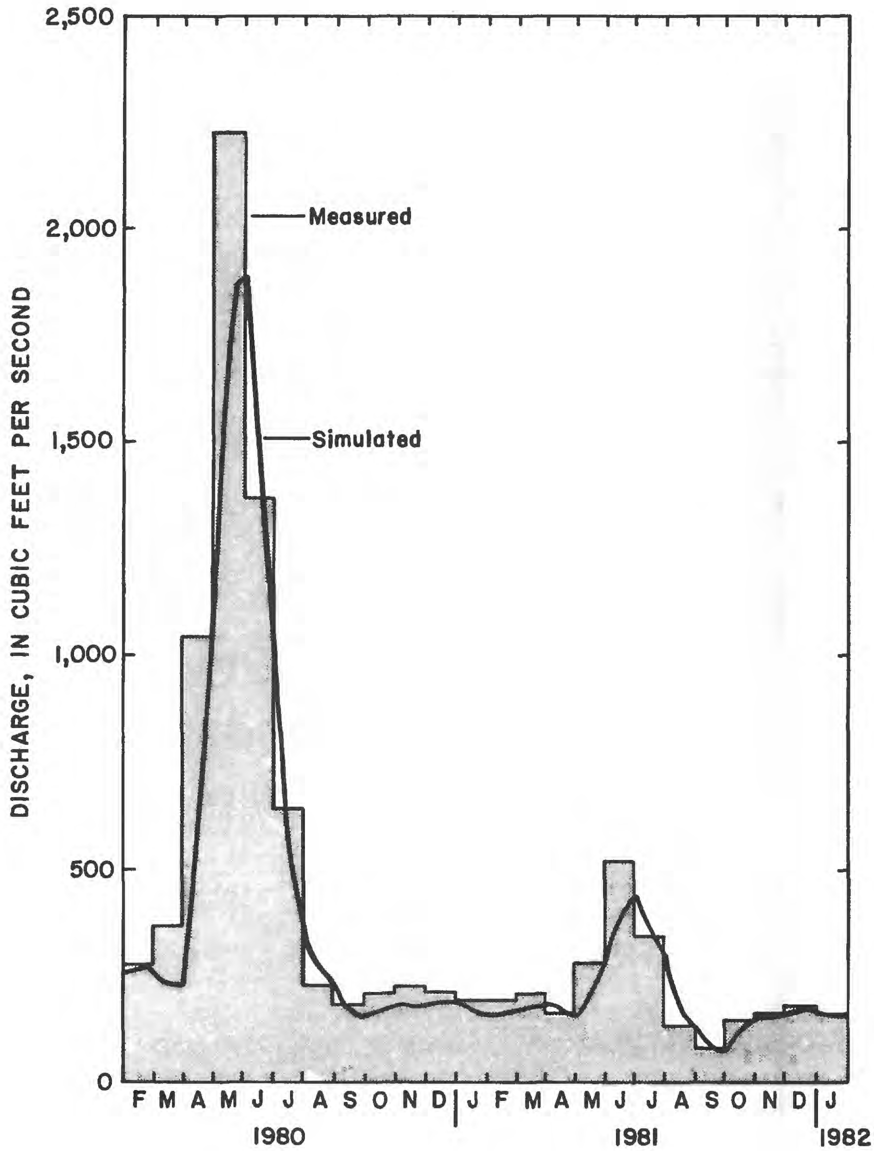

Figure 13.--Measured and simulated streamflow at the downstream boundary of the Cokeville study area.

Table 9.--*Calculated ground-water budget for the 1980 and 1981 simulation, Cokeville study area*

Date	Stream leakage	Storage	Net under-flow	Irrigation	Evapotran-spiration	Pumpage
	Ground-water recharge (+) or discharge (-), in cubic feet per second					
Feb. 1980	- 7.2	- 8.4	+15.6	0.0	0.0	0.0
Mar. 1980	- 10.5	- 5.1	+15.6	.0	.0	.0
Apr. 1980	+ 14.0	- 93.0	+15.6	+ 72.7	- 9.3	.0
May 1980	+ 21.2	-155	+15.6	+134	-13.0	-2.26
June 1980	- 92.0	-173	+15.6	+282	-27.6	-4.79
July 1980	-140	- 68.8	+15.6	+232	-32.4	-6.41
Aug. 1980	-142	- 40.3	+15.6	+106	-14.9	-5.00
Sept. 1980	-110	+ 83.4	+15.6	+ 30.4	-16.7	-2.66
Oct. 1980	- 77.5	+ 72.6	+15.6	.0	-10.7	.0
Nov. 1980	- 64.8	+ 49.2	+15.6	.0	.0	.0
Dec. 1980	- 52.0	+ 36.4	+15.6	.0	.0	.0
Jan. 1981	- 45.5	+ 29.9	+15.6	.0	.0	.0
Feb. 1981	- 38.1	+ 22.5	+15.6	.0	.0	.0
Mar. 1981	- 31.5	+ 15.9	+15.6	.0	.0	.0
Apr. 1981	- 40.3	+ 24.7	+15.6	.0	.0	.0
May 1981	- 71.6	-212	+15.6	+279	-10.4	-2.26
June 1981	-190	-148	+15.6	+364	-36.4	-4.79
July 1981	-118	- 42.0	+15.6	+178	-27.2	-6.41
Aug. 1981	-100	+ 46.9	+15.6	+ 61.0	-18.5	-5.00
Sept. 1981	- 75.5	+ 62.3	+15.6	+ 15.4	-15.1	-2.66
Oct. 1981	- 17.0	- 12.1	+15.6	.0	-10.7	.0
Nov. 1981	- 11.4	- 4.2	+15.6	.0	.0	.0
Dec. 1981	- 10.7	- 4.9	+15.6	.0	.0	.0
Jan. 1982	- 11.5	- 4.1	+15.6	.0	.0	.0

The root-mean-square deviations (table 10) show that errors in estimates of irrigation recharge had a greater effect on model results than any errors that occurred in the estimate of specific yield. Nevertheless, errors in either parameter did not significantly affect the simulated streamflow.

Table 10.--*Results of sensitivity analysis of Cokeville flow model for the 1980 and 1981 simulation*

Irrigation recharge (percentage of diverted flow)	Root-mean-square deviation of streamflow for indicated specific yield (cubic feet per second)			
	Specific yield (dimensionless)			
	0.05	0.10	0.15	0.20
0	13.1	14.3	15.3	28.6
25	12.8	11.4	12.3	19.5
40	12.5	10.5	9.3	10.6
55	10.9	10.5	10.2	10.0

EFFECTS OF GROUND-WATER WITHDRAWALS IN THE COKEVILLE STUDY AREA

During most years, the rate of ground-water pumping along the Bear River is less than the accuracy of streamflow measurements. Existing pumping rate is less than 8 ft^3/s, whereas the minimum discharge of the Bear River at Border, Wyo. (station 10039500), averaged for all years of record, is 91 ft^3/s of streamflow (Lines and Glass, 1975). Low-flow measurements along the Bear River generally are rated by hydrographers as either fair or poor, and measurement error probably is greater than 10 percent throughout most low-flow periods.

Therefore, during years of average or above-average streamflow, current ground-water withdrawals do not measurably reduce streamflow. For effects of pumping to be measurable, streamflow of the Bear River must be significantly less than 80 ft^3/s. If new wells are drilled and pumpage increases greatly, streamflow may be measurably reduced during years of average streamflow.

During years of less-than-average streamflow, the amount of water diverted in the Cokeville study area is reduced to meet commitments of the interstate compact. Reductions in the amount of water diverted were common during dry years prior to extensive well development. For example, streamflow was very small during 1977, and the surface-water diversions--B.Q. Eastside Ditch, B.Q. West Slough, and Pixley Ditch--were not operated during some months in order to meet compact commitments. Most ground-water development has occurred in the Cokeville study area since 1977.

The Cokeville stream-aquifer model was used to evaluate possible effects of existing ground-water development during a year of less-than-average streamflow. Because streamflow for 1977 was the lowest recorded, streamflow at model boundaries and diversion rates for that year were used in the analysis (Bear River Commission, 1978). Two simulations were made. The first was made assuming no ground-water pumping; the second simulation was made using pumping rates given in table 11. The pumping rates in this table were obtained by calculating the consumptive-use requirements of irrigated lands based on long-term average air temperature. Supplemental ground water was needed to meet consumptive-use requirements of land irrigated by the Covey Canal. Other model-input data, such as evapotranspiration, were obtained using procedures described in previous sections of this report.

The effects of ground-water withdrawals on the flow of the Bear River at Border, Wyo., were identified by comparing the streamflow calculated in the simulation with well pumpage to the streamflow calculated in the simulation without pumpage (fig. 14). The largest effect on streamflow occurred in August, which correlates with the period of maximum pumping, July and August. August streamflow of the Bear River was reduced by 3.4 ft^3/s. The effects of pumping continued throughout the winter, but the reduction in streamflow was less than 0.5 ft^3/s by the following April.

With one exception, pumping did not cause streams to become dry or reduce the amount of water available for diversion. The one exception occurred during August. In the simulation without pumping, the Cook Canal diverted water from the Bear River at a rate of 31.8 ft^3/s. As a result of reduced streamflow during the simulation with pumping, the total flow of the Bear River at the Cook Canal was 28.2 ft^3/s. The Cook Canal diverted the entire flow of the Bear River, causing the stream to become dry for a short distance. Ground-water discharge to the Bear River prevented the stream from being dry at the Idaho-Wyoming border.

Because part of the water pumped by wells in the simulation was derived from water that otherwise would have been transpired by phreatophytes, the total volume pumped was greater than the volume of streamflow reduction. Approximately 84 percent of the water pumped by wells resulted in a decrease in streamflow. The remaining 16 percent resulted in a decrease in water lost to phreatophytes.

EFFECTS OF GROUND-WATER WITHDRAWALS IN THE EVANSTON STUDY AREA

Hydrogeologic characteristics of aquifers in the Evanston study area could not be estimated with the same accuracy as in the Cokeville study area. Transmissivity estimates are based on very few field data, while no field data are available to estimate specific yield. There were insufficient field data to map potentiometric surfaces accurately or to estimate transmissivity and specific yield. Owing to the limited amount of data, the hydrologic significance of aquifer boundaries and changes in the distribution of ground-water recharge and discharge are not well described.

Because hydrologic knowledge of the Evanston study area is limited, the methodology used to evaluate effects of ground-water withdrawals is correspondingly simplistic. Complex computer models, which require a knowledge of the distribution of aquifer properties and ground-water recharge and discharge, are not appropriate for application in the Evanston study area. Analytical techniques that assume relatively simple boundary conditions and uniform aquifer properties are appropriate. The streamflow-depletion method, described next, is such a technique.

Table 11.--*Ground-water pumping rates used in the Cokeville predictive analysis of a 1-year drought*

Well number	Average pumping rate (cubic feet per second)				
	May	June	July	August	September
23-119-32bd	0.42	0.89	1.19	0.93	0.49
23-120-13db	.32	.67	.90	.70	.37
23-120-25ca	.56	1.18	1.58	1.23	.66
24-119-05cc	.08	.17	.23	.18	.09
24-119-31db	.37	.79	1.05	.82	.44
25-119-17cb	.39	.82	1.10	.86	.46
25-119-17db	.13	.27	.37	.28	.15
Covey Canal (seven wells with supplemental water rights)	.05	.21	.93	.0	.14
Total	2.32	5.00	7.35	5.00	2.80

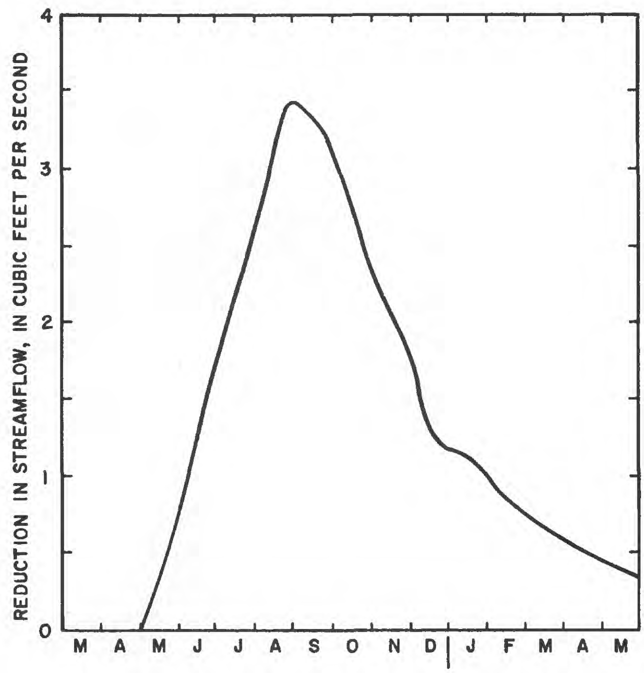

Figure 14.--Reduction in streamflow due to depletion by ground-water withdrawals during a 1-year drought.

Streamflow-Depletion Method

An analytical method, called the streamflow-depletion method (Jenkins, 1968a; 1968b) was used in the Evanston study area to evaluate effects of pumping on streamflow. This method calculates rate and volume of pumped water that is obtained from streamflow as a function of pumping time and a streamflow-depletion factor. The streamflow-depletion factor is related to transmissivity, specific yield, and distance between the stream and pumping well. The larger the streamflow-depletion factor, the longer it takes for pumping to affect streamflow. Effects of hydrologic boundaries are included in the streamflow depletion by using image-well theory. For no-flow boundaries, the result is a decrease in streamflow-depletion factor from that calculated when no boundary is present. The streamflow-depletion factor at the boundary is decreased by half. The influence of a semipermeable streambed is included by increasing the apparent distance between the stream and well to account for the difference between hydraulic conductivity of the streambed and alluvial aquifer (Hantush, 1965). A similar procedure is used to account for the difference between transmissivity of the alluvial and Wasatch aquifers.

Although reliable estimates of the streamflow-depletion factor used in the analytical method cannot be made, streamflow-depletion factors corresponding to overestimates and underestimates of aquifer properties have been mapped in the Evanston study area (figs. 15 and 16). The overestimates and underestimates of streamflow-depletion factors, when used in the analytical method of Jenkins (1968a), give reasonable ranges for the volume and timing of ground-water pumpage derived from streamflow.

Values of transmissivity and storage coefficient used in figure 15 to overestimate the streamflow-depletion factor were 100,000 ft/d and 0.2 for the alluvial aquifer, and 1,000 ft/d and 0.1 for the Wasatch aquifer. Values of transmissivity and storage coefficient used in figure 16 to underestimate the streamflow-depletion factor were 200,000 ft/d and 0.1 for the alluvial aquifer, and 3,000 ft/d and 0.1 for the Wasatch aquifer.

The streamflow-depletion factors for the alluvial aquifer (figs. 15 and 16) generally are less than 20 days, implying that wells pumping from the alluvial aquifer will affect streamflow quite rapidly. Specifically, after as little as 20 days of pumping, streamflow in the Bear River will be reduced by at least 28 percent of the pumped volume. Similarly, the effect of pumping on streamflow, after actual pumping has stopped, will dissipate rapidly.

The streamflow-depletion factors for the Wasatch aquifer (figs. 15 and 16) are much larger than for the alluvial aquifer--implying that wells pumping from the Wasatch aquifer will affect streamflow only after several months. For example, if an overestimate of streamflow-depletion factor is used (fig. 15) at NW1/4 NW1/4 NW1/4 sec. 11, T. 14 N., R. 120 W., the predicted streamflow would not be significantly affected until more than 2 years after pumping had occurred. In contrast, if the underestimate of streamflow-depletion factor is used (fig. 16), predicted streamflow would be measurably reduced after 6 months. This difference could be important in the management of water wells with temporary-use permits. Temporary-use permits are typically granted by the Wyoming State Engineer for water wells used during oil and gas drilling.

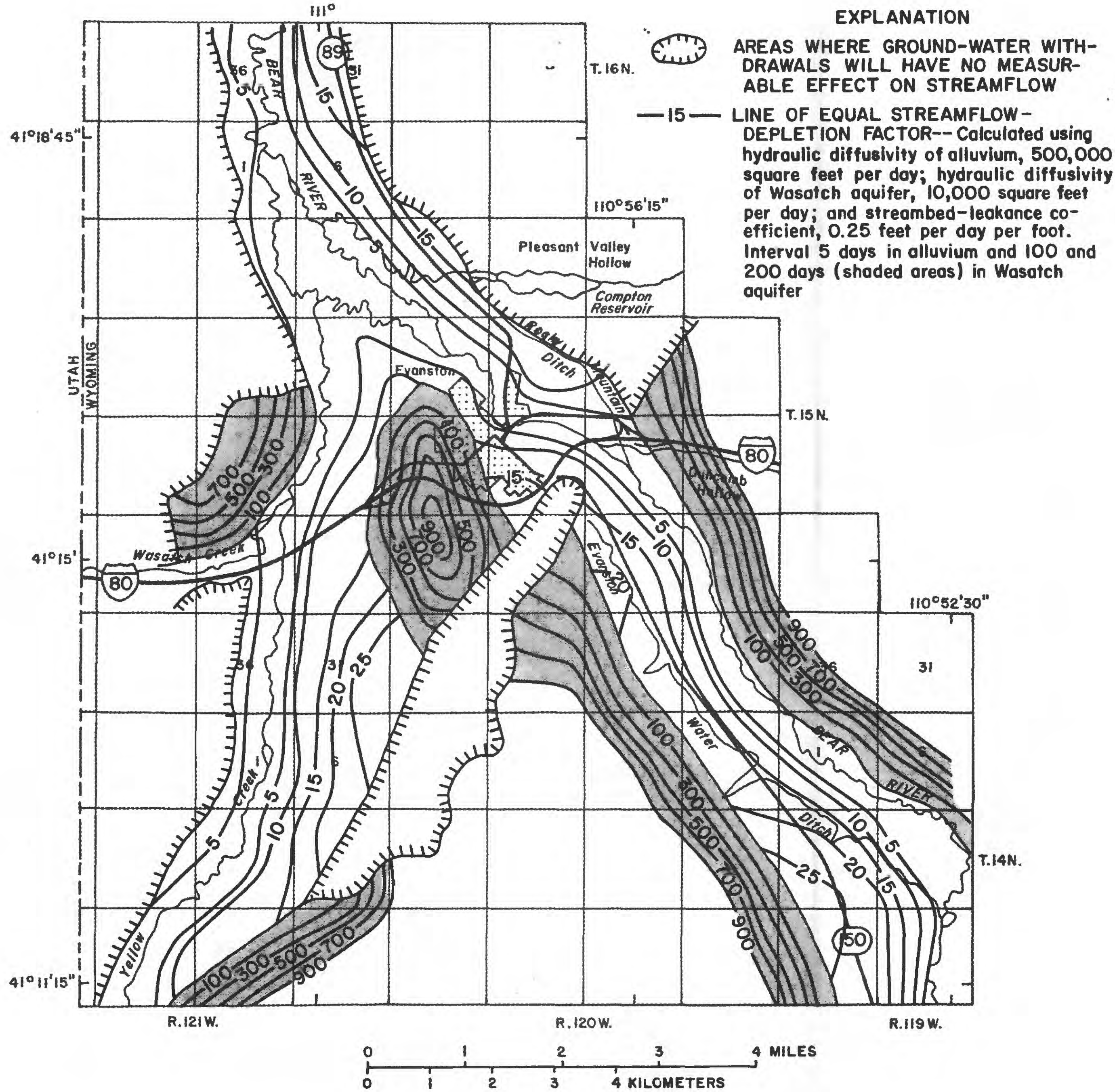

Figure 15.--Overestimates of streamflow-depletion factor for aquifers in the Evanston study area.

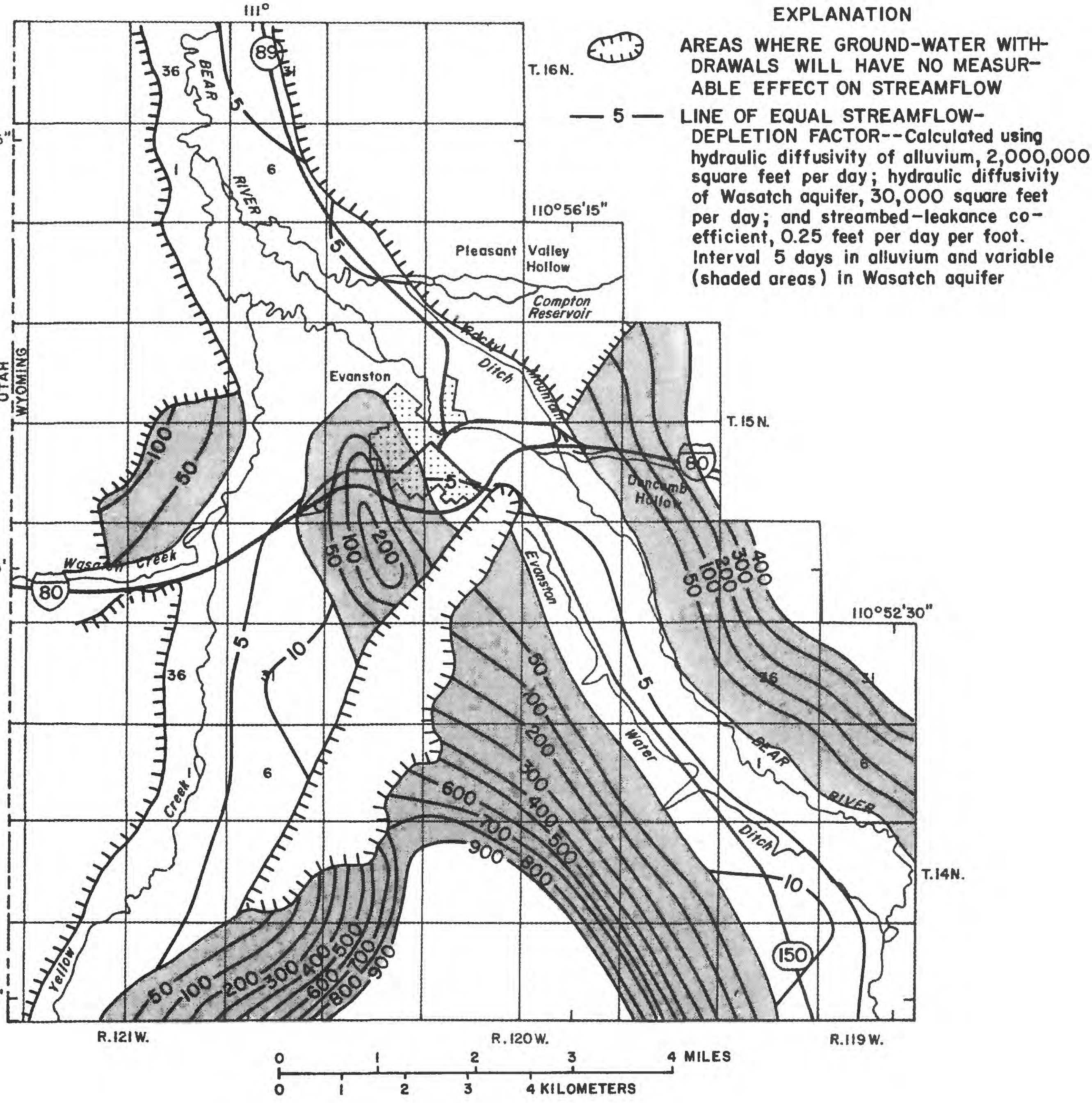

Figure 16.--Underestimates of streamflow-depletion factor for aquifers in the Evanston study area.

Well pumpage in several areas that are separated from the alluvial aquifer by zones of nearly impermeable rock (fig. 15) will not measurably reduce streamflow. Wells in an area immediately south of the Medicine Butte Fault (the large, hatchured area south of Evanston, figs. 15 and 16) are completed in the Hams Fork Conglomerate Member of the Evanston Formation or older strata. The nearly impermeable upper members of the Evanston Formation isolate the Bear River from effects of pumping. Well development in extensive areas northeast and west of Evanston also will have no measurable effect on streamflow. These wells are completed in the Wasatch aquifer, but the presence of the Fowkes Formation or Evanston Formation prevents movement of measurable amounts of water between the alluvium and Wasatch Formation.

Well pumping from perched zones within the Wasatch aquifer will not measurably affect streamflow. Distinguishing between wells that are open to perched zones and wells that are open to the main saturated zone of the Wasatch is possible by reviewing well-completion data. If the altitude of the open interval of a well is significantly higher than the altitude of nearby streams, the well probably obtains water from a perched zone. If the altitude of the open interval is similar or less than the altitude of nearby streams, the well probably is in direct hydrologic connection with the streams. Perched zones have not been observed in the Wasatch aquifer, north of the Medicine Butte Fault, between the Bear River and Yellow Creek.

Pumping by Evanston Municipal Wells

The streamflow-depletion method was used to evaluate effects of pumping the Evanston municipal wells. Wells were assumed to be pumped from May through September at rates similar to those recorded from 1981-83 (table 12). Estimates of streamflow-depletion factor were obtained for each well by interpolating between contour lines in figures 15 and 16. The streamflow-depletion factor is proportional to the squared distance between stream and well. Therefore, linear interpolation between values of the square root of streamflow-depletion factor was used. The effect of a variable pumping rate was evaluated by applying image-well theory. Jenkins (1968a) gives examples of the use of image-well theory when evaluating the effects of pumping on streamflow. The total reduction in streamflow resulting from pumping was obtained by adding the individual effects of each well.

The reduction in streamflow due to pumping the Evanston municipal wells is greatest during the pumping season, but streamflow continues to be affected for an extended period after pumping stops (fig. 17). Most of these reductions are the result of pumping from wells completed in the alluvial aquifer, but most of the reduction in streamflow after pumping stops can be traced to wells completed in the Wasatch aquifer. The effect of pumping on streamflow is likely to be less than 0.25 ft^3/s 100 days after pumping stops.

A qualitative measure of the error in the analysis is provided by the difference between curves (fig. 17) using overestimates and underestimates of streamflow-depletion factor. The error in the analysis generally is less than 20 percent of the reduction in streamflow that was calculated using an overestimate of streamflow-depletion factor. If additional wells, open to the Wasatch Formation, were included in the analysis, the error would be greater.

Table 12.--*Average monthly ground-water pumping rates used in the analysis of Evanston municipal wells*

Well number	Aquifer	Monthly mean pumping rate (cubic feet per second)				
		May	June	July	August	September
15-120-17ac	Alluvial	0.00	0.26	0.48	0.06	0.00
15-120-20ab	Wasatch	.02	.48	.65	.37	.01
15-120-20bd	Wasatch	.01	.24	.29	.26	.06
15-120-21bb	Alluvial	.06	.69	.87	.61	.02
15-120-21bd	Alluvial	.01	.31	.48	.21	.00
15-120-21cb	Alluvial	.01	.05	.02	.00	.00

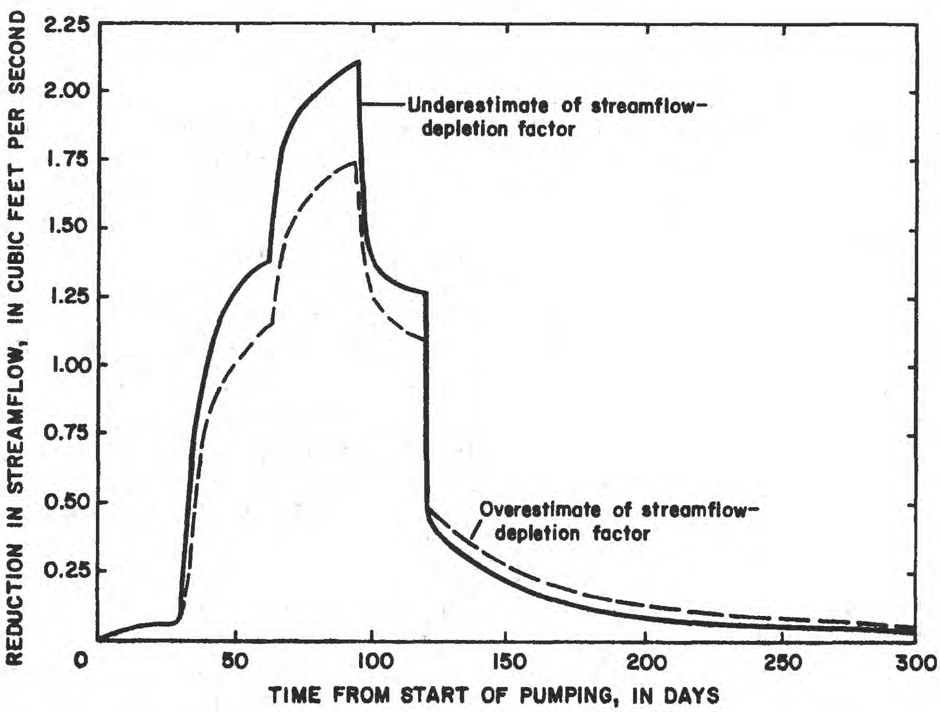

Figure 17.--Reduction in streamflow due to depletion by ground-water withdrawals from Evanston municipal wells.

Estimates of the effects of pumping on streamflow in the Evanston area can be decreased by 16 percent to account for ground-water withdrawals that otherwise would have been evapotranspired by phreatophytes. The streamflow-depletion method does not include evapotranspiration. Phreatophytes are common in both the Cokeville and Evanston study areas, and model analysis in the Cokeville area showed that approximately 16 percent of pumped water was obtained from water that would normally be evapotranspired by phreatophytes. Therefore, the effects of pumping on streamflow given in figure 17 are overestimated. Reducing the estimates by 16 percent to account for evapotranspiration is reasonable.

SUMMARY

The stream-aquifer hydrology near two major centers of ground-water pumping in the upper Bear River valley of Utah and Wyoming was described. Water managers need the information for quantifying the relation of ground-water movement to streamflow. The investigation, a cooperative effort by the U.S. Geological Survey and the Wyoming State Engineer, focused on pumping centers near Cokeville, Wyo. (where well water is obtained from alluvium) and Evanston, Wyo. (where wells obtain water from alluvium and bedrock formations).

Streams in the Cokeville study area include the Bear River, Smiths Fork, and several small tributaries; the principal aquifer is in alluvial deposits adjacent to the streams. The thickness of the alluvial aquifer near Cokeville has not been determined, but it is at least 450 ft. Aquifers in bedrock are hydraulically isolated from the alluvium, and are not part of the stream-aquifer system.

A steady-state ground-water budget of the Cokeville study area was constructed from field data and was successfully simulated using a finite-element model. The simulated budget showed the Bear River to gain 36.8 ft^3/s from the alluvial aquifer; tributaries lost 21.2 ft^3/s by stream leakage. Underflow across the upstream boundary accounted for 18.5 ft^3/s; underflow across the downstream boundary accounted for 17.8 ft^3/s. An additional 14.9 ft^3/s occurred as underflow through alluvium along small tributaries of the Bear River.

Estimates of aquifer properties and the distribution of ground-water recharge were obtained by steady-state and transient simulations of the Cokeville stream-aquifer system. The 1980 and 1981 irrigation seasons were selected for calibration of the transient simulation. Transmissivity estimates ranged over several orders of magnitude, but values greater than 100,000 ft^2/d were obtained for much of the study area. Specific yield was estimated to be 0.15. Steady-state ground-water recharge was in the form of stream leakage and underflow, while large amounts of seasonal recharge occurred in areas irrigated by surface water. The calculated ground-water budget for the 1980 and 1981 simulation shows that the main source of recharge during most of the irrigation season is flood-irrigated fields, while the main area of discharge is the Bear River. Also, ground-water pumpage is small, in comparison with total ground-water discharge.

Because current (1982) rates of ground-water withdrawal in the upper Bear River valley are small, the effects of pumping on streamflow, particularly during years of average or greater-than-average streamflow, are less than the accuracy of discharge measurements. For this reason, no detailed analysis of the effects of pumping during years of average or greater-than-average streamflow is included in this report. If pumping in the future increases dramatically, such detailed analysis may be justified.

Computer simulation of the Cokeville stream-aquifer system demonstrated the effects of ground-water pumping during a year of less-than-average streamflow. Simulation using 1977 data showed that streamflow would be reduced by a maximum of 3.4 ft^3/s during August, which correlates with the period of maximum pumping, July and August. By the start of the following irrigation season, the effect of pumping during the previous year would be less than 0.5 ft^3/s. Approximately 84 percent of the water pumped by wells was derived from water that otherwise would have been discharged to the river, and 16 percent from water that otherwise would have been lost to phreatophytes.

Streams in the Evanston study area include the Bear River and Yellow Creek; aquifers are in the alluvial deposits and the Wasatch Formation. Aquifers in older strata are not part of the stream-aquifer system. Relatively impermeable rocks located on upthrown sides of normal faults have isolated the Wasatch Formation from the stream-aquifer system in some locations; in general, however, the alluvial and Wasatch aquifers are in direct hydraulic connection. Perched zones within the Wasatch aquifer are not considered part of the stream-aquifer system.

Seepage-run data were used to identify stream-aquifer relations in the Evanston study area, but a complete ground-water budget could not be constructed with existing data. Seepage-run data indicate the Bear River gains water from the alluvial aquifer, and Yellow Creek loses water. Because hydraulic-head data are insufficient, reliable computer models of the Evanston stream-aquifer system that, in turn, could be used to estimate rates of underflow and irrigation recharge, could not be developed.

The transmissivity of aquifers in the Evanston area could not be determined accurately from available data, but probably varies over several orders of magnitude. Limited field data indicate transmissivity of the alluvial aquifer ranges from 14,000 to 72,000 ft^2/d. The alluvium is similar in character to alluvium near Cokeville; thus values of transmissivity should be similar. In contrast, the transmissivity of the Wasatch aquifer probably is at least one order of magnitude less than that of the alluvial aquifer.

An analytical streamflow-depletion method, applied to pumpage from municipal wells of Evanston, showed that the largest reduction in streamflow occurred during the pumping season, and streamflow is affected for an extended period after pumpage stops. Most of the reduction was the result of pumping from wells completed in the alluvial aquifer. Effects of pumping the Wasatch aquifer occurred over a longer time. The total effect of pumping, 100 days after pumping stopped, is likely to be less than 0.25 ft^3/s.

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