Hydrogeology and Simulation of Flow between the Alluvial and Bedrock Aquifers in the Upper Black Squirrel Creek Basin, El Paso County, Colorado

by Kenneth R. Watts

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CONVERSION FACTORS, ABBREVIATIONS, AND VERTICAL DATUM

Multiply	By	To obtain
acre	0.4047	hectare
acre-foot (acre-ft)	1,233	cubic meter
acre-foot per year (acre-ft/yr)	1,233	cubic meter per annum
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter (m)
foot per day (ft/d)	0.3048	meter per day
foot per mile (ft/mi)	0.1894	meter per kilometer
foot per second (ft/s)	30.48	centimeter per second (cm/s)
foot per year (ft/yr)	0.3048	meter per annum
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	2.54	centimeter (cm)
mile (mi)	1.609	kilometer
square foot (ft^2)	0.09290	square meter
square foot per day (ft^2/d)	0.09290	square meter per day
square inch per pound (in^2/lb)	0.01422	square centimeters per gram
square mile (mi^2)	2.590	square kilometer

Degree Celsius (°C) may be converted to degree Fahrenheit (°F) by using the following equation:

$$°F = 9/5\ (°C) + 32.$$

Degree Fahrenheit (°F) may be converted to degree Celsius (°C) by using the following equation:

$$°C = 5/9\ (°F - 32).$$

The following terms and abbreviations also are used in this report:

calorie per centimeter per second per degree Celsius (cal/cm/s/°C)

calorie per gram per degree Celsius (cal/gm/ °C)

gram per cubic centimeter (g/cm^3)

microsiemens per centimeter at 25 degrees Celsius (µS/cm)

milligram per liter (mg/L)

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

GLOSSARY

Aquifer.—"***a formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells or springs" (Lohman and others, 1972, p. 2).

Cell.—A block in the three-dimensional mathematical mesh used to subdivide the aquifer system (McDonald and Harbaugh, 1988, p. 2–2).

Conductance.—The product of hydraulic conductivity and cross-sectional area of flow divided by the length of the flow path (McDonald and Harbaugh, 1988, p. 2–11).

Confining unit.—"***a body of "impermeable" material stratigraphically adjacent to one or more aquifers. In nature, however, its hydraulic conductivity may range from nearly zero to some value distinctly lower than that of the aquifer" (Lohman and others, 1972, p. 5–6). If a confining unit is permeable, it is referred to as a leaky confining unit. Although a confining unit may have very small permeability, it may store substantial volumes of water, which may flow to adjacent aquifers under a sufficiently large hydraulic gradient.

Constant-head boundary.—A boundary in the model represented by a cell in which water levels are assumed to remain constant.

Drain.—A sink or head-dependent flow boundary in the model hydrologic feature which can receive discharge from an aquifer but which can not recharge the aquifer. Drains are used to represent springs and seepage faces.

Evapotranspiration.—The combined loss of water from a given area by evaporation from the land and transpiration by plants.

General-head boundary.—A boundary in the model, similar to a constant head in that the water level at the node is assumed to remain constant, except that it is located at some distance from the actual boundary of the model.

Gradient of head (∇).—See "hydraulic gradient."

Head-dependent flow.—Flow across a cell boundary that is a function of the difference between heads (water levels) in adjacent cells and the conductance between the centers of the adjacent cells.

Homogeneity and *heterogeneity.*—If a property, for example, hydraulic conductivity ($K_{x,y,z}$) is independent of position within a hydrogeologic unit, then the unit is said to be homogeneous; if the property varies with position, then the unit is heterogeneous. Freeze and Cherry (1979, p. 30–32) describe three major types of heterogeneity: (1) Layered heterogeneity, (2) discontinuous heterogeneity, and (3) trending heterogeneity. Layered heterogeneity is common in sedimentary sequences of alternating fine- and coarse-grained rocks. Discontinuous heterogeneity occurs across faults, large-scale stratigraphic features, and contacts between unconsolidated deposits and rock. Trending heterogeneity occurs when a property is more or less a regular function of lateral position within a hydrogeologic unit. Because most geologic formations exhibit some spatial variation in properties, there probably are no truly homogeneous hydrogeologic units. An alternative definition of a homogeneous hydrogeologic unit is one in which the mean value of the property is constant in space (Freeze and Cherry, 1979, p. 31).

Hydraulic conductivity ($K_{x,y,z}$).—"***the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow" through the porous medium (Lohman and others, 1972, p. 4). Hydraulic conductivity primarily is a function of the size and distribution of pore space.

Hydraulic diffusivity (T/S or $K_{x,y}/S_y$).—The hydraulic diffusivity of a porous media is the ratio of transmissivity (T) to storage coefficient (S) for confined conditions and the ratio of hydraulic conductivity ($K_{x,y}$) to specific yield (S_y) for unconfined conditions (Lohman and others, 1972, p. 8).

Hydraulic gradient (dh/dl).—"***the change in static head per unit of distance in a given direction. If not specified, the direction generally is understood to be that of the maximum rate of decrease in head. The *gradient of head* is a mathematical term which refers to the vector denoted by ∇h or grad h, whose magnitude dh/dl is equal to the maximum rate of change in head and whose direction is that in which the maximum rate of increase occurs. The hydraulic gradient and the gradient of head are equal but of opposite sign" (Lohman and others, 1972, p. 8–9).

Infiltration.—The downward flow of water into the soil or rock.

Isotropy and *anisotropy.*—If all significant properties, for example, hydraulic conductivity, are independent of direction of measurement, then the system is isotropic (Lohman and others, 1972, p. 9); a hydrogeologic unit is anisotropic if significant properties vary with direction of measurement. Primary causes of anisotropy in sedimentary rocks and unconsolidated sediments are stratification and the orientation of clay minerals.

Porosity (η).—"The porosity of a rock or soil is its property of containing interstices or voids and may be expressed quantitatively as the ratio of the volume of its interstices to its total volume," and "*effective porosity* refers to the amount of interconnected pore space available for fluid transmission" (Lohman and others, 1972, p. 10).

Specific discharge (q_z).—"***the rate of discharge of ground water per unit area measured at right angles to the direction of flow" (Lohman and others, 1972, p. 11–12). Specific discharge is sometimes referred to as Darcian velocity.

Specific storage (S_s).—"***the volume of water released from or taken into storage per unit volume of the porous medium per unit change in head" (Lohman and others, 1972, p. 13). Specific storage is a function of the porosity (η), the specific weight of water (γ_w), the bulk modulus of elasticity of water (E_w), and the constrained modulus of elasticity of the porous medium (E_k). Specific storage, as defined by Lohman (1979, p. 9), is given as:

$$S_s = \eta\gamma_w \left(\frac{1}{E_w} + \frac{C}{\eta E_k}\right) \tag{1}$$

where C equals unity (1) for uncemented granular material or C approximately equals η in incompressible porous media, such as limestone in which tubular solution channels are present. C lies between these limits for sandstones.

Specific yield (S_y).—"***the ratio of (1) the volume of water which the rock or soil, after being saturated, will yield by gravity to (2) the volume of rock of (sic) soil***. It is equal to porosity minus specific retention" (Lohman and others, 1972, p. 12).

Storage coefficient (S).—"***the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. In a confined water body, the water derived from storage with decline in head comes from expansion of the water and compression of the aquifer***." (Lohman and others, 1972, p. 13). The storage coefficient is the integration of the saturated thickness (b) and specific storage (S_s) of a confined water body, $S = S_s b$. In unconfined aquifers, the storage coefficient is virtually equal to the specific yield.

Transmissivity (T).—"Transmissivity is the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient***." It is "equal to an integration of the hydraulic conductivities across the saturated part of the aquifer perpendicular to the flow paths" (Lohman and others, 1972, p. 13).

Vertical leakance (K_z/b).—The ratio of vertical hydraulic conductivity (K_z) with the thickness (b) of the hydrostratigraphic unit (Lohman, 1979, p. 30). Units of leakance are per time.

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Abstract

Anticipated increases in pumping from the bedrock aquifers in El Paso County potentially could affect the direction and rate of flow between the alluvial and bedrock aquifers and lower water levels in the overlying alluvial aquifer. The alluvial aquifer underlies about 90 square miles in the upper Black Squirrel Creek Basin of eastern El Paso County. The alluvial aquifer consists of unconsolidated alluvial deposits that unconformably overlie siltstones, sandstones, and conglomerate (bedrock aquifers) and claystone, shale, and coal (bedrock confining units) of the Denver Basin. The bedrock aquifers (Dawson, Denver, Arapahoe, and Laramie-Fox Hills aquifers) are separated by confining units (upper and lower Denver and the Laramie confining units) and overlie a relatively thick and impermeable Pierre confining unit. The Pierre confining unit is assumed to be a no-flow boundary at the base of the alluvial/bedrock aquifer system.

During 1949–90, substantial water-level declines, as large as 50 feet, in the alluvial aquifer resulted from withdrawals from the alluvial aquifer for irrigation and municipal supplies. Average recharge to the alluvial aquifer from infiltration of precipitation and surface water was an estimated 11.97 cubic feet per second and from the underlying bedrock aquifers was an estimated 0.87 cubic foot per second.

Water-level data from eight bedrock observation wells and eight nearby alluvial wells indicate that, locally, the alluvial and bedrock aquifers probably are hydraulically connected and that the alluvial aquifer in the upper Black Squirrel Creek Basin receives recharge from the Denver and Arapahoe aquifers but locally recharges the Laramie-Fox Hills aquifer.

Subsurface-temperature profiles were evaluated as a means of estimating specific discharge across the bedrock surface (the base of the alluvial aquifer). However, assumptions of the analytical method were not met by field conditions and, thus, analyses of subsurface-temperature profiles did not reliably estimate specific discharge across the bedrock surface. The vertical hydraulic diffusivity of a siltstone and sandstone in the lower Denver confining unit was estimated, by an aquifer test, to be about 8×10^{-4} square foot per day.

Physical and chemical characteristics of water from the bedrock aquifers in the study area generally differ from the physical and chemical characteristics of water from the alluvial aquifer, except for the physical and chemical characteristics of water from one bedrock well, which is completed in the Laramie-Fox Hills aquifer. In the southern part of the study area, physical and chemical characteristics of ground water indicate downward flow of water from the alluvial aquifer to the Laramie-Fox Hills aquifer.

A three-dimensional numerical model was used to evaluate flow of water between the alluvial aquifer and underlying bedrock. Simulation of steady-state conditions indicates that flow from the bedrock aquifers to the alluvial aquifer was about 7 percent of recharge to the alluvial aquifer, about 0.87 cubic foot per second. The potential effects of withdrawal from the alluvial and bedrock aquifers at estimated (October 1989 to September 1990) rates and from the bedrock aquifers at two larger hypothetical rates were simulated for a 50-year projection period. The model simulations indicate that water levels in the alluvial aquifer will decline an average of 8.6 feet after 50 years of pumping at estimated October 1989 to September 1990 rates. Increases in withdrawals from the bedrock aquifers in El Paso County were

simulated to: (1) Capture flow that currently discharges from the bedrock aquifers to springs and streams in upland areas and to the alluvial aquifer, (2) induce flow downward from the alluvial aquifer, and (3) accelerate the rate of water-level decline in the alluvial aquifer.

INTRODUCTION

Recent and anticipated population growth in El Paso County, Colorado, has caused concern among local water users and water-resource managers regarding the potential effects of anticipated increases in pumping from the Denver Basin bedrock *aquifers* on the alluvial aquifer in the upper Black Squirrel Creek Basin. The U.S. Geological Survey, in cooperation with the Cherokee Metropolitan District; the Colorado Springs Utilities, Water Resources Department; and the Upper Black Squirrel Creek Ground Water Management District, began a study in 1987 to quantitatively evaluate ground-water flow between the alluvial and bedrock aquifers in the upper Black Squirrel Creek Basin.

Results from a previous study of the alluvial aquifer (Buckles and Watts, 1988) and of the potential effects of pumpage from the bedrock aquifers of northern El Paso County (Banta, 1989) were used extensively in this study. The emphasis of this study is on the hydrogeology of the upper Black Squirrel Creek Basin and provides a current (1991) analysis, using a model capable of simulating flow between the bedrock aquifers and the alluvial aquifer, in the upper Black Squirrel Creek Basin.

Purpose and Scope

This report describes the results of a study to refine knowledge of the hydrogeology of the alluvial aquifer and underlying bedrock aquifers and *confining units* in the upper Black Squirrel Creek Basin and to quantitatively evaluate the potential hydrologic effects to the alluvial aquifer of anticipated increases in withdrawals from the bedrock aquifers in El Paso County. The report presents analyses of hydrogeologic data collected during April 1987 through September 1990 and results from a numerical model of flow in the alluvial/bedrock aquifer system.

Hydrogeologic data presented in this report include: (1) 1987–90 water-level data for selected wells that are completed in the alluvial aquifer and in underlying bedrock aquifers and confining units, (2) subsurface-temperature profiles that were measured in the eight bedrock wells during November 1987, (3) aquifer-test data measured during September 1989, and (4) water-quality data for samples collected from the eight bedrock observation wells during June and July 1987. Selected hydrogeologic data for the alluvial aquifer that were collected during previous studies or as part of ongoing data-collection programs also are presented in this report.

Analyses of the hydrogeologic data were used to refine understanding of the hydraulic connection between the alluvial aquifer and underlying bedrock aquifers and confining units. These analyses, which are described in this report, include: (1) Comparisons of 1987–90 water-level hydrographs for selected alluvial wells with those of the eight bedrock observation wells; (2) analyses of eight subsurface-temperature profiles, using the analytical model of Bredehoeft and Papadopulos (1965) to estimate vertical flow through the bedrock that underlies the alluvial aquifer; (3) analysis of an aquifer test, using the analytical models of Neuman (1975) to determine the ratio of vertical to horizontal *hydraulic conductivity* of the alluvial aquifer and using the analytical model (the ratio method) of Neuman and Witherspoon (1972) to determine the vertical *hydraulic diffusivity* of a siltstone and sandstone that underlies the alluvial aquifer; and (4) comparisons of selected water-quality data from the alluvial aquifer with water-quality data from the eight bedrock observation wells.

This report also presents the results from a three-dimensional model of flow in the alluvial/bedrock aquifer system that was developed to evaluate the potential hydrologic effects to the alluvial aquifer of hypothetical withdrawals from the bedrock aquifers in El Paso County. Previously completed field work and reports, specifically those by Buckles and Watts (1988) on the hydrogeology of the alluvial aquifer and by Banta (1989) on the potential hydrologic effects of withdrawals from the bedrock aquifers of northern El Paso County, served as a background for this report.

Location and Description of the Study Area

The study area (fig. 1) is located in the upper Black Squirrel Creek Basin east of Colorado Springs, near the southern edge of the Denver Basin (Banta, 1989, fig. 1) and includes an area of about 176 mi^2. About 90 mi^2 of the study area is underlain by the alluvial aquifer (fig. 2).

The climate of the study area is semiarid; mean annual precipitation ranges from about 12 in. in the southern part of the study area to more than 14 in. in the northern part of the study area (Livingston and others, 1976, fig. 5). Mean annual evaporation from free-

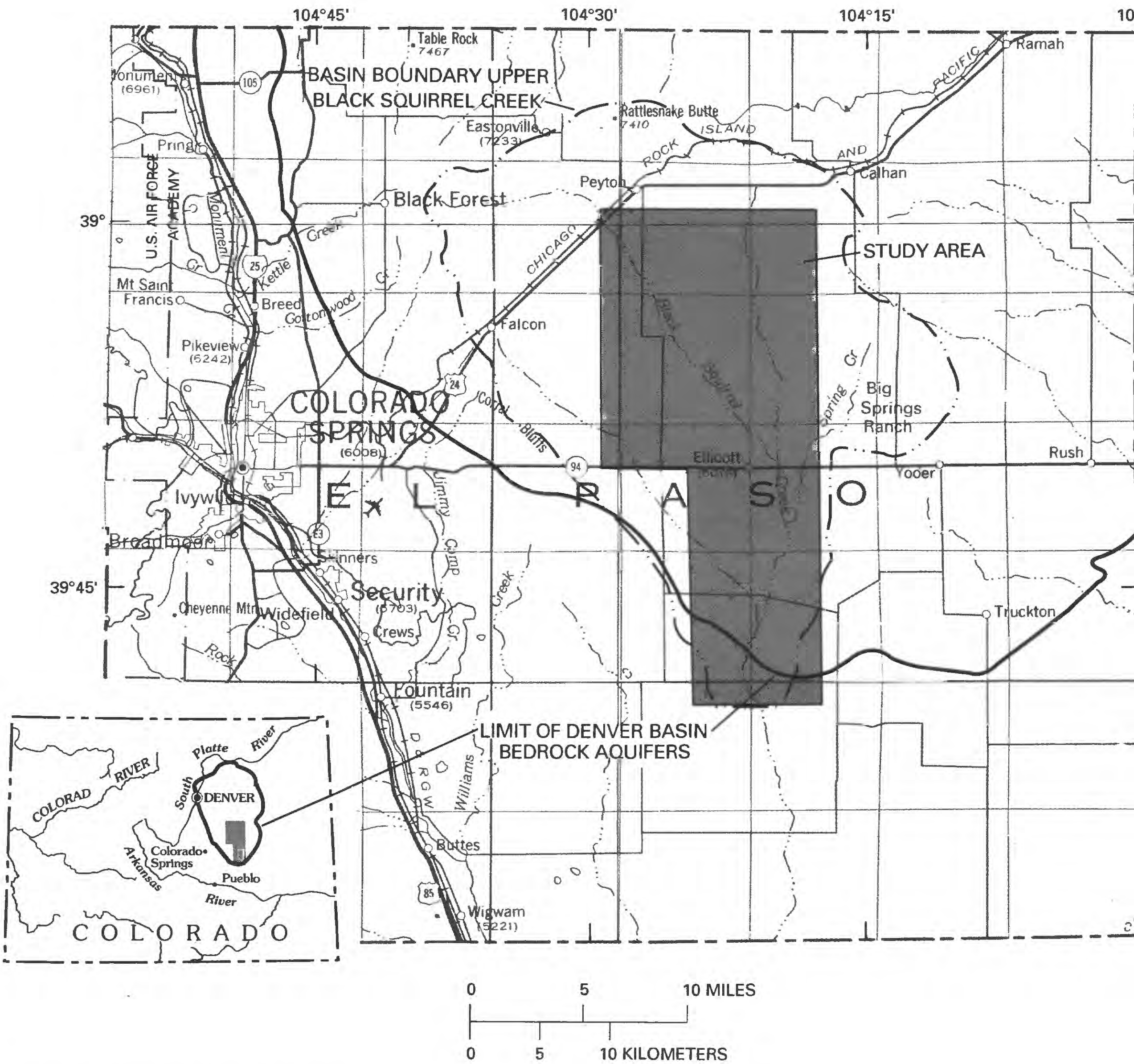


Figure 1. Location of the study area.

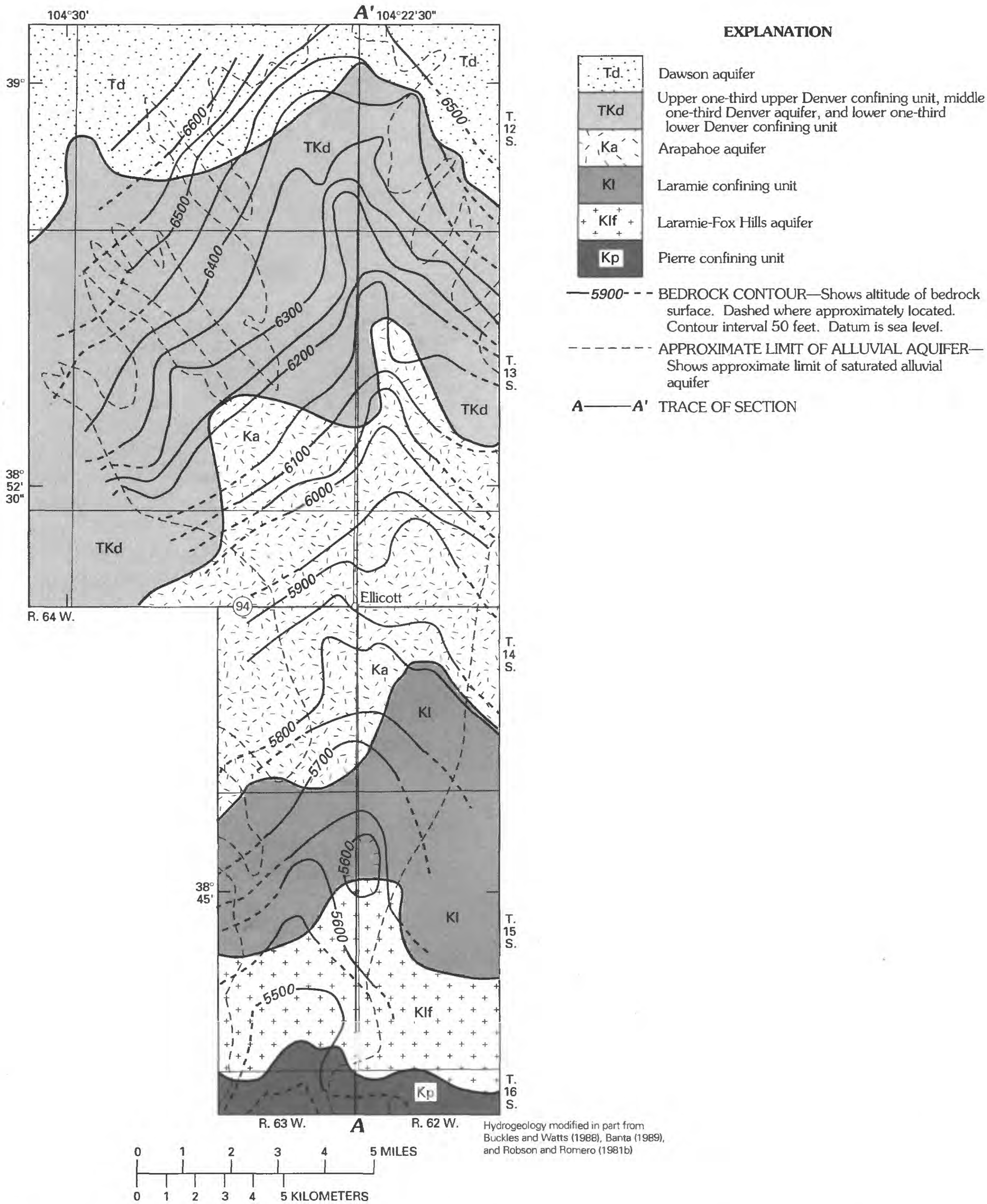


Figure 2. Generalized configuration and altitude of the bedrock surface, approximate extent of alluvial aquifer, and approximate limits of underlying hydrogeologic units.

water surfaces in the study area ranges from 50 to 70 in. (Hansen and others, 1978). Mean monthly air temperature ranges from 28 to 30°F in January and from 65 to 75°F in July (Hansen and others, 1978).

Black Squirrel Creek is an ephemeral tributary of Chico Creek, which is a tributary of the Arkansas River. Streambeds within the study area generally are dry and have sandy bottoms; runoff generally infiltrates into the sandy streambeds within the area that is underlain by the alluvial aquifer. Occasionally, after intense precipitation, some surface water is discharged from the study area. After an estimated rainfall of 14 in. on June 17, 1965, peak discharge on Black Squirrel Creek near Ellicott was an estimated 141,000 ft^3/s (Snipes and others, 1974, p. 44).

Land use in the upper Black Squirrel Creek Basin primarily is agricultural. However, urban and suburban land uses in the basin have increased in recent years. Agricultural land use includes irrigated and nonirrigated cropland, pasture, hayland, and range. During 1964–84, the principal use of water from the alluvial aquifer in the study area was to irrigate crops. During 1964–84, average annual crop consumptive use (withdrawal minus return flow) from the alluvial aquifer was an estimated 9.3 ft^3/s or about 6,750 acre-ft/yr (Buckles and Watts, 1988). Because most irrigation withdrawals occur in April–September (the irrigation season), average consumptive use was actually about 18.6 ft^3/s during the irrigation season; presumably no withdrawals for irrigation occurred during October–March. During 1964–84, an average of 3.8 ft^3/s or about 2,750 acre-ft/yr (Buckles and Watts, 1988) of ground water was exported from the basin for municipal use. There is limited use of ground water from the bedrock aquifers in the study area for domestic, stock, and irrigation supplies.

Acknowledgments

The author acknowledges the cooperation of the following individuals for allowing installation of observation wells on their property: Mr. and Mrs. Joe Farmer, Mr. and Mrs. Francis Guthrie, Mr. and Mrs. Everett Handle, Mr. Rodney Preisser, Mr. Stuart Loosely of the Cherokee Metropolitan District, and Ms. Hilma Tollefson. In addition, the author wishes to acknowledge the following volunteers who aided in data collection and analyses during this study: Mrs. Marilyn Penley, for her assistance during an aquifer test; Ms. Milly Powell, Professor of Geology at the University of Southern Colorado, Pueblo, who described drill cuttings; and Mr. Richard McCullough, U.S. Geological Survey retiree and volunteer, who interpreted geophysical logs of test holes and wells.

HYDROGEOLOGIC SETTING

The study area is underlain by a sequence of sedimentary rocks of Cenozoic, Mesozoic, and Paleozoic age, which reach a maximum thickness of about 15,000 ft northwest of the study area near the Elbert-El Paso county line (Robson and Banta, 1987, pl. 1). However, this study is concerned only with ground-water conditions in unconsolidated deposits of Quaternary age and sedimentary rocks of Tertiary and Cretaceous age that overlie the Pierre Shale of Cretaceous age.

Unconsolidated alluvial and eolian deposits of Quaternary age unconformably overlie the slightly dipping sedimentary rocks of Tertiary and Cretaceous age in the upper Black Squirrel Creek Basin on the south-central flank of the Denver Basin. The unconsolidated silty and clayey sand and gravel in alluvial and eolian deposits, hereinafter referred to as unconsolidated Quaternary deposits, are the principal source of ground water pumped by irrigation, municipal, and domestic wells in the study area. Water-bearing siltstones, sandstones, and conglomerates, which underlie the unconsolidated Quaternary deposits and overlie the Pierre Shale, are secondary sources of ground water in the study area. Claystone, shale, and coal beds, which separate the water-bearing rocks, are confining or leaky confining units.

The stratigraphic and structural relations and the hydraulic properties of the heterogeneous sequence of unconsolidated deposits and underlying rocks affect flow of water in the layered aquifer system. Discussions of geology in this report are restricted to stratigraphic and structural geology and how they relate to ground-water conditions in the unconsolidated deposits and rocks that overlie Pierre Shale.

Stratigraphy

The stratigraphic sequence considered in this report includes the unconsolidated Quaternary deposits and rocks of Tertiary and Cretaceous age that overlie the Pierre Shale of Cretaceous age (table 1). Strata capable of yielding usable quantities of potable water in the study area occur in: Unconsolidated alluvial and eolian deposits of Quaternary age, Dawson Arkose of Tertiary age, Denver Formation of Cretaceous and Tertiary age, and Arapahoe and Laramie Formations and Fox Hills Sandstone of Cretaceous age. Descrip-

Table 1. Generalized correlation and description of geologic units, hydrogeologic units, and layer numbers in the numerical model of the upper Black Squirrel Creek Basin

System	Geologic unit	Description of geologic unit	Hydrogeologic unit	Layer number in numerical model
QUATERNARY	Flood-plain alluvium	Poorly sorted clay, silt, sand, and gravel; 0 to 15 feet thick along streams.	Upper Black Squirrel Creek alluvial aquifer	1
	Piney Creek alluvium	Clayey and sandy silt and silty sand; 0 to 15 feet thick.		
	Eolian deposits	Fine to very coarse sand; 0 to 40 feet thick.		
	Valley-fill alluvium	Sand and gravel; 0 to 200 feet thick.		
TERTIARY	Dawson Arkose	Interbedded conglomerate, sandstone, and shale; 200 to 900 feet thick, where not removed by erosion.	Dawson aquifer	2
	Denver Formation	Shale in upper and lower one-third; interbedded shale, claystone, siltstone, and sandstone in middle one-third; 600 to 1,000 feet combined thickness, where not removed by erosion.	Upper Denver confining unit	3
			Denver aquifer	4
			Lower Denver confining unit	5
CRETACEOUS	Arapahoe Formation	Interbedded conglomerate, sandstone, siltstone, and shale; 400 to 700 feet thick, where not removed by erosion.	Arapahoe aquifer	6
	Laramie Formation	Upper gray to black shale with coal seams and some siltstone and sandstone; 400 to 500 feet thick, where not removed by erosion. Lower fine- to medium-grained sandstone containing interbedded siltstone and shale; 50 to 100 feet thick, where not removed by erosion.	Laramie confining unit	7
	Fox Hills Sandstone	Upper very fine-grained silty sandstone; 40 to 50 feet thick, where not removed by erosion. Lower shaley siltstone containing interbedded shale; 100 to 150 feet thick, where not removed by erosion.	Laramie-Fox Hills aquifer where permeable	8
	[1]Pierre Shale	Gray to black shale; 5,000 feet thick. Locally, upper interbedded sandstone, siltstone, and shale; 100 to 200 feet thick.	Basal confining unit	Not represented in model

[1]Where permeable, the upper part of the Pierre Shale was included as part of the Laramie-Fox Hills aquifer (Robson, 1987; Banta, 1989).

tions of the stratigraphy in this report are modified from Buckles and Watts (1988) for the unconsolidated Quaternary deposits and from Robson (1987) and Banta (1989) for the rocks of Tertiary, Tertiary and Cretaceous, and Cretaceous age. Maximum thickness of the bedrock considered in this study is about 2,200 ft (Robson, Wacinski, and others, 1981, sheet 1, fig. 4). In the southern part of the study area, the bedrock has been removed by erosion and the unconsolidated Quaternary deposits are in contact with the Pierre Shale (fig. 3).

The unconsolidated Quaternary deposits in the upper Black Squirrel Creek Basin consist of: Flood-plain alluvium along stream channels, Piney Creek alluvium, eolian deposits of sand and silt, and older valley-fill alluvium. Flood-plain alluvium along stream channels consists of less than 15 ft of poorly sorted clay, silt, sand, and gravel. The Piney Creek alluvium of Holocene age ranges in thickness from 0 to 15 ft and consists of clayey and sandy silt and silty sand. In some areas, the Piney Creek alluvium overlies eolian deposits and, in other areas, valley-fill alluvium. The eolian deposits of Holocene age range in thickness from 0 to 40 ft and consist of fine to very coarse-grained sand. The valley-fill alluvium of Pleistocene age ranges in thickness from 0 to 200 ft and consists of sand and gravel. Prior to development and use of irrigation wells in the area during the 1950's and 1960's, saturated thickness of the unconsolidated Quaternary deposits locally was more than 120 ft (McGovern and Jenkins, 1966; Bingham and Klein, 1974). As of March 1984, saturated thickness of the unconsolidated deposits was 120 ft or more only near the southern boundary of the study area (Buckles and Watts, 1988, fig. 6).

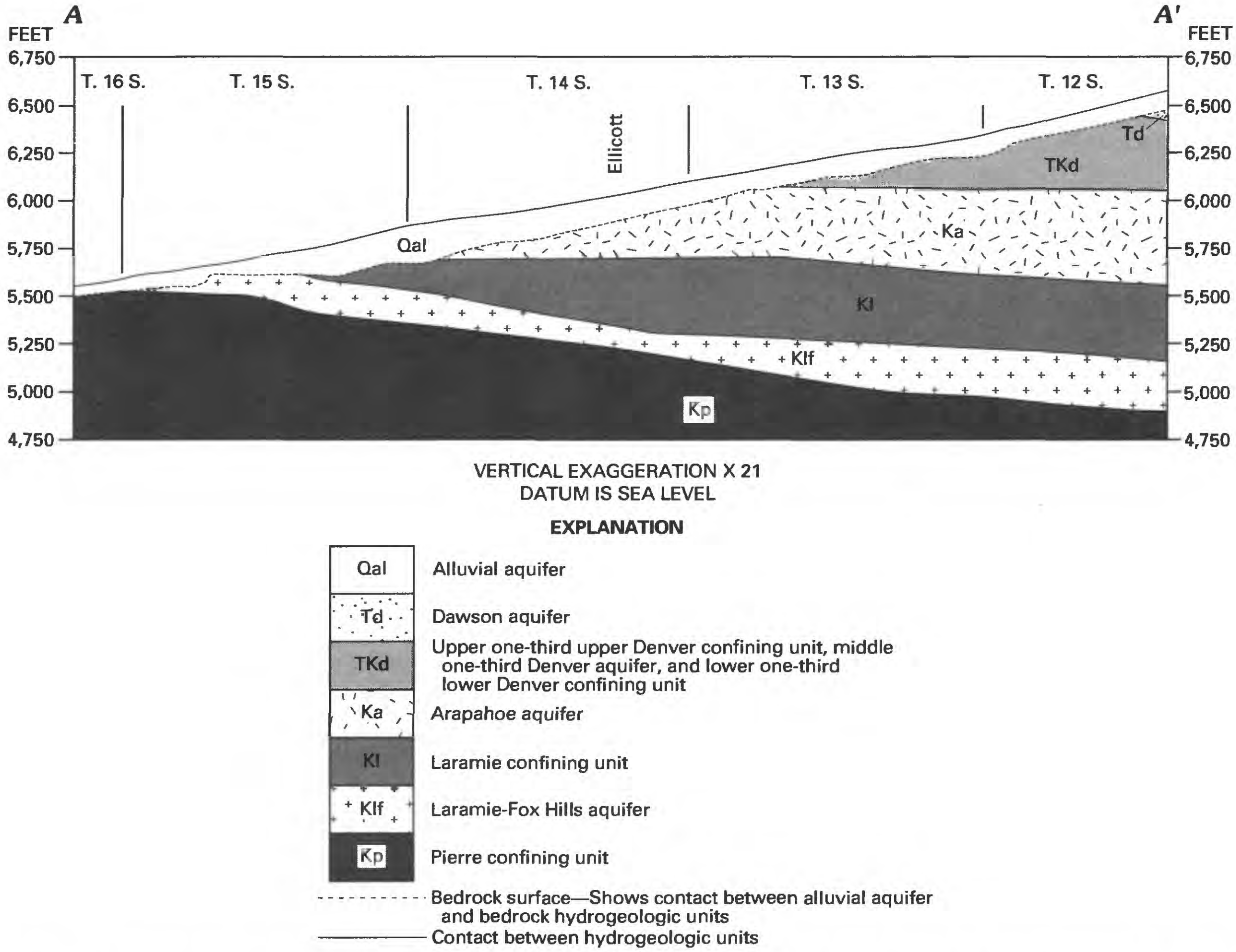


Figure 3. Generalized hydrogeologic section in the upper Black Squirrel Creek Basin (trace of section shown in figure 2).

The Dawson Arkose of Tertiary age generally ranges in thickness from 200 to 900 ft and overlies the Denver Formation. The Dawson Arkose consists of interbedded conglomerate, sandstone, and shale. Water-yielding conglomerate and sandstone range in thickness from 100 to 400 ft (Banta, 1989, p. 7). The Dawson Arkose is in contact with the alluvial deposits of the upper Black Squirrel Creek Basin only in the northern part of the study area.

The Denver Formation of Tertiary and Cretaceous ages ranges in thickness from 600 to 1,000 ft and overlies the Arapahoe Formation. The upper one-third and lower one-third of the Denver Formation are predominantly shale. The middle one-third of the Denver Formation consists of interbedded shale, claystone, siltstone, and sandstone. The thickness of water-yielding siltstone and sandstone in the middle one-third of the formation generally ranges from 100 to 300 ft (Banta, 1989, p. 7).

The Arapahoe Formation of Cretaceous age ranges in thickness from 400 to 700 ft and overlies the Laramie Formation. The Arapahoe Formation consists of interbedded conglomerate, sandstone, siltstone, and shale. Thickness of water-yielding conglomerate, sandstone, and siltstone ranges from 200 to 300 ft (Banta, 1989, p. 7).

The Laramie Formation of Cretaceous age consists of an upper 400 to 500 ft of gray to black shale containing interbedded coal seams, as much as 10 ft thick, some siltstone and sandstone, and a lower fine- to medium-grained sandstone that is 50 to 100 ft thick containing interbedded siltstone and shale. Thickness of water-yielding sandstone and siltstone in the lower part of the Laramie Formation is 50 to 100 ft. Locally, a 5- to 20-ft-thick shale bed underlies the lower sandstone unit (Robson, 1987, p. 7). The Fox Hills Sandstone of Cretaceous age is 150 to 200 ft thick and consists of an upper very fine-grained silty sandstone that is 40 to 50 ft thick and a lower shaley siltstone containing interbedded shale that is 100 to 150 ft thick (Robson, 1987, p. 7).

The Pierre Shale of Cretaceous age consists of as much as 5,000 ft of gray to black massive shale beds. Locally, the upper 100 to 200 ft of the Pierre Shale can contain sandstones that may be hydraulically connected with the Fox Hills Sandstone (Robson, 1987, p. 7).

Because a geologic unit can include several rock types that have different hydraulic properties, hydrogeologic units in table 1 are defined to include adjacent strata with similar hydraulic properties. Hydrogeologic units are classified as either aquifers or confining units, depending on the relative amount of water-bearing or confining rocks within them. For example, a sandstone that contains some shales would be considered an aquifer, and a shale that contained some sandstones would be considered a confining unit. The lithology of the alluvial aquifer primarily is sand and gravel but can include some silt and clay; the lithology of the bedrock aquifers primarily is siltstone, sandstone, and conglomerate but can include some claystones, shales, and coal beds; and the lithology of the bedrock confining units primarily is claystone, shale, and coal but can include some siltstone and sandstone.

The bedrock hydrogeologic units that underlie the alluvial aquifer (Robson, 1987; Banta, 1989) include, in descending hydrogeologic order: The Dawson aquifer, upper Denver confining unit, Denver aquifer, lower Denver confining unit, Arapahoe aquifer, Laramie confining unit, Laramie-Fox Hills aquifer, and Pierre (basal) confining unit (table 1). The Pierre confining unit underlies the Laramie-Fox Hills aquifer and consists of 5,000 ft or more of relatively impermeable shale. In this report, as in previous studies (Robson 1987; Banta, 1989), the Pierre confining unit is considered to be an impermeable lower limit for ground-water flow in the Denver Basin.

The alluvial aquifer consists of the contiguous, unconsolidated alluvial and eolian deposits in the upper Black Squirrel Creek Basin. Saturated thickness ranges from 0 to 120 ft, with the larger thicknesses occurring in channels eroded into the bedrock surface (fig. 2).

Structural Relations

The angular unconformity between the unconsolidated deposits and underlying bedrock affects flow of water between the alluvial and bedrock aquifers; flow across the unconformity consists of horizontal and vertical components. The approximate extent of the alluvial aquifer and approximate limits of the underlying bedrock hydrogeologic units (fig. 2) were modified from Robson (1987, fig. 4), Buckles and Watts (1988, fig. 6), and Banta (1989, pls. 1 and 2). Because of limited subsurface data, the limits of the bedrock hydrogeologic units are poorly defined. The surface at the base of the alluvial aquifer also is the upper bedrock surface and is an erosional unconformity.

The dip of the bedrock at outcrop in the study area generally is too small for direct measurement; therefore, the dip of the bedrock was estimated from maps of the altitude of the bases of the Dawson, Denver, Arapahoe, and Laramie-Fox Hills aquifers (Banta, 1989, pls. 1 and 2). By definition, the basal surface of an aquifer might not coincide with a stratigraphic contact (surface); therefore, dip estimated from hydrogeologic maps might differ from dip estimated from geologic maps of the corresponding stratigraphic units. The estimated dip of the bedrock aquifers in the study area ranges from about 30 to 60 ft/mi (about 0.3 to 0.7°) to the north-northwest. The slope of the bedrock surface (the unconformity at the base of the alluvial aquifer) has a regional slope of about 50 ft/mi (about 0.5°) to the south. Locally, the slope of the bedrock surface exceeds 100 ft/mi. Because the bedrock aquifers dip to the north-northwest and the bedrock surface slopes to the south, the dip of the bedrock aquifers, when measured relative to the bedrock surface (fig. 2), is about the sum of the dip of the bedrock aquifers and the slope of the erosional surface, or about 80 to 110 ft/mi (about 0.9 to 1.2°). The geometry of the bedrock hydrogeologic units, which is defined by maps of the altitudes of tops and bases and by isopachus (equal thickness) maps of the bedrock aquifers and confining units (Robson and Romero, 1981a; Robson, Romero, and Zawistowski, 1981; Robson, Wacinski, and others, 1981; Banta, 1989), were not modified for this study, except in the area in which the alluvial aquifer overlies the bedrock aquifers and confining units (fig. 2). The bedrock surface locally is better defined by data than are the surfaces at the top and bases of the bedrock hydrogeologic units. Therefore, in the area in which the bedrock aquifers underlie the alluvial aquifer, the altitude of the tops and bases of the bedrock hydrogeologic units were assumed to not exceed the altitude of the bedrock surface (fig. 2).

Aquifers and Confining Units

The alluvial aquifer consists of moderately permeable sand and sandy gravel that contains silt and clay. The bedrock aquifers consist of slightly to moderately permeable siltstone, sandstone, and conglomerate that contain interbedded shale, claystone, and lignite. The bedrock confining units (table 1) consist of relatively impermeable shale, claystone, and coal beds that contain interbedded siltstone and sandstone.

Values of the hydraulic conductivity, *specific storage*, and *specific yield* of the alluvial and bedrock aquifers and bedrock confining units are listed in table 2. Where noted, these values were derived previously during calibration of numerical models of flow (Robson, 1987; Buckles and Watts, 1988; Banta, 1989) and are assumed to represent average regional values. Values of the hydraulic properties listed in table 2, which were determined from field measurements, are noted under the column heading "Type of analysis" as aquifer test(s).

Reported values of the horizontal hydraulic conductivity of the alluvial aquifer range from 48 to 147 ft/d. However, the larger values were determined by McGovern and Jenkins (1966) using the straight-line solution of Cooper and Jacob (1946) for analysis of drawdown or recovery data from the pumped wells. In a discussion of the Cooper and Jacob straight-line solution, Lohman (1979, p. 22–23) states that, "In an unconfined aquifer that *drains* very slowly or incompletely, or both, however, the results obtained by use of equation 56 (the Cooper and Jacob straight-line solution) may be badly in error." Because the aquifer test done as part of this study indicated that the alluvial aquifer exhibits delayed gravity effects and is also anisotropic, the values of horizontal hydraulic conductivity for the alluvial aquifer that were estimated by McGovern and Jenkins (1966) might be larger than the hydraulic conductivity of the alluvial aquifer. The value of horizontal hydraulic conductivity of 64 ft/d (table 2) that was estimated by Buckles and Watts (1988) during calibration of a steady-state flow model is approximately equal to the value determined by an aquifer test done during this study and is assumed to represent an average value for the alluvial aquifer.

The vertical hydraulic conductivity of the alluvial aquifer has only been determined at one location in the study area and is estimated to be 3 ft/d (table 2). The ratio of horizontal to vertical hydraulic conductivity from this test is 16 to 1. Freeze and Cherry (1979, p. 23), in a discussion of unconsolidated alluvial deposits, state, "When the average properties of large volumes are considered, the bedded character of fluvial deposits imparts a strong *anisotropy* to the system."

The average specific yield of the alluvial aquifer of 18 percent was estimated by Buckles and Watts (1988) during calibration of a model of 1964–84 transient flow in the alluvial aquifer. Comparisons of 1974–84 water-level change in the alluvial aquifer with simulated 1974–84 water-level change indicate that the ratios of simulated pumping to specific yield used by Buckles and Watts (1988) were approximately correct. It is difficult to determine specific yield of the alluvial aquifer by aquifer-test methods because delayed gravity effects require a long-term test, perhaps as long as 30 days at a constant pumping rate; well interference, boundary effects, and the inability to maintain a constant discharge also may affect long-term tests.

The specific storage of the alluvial aquifer was determined by aquifer test at one site to be about 5×10^{-5} per ft (table 2). The product of specific storage and saturated thickness equals the *storage coefficient* of the alluvial aquifer, which at the test site was about 3×10^{-3} (5×10^{-5} per ft $\times$ 60 ft $= 3 \times 10^{-3}$).

Values of the horizontal hydraulic conductivity of the bedrock aquifers were estimated by Robson (1983; 1987) and Banta (1989) through use of aquifer tests, specific-capacity tests, and laboratory analyses of undisturbed rock samples. The horizontal hydraulic conductivity of water-yielding materials in the Dawson aquifer ranges from 0.01 to 6.2 ft/d (Robson, 1987); in the Denver aquifer, from 0.01 to 8.5 ft/d; in the Arapahoe aquifer, from 0.002 to 10 ft/d; and in the Laramie-Fox Hills aquifer, from 0.01 to 7.2 ft/d (Banta, 1989, p. 9). Values of vertical hydraulic conductivity of the bedrock confining units were estimated by Banta (1989, p. 9) during calibration of a three-dimensional flow model of the Denver Basin bedrock aquifers. The estimated vertical hydraulic conductivity of the upper Denver confining unit is 4.1×10^{-5} ft/d; the lower Denver confining unit is 1.3×10^{-5} ft/d; and the Laramie confining unit is 6.2×10^{-7} ft/d (Banta, 1989, p. 25, table 2). The ratio of horizontal hydraulic conductivity (0.01 to 8.5 ft/d) of the Denver aquifer to the vertical hydraulic conductivity (8×10^{-4} ft/d) of the Denver aquifer ranges from about one to four orders of magnitude.

Specific-yield values of the bedrock aquifers primarily were estimated by laboratory measurements of undisturbed samples of water-yielding materials (Robson, 1987, p. 15) and for the Dawson aquifer, average 18 percent and range from 3.6 to 34 percent; for the Denver aquifer, average 14 percent and range from 0.2 to 29 percent; for the Arapahoe aquifer, average 18 percent and range from 3.3 to 33 percent; and for the Laramie-Fox Hills aquifer, average 20 percent and range from 4.8 to 38 percent (table 2). The estimated specific storage of the water-yielding rocks in the bedrock aquifers is about 2×10^{-6} per ft (Robson, 1983; 1987). The specific storage of the bedrock confining units is estimated to range from about 6×10^{-8} to 5×10^{-7} per ft, based on an estimated *porosity* of 1 to

Table 2. Hydraulic properties of the alluvial and bedrock aquifers and bedrock confining units in the upper Black Squirrel Creek Basin

[--, not estimated; e, estimated]

Hydrogeologic unit	Layer number in numerical model	Hydraulic conductivity (feet per day)		Specific storage (per foot)	Specific yield (percent)		Type of analysis	Source of data
		Horizontal	Vertical		Range	Average		
Alluvial aquifer	1	84–147	--	--	--	--	[1]Aquifer tests	McGovern and Jenkins (1966)
		64e	--	--	--	18e	Numerical model	Buckles and Watts (1988)
		48	3	5×10^{-5}	--	15	Aquifer test	This study
Dawson aquifer	2	0.01–6.2	--	2×10^{-6}e	3.6–34	18	Aquifer tests	Robson (1987); Banta (1989)
Upper Denver confining unit	3	--	4.1×10^{-5}e	--	--	--	Numerical model	Banta (1989)
Denver aquifer	4	.01–8.5	--	2×10^{-6}e	0.2–29	14	Aquifer tests	Robson (1987); Banta (1989)
		--	8×10^{-4}	8×10^{-7}	--	--	Aquifer test	This study
Lower Denver confining unit	5	--	1.3×10^{-5}e	--	--	--	Numerical model	Banta (1989)
Arapahoe aquifer	6	.002–10	--	2×10^{-6}e	3.3–33	18	Aquifer tests	Robson (1987); Banta (1989)
Laramie confining unit	7	--	6.2×10^{-7}e	--	--	--	Numerical model	Banta (1989)
Laramie-Fox Hills aquifer	8	.01–7.2	--	2×10^{-6}e	4.8–38	20	Aquifer tests	Robson (1987); Banta (1989)

[1] Test results probably subject to large error because delayed gravity effects were not considered.

5 percent and an estimated compressibility of 1×10^{-6} to 1×10^{-7} in^2/lb. Although specific storage of the bedrock confining units is relatively small, in comparison with that of the bedrock aquifers, over long periods of time, water is released from storage in the bedrock confining units to adjacent aquifers.

Recharge and Discharge Conditions

Recharge to the alluvial aquifer includes: (1) *Infiltration* of precipitation and surface water, (2) irrigation return flow, and (3) upward flow from the bedrock. During 1964–84, estimated average annual recharge from infiltration and upward flow totaled about 13 ft^3/s or about 9,450 acre-ft/yr (Buckles and Watts, 1988, p. 41). The estimated rate of recharge (Buckles and Watts, 1988) is larger in the northern part of the study area than in the southern part (fig. 4). In the northern part of the study area, average annual precipitation is largest, infiltration of surface water (runoff) from upland areas is more likely, and upward flow from the Denver and Arapahoe aquifers to the alluvial aquifer might occur. In the southern part of the study area, average annual precipitation is smallest, infiltration of surface water (runoff) is less likely, and downward flow to the Laramie-Fox Hills might occur (fig. 4).

Discharge from the alluvial aquifer includes: (1) Pumpage for irrigation and municipal supply, (2) *evapotranspiration*, (3) discharge to streams, (4) underflow, and (5) downward flow to the bedrock. The estimated 1964–84 discharge of irrigation and municipal wells averaged about 12.4 ft^3/s (about 9,000 acre-ft/yr) and included an average of 8.6 ft^3/s (about 6,250 acre-ft/yr) of estimated irrigation pumpage, estimated as crop consumptive use, and an average of 3.8 ft^3/s (about 2,750 acre-ft/yr) of municipal pumpage that was exported from the basin (Buckles and Watts, 1988, p. 16).

Evapotranspiration of ground water occurs in areas in which the water table is near land surface and in the root zone of phreatophytes (fig. 4). In one area that is about 7 to 8 mi north of Ellicott, the topographic map (Haegler Ranch, 1954) shows an area of marshy ground and numerous springs; in another area about 4 to 6 mi southeast of Ellicott and along Black Squirrel Creek, ground water was less than 10 ft below land surface in 1964 (McGovern and Jenkins, 1966). Prior to large water-level declines, evapotranspiration discharged water from the alluvial aquifer in these areas and in areas where crops were naturally subirrigated. Buckles and Watts (1988, p. 39–40) state that, "Prior to 1950 when conditions in the alluvial aquifer were steady state, simulated ground-water evapotranspiration was about 43 percent of the total outflow (table 5). By 1964, simulated ground-water evapotranspiration represented less than 10 percent of the total outflow, and by 1984, it was less than 3 percent of total outflow***."

Discharge to streams (drains) was a minor component of the water budget of the alluvial aquifer and is less than 1 percent of the 1964–84 average discharge from the alluvial aquifer (Buckles and Watts, 1988, p. 47). Underflow out of the alluvial aquifer at the southern limit of the study area is estimated to discharge about 6.7 ft^3/s (about 4,850 acre-ft/yr) from the alluvial aquifer. Flow across the southern limit of the study area is assumed to have remained fairly constant because there are few irrigation and municipal wells in the southern part of the study area and, thus, water levels have remained relatively steady. Discharge from the alluvial aquifer, as downward flow to the bedrock, occurs locally and is most likely in the southern part of the study area where the alluvial aquifer overlies the Laramie-Fox Hills aquifer.

FLOW BETWEEN THE ALLUVIAL AND BEDROCK AQUIFERS

Previous estimates of the rate of flow between the alluvial and bedrock aquifers in the upper Black Squirrel Creek Basin vary in magnitude and direction. Estimates of discharge from the alluvial aquifer to the bedrock aquifers ranged from 4.1 to 7.4 ft^3/s or about 3,000 to 5,350 acre-ft/yr (Erker and Romero, 1967; Goeke, 1970; Waltz and Sunada, 1972). Estimates of discharge from the bedrock aquifers to the upper Black Squirrel Creek Basin ranged from 1.7 to 1.8 ft^3/s or about 1,200 to 1,300 acre-ft/yr (Robson, 1987; Banta, 1989). Buckles and Watts (1988) reported that, locally, discharge from the bedrock aquifers recharged the alluvial aquifer in the upper Black Squirrel Creek Basin. However, their model of flow in the alluvial aquifer did not specifically account for the hydraulic connection between the alluvial and bedrock aquifers and, thus, the rate of flow between the alluvial and bedrock aquifers was not estimated.

Although flow between the alluvial aquifer and underlying bedrock cannot be measured directly, the rate of flow can be calculated as the product of the vertical *hydraulic gradient* at the bedrock surface and the vertical hydraulic conductivity of the bedrock. Flow also can sometimes be determined indirectly from analysis of the curvature of subsurface-temperature pro-

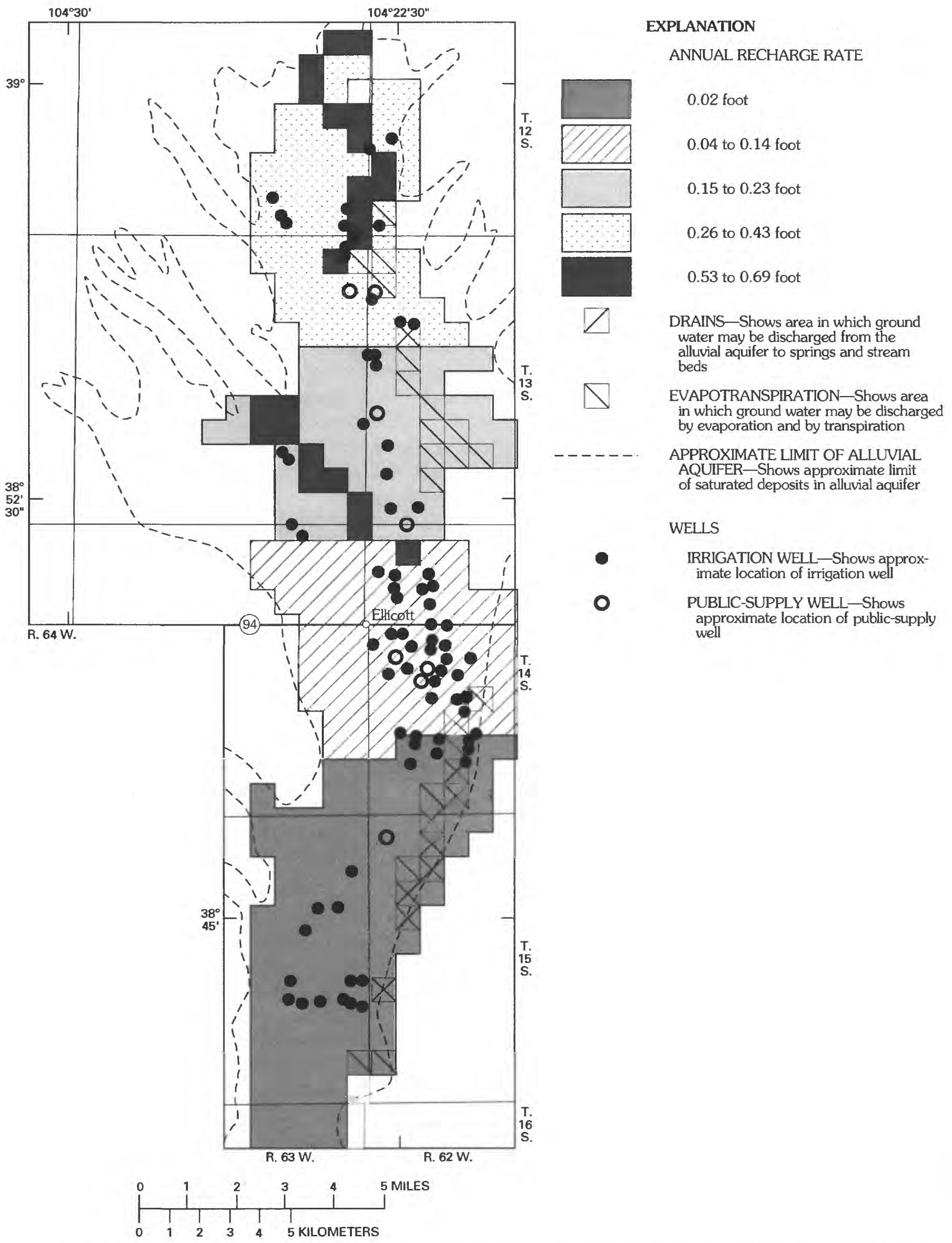


Figure 4. Estimated recharge to the alluvial aquifer, discharge areas, and locations of irrigation and municipal wells.

files. Comparisons of physical and chemical characteristics of water from adjacent hydrogeologic units can sometimes be used to qualitatively infer the direction of flow.

Water Levels and Hydraulic Gradients

Water levels in wells completed in the alluvial aquifer (alluvial wells) are measured routinely as part of ongoing data-collection programs by the Cherokee Metropolitan District (37 wells biweekly) and by the U.S. Geological Survey (29 wells monthly). Locations of eight selected alluvial observation wells with water-level measurements, that are discussed in this report, are shown in figure 5. The U.S. Geological Survey also measures water levels monthly in the eight bedrock observation wells (fig. 5) that were installed during this study. The system of numbering wells, used throughout this report, is based on the Bureau of Land Management's system and is explained in the "System of Numbering Wells" section at the back of the report.

Water levels in the alluvial aquifer locally declined more than 50 ft when pumping for irrigation and municipal use increased substantially during 1949–90. During 1964–74, water levels in the alluvial aquifer declined 20 to 35 ft in the areas of greatest pumping, northeast and southeast of Ellicott (fig. 6A, modified from Bingham and Klein, 1974, pl. 1), and during 1974–84, water levels in the alluvial aquifer declined 21 to 30 ft in an area about 1 mi wide and 4 to 7 mi north of Ellicott (fig. 6B, modified from Buckles and Watts, 1988, fig. 11).

Water-level hydrographs (figs. 7A–14A) for the eight alluvial/bedrock well pairs (fig. 5) can be used to calculate the potential for vertical flow; they also can be used to qualitatively describe the degree of hydraulic connection between the alluvial aquifer and underlying bedrock. The relative accuracy of water-level altitudes is only 10 ft for alluvial/bedrock well pair SC01206230BBC and SC01206219CCC2 (fig. 7A) but is 0.01 ft for the remaining seven well pairs (figs. 8A–14A).

Water levels in alluvial well SC01206230BBC (fig. 7A) show a seasonal pattern of water-level decline during the spring and summer irrigation season, followed by recovery during the fall and winter. Water levels in the upper Denver confining unit at well SC01206219CCC2 (fig. 7A) show little change in water levels after March 1988. The lack of correlation between water-level changes in these wells indicates that there is little hydraulic connection between the alluvial aquifer and upper Denver confining unit at these wells.

Water levels in alluvial wells SC01306207BCB2 (fig. 8A), SC01306230ACC2 (fig. 9A), and SC01406205CAA4 (fig. 10A) show cyclic fluctuations caused, in part, by pumping of nearby municipal wells; water levels in nearby bedrock wells SC01306207BCB4 (lower Denver confining unit), SC01306230ACC4 (Arapahoe aquifer), and SC01406205CAA5 (Arapahoe aquifer) show similar patterns of fluctuation. Cyclic pumping of large-capacity wells completed in the alluvial aquifer may cause temporary reversals in the direction of flow between the alluvial and bedrock aquifers. The similar patterns of water-level fluctuation in the alluvial aquifer and underlying bedrock hydrogeologic units indicate that either there is a hydraulic connection or that the hydrogeologic units respond to similar patterns of recharge and discharge. Because withdrawals by wells are a substantial discharge from the alluvial aquifer and there are few large-capacity wells in the upper Black Squirrel Creek Basin that are completed in bedrock hydrogeologic units, except the Laramie-Fox Hills aquifer, it is likely that the similar water-level fluctuations result from hydraulic connection between hydrogeologic units.

Water levels in alluvial well SC01406216CDB and bedrock well SC01406216CDB2 (Arapahoe aquifer) have similar trends of declining water levels (fig. 11A). Even though these wells are in an area in which the alluvial and Arapahoe aquifers are not pumped intensively, water levels declined about 2 ft during 1987–90. The declines probably resulted from regional water-level declines (depletion of storage) in the alluvial aquifer. The similarity of water-level trends in the alluvial well and bedrock (Arapahoe aquifer) well indicates that the alluvial and Arapahoe aquifers probably are hydraulically connected.

Water levels in the alluvial well SC01406229BBBA (fig. 12A) show an annual water-level decline of about 1 to 2 ft during 1987–90, whereas the seasonal water-level fluctuations in the Laramie-Fox Hills aquifer at bedrock well SC01406229BBB3 were about 70 to 80 ft. The large water-level fluctuations in bedrock well SC01406229BBB3 probably are caused by pumpage of nearby irrigation wells, which are completed in the Laramie-Fox Hills aquifer. The hydraulic connection between the alluvial and Laramie-Fox Hills aquifer at these wells is limited because the Laramie confining unit is about 170 ft thick. Where the Laramie confining unit is sufficiently thick, short-term fluctuations in the water levels in either the alluvial aquifer or the Laramie-Fox Hills aquifer will not substantially affect water levels in the other aquifer.

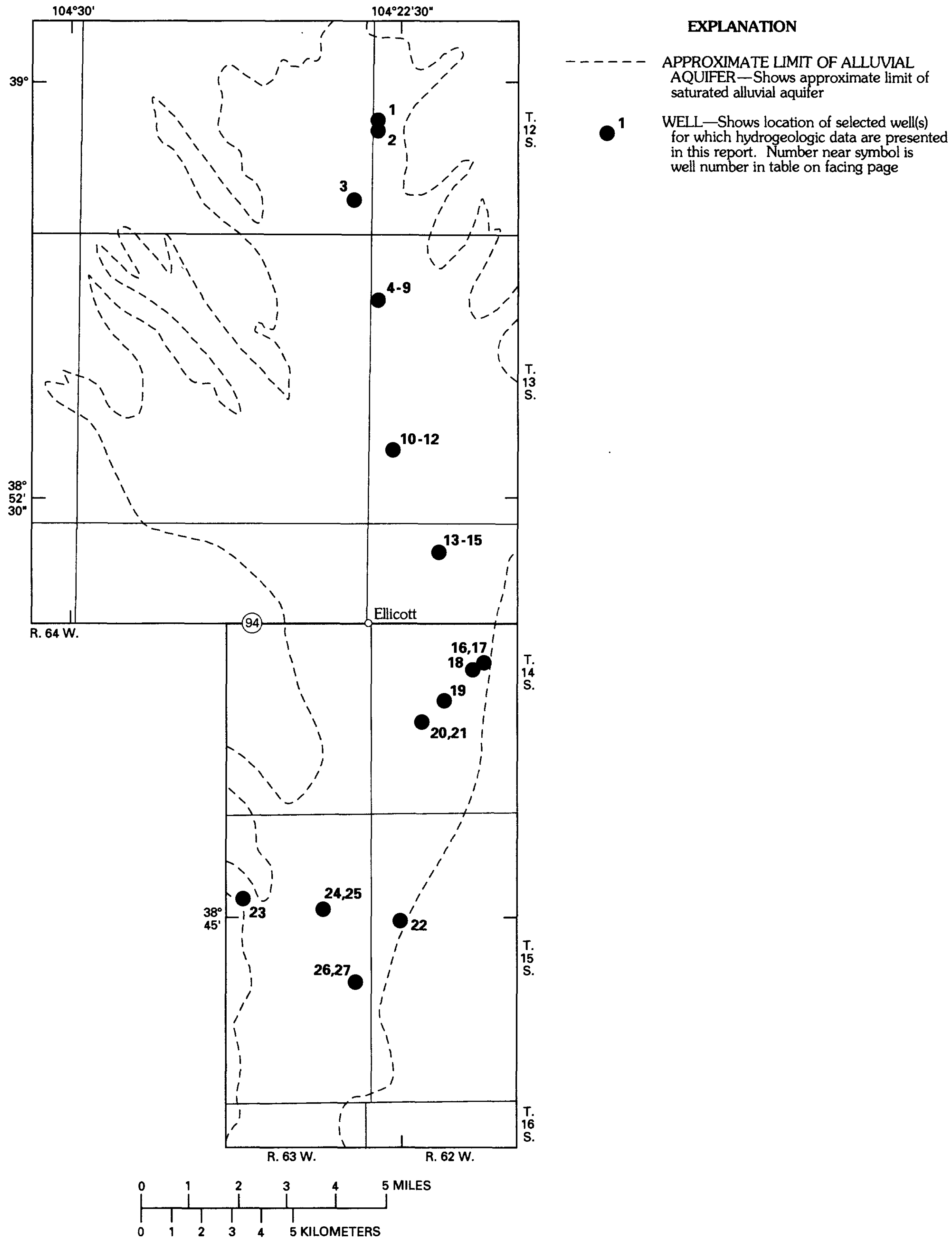


Figure 5. Locations of selected alluvial and bedrock wells in the upper Black Squirrel Creek Basin.

Well number	Local number	Hydrogeologic unit	Type of hydrogeologic data			
			Water level	Temperature profile	Aquifer test	Water quality
1	SC01206219CCC2	Upper Denver confining unit	X	X	--	X
2	SC01206230BBC	Alluvial aquifer	X	--	--	--
3	SC01206336ACC	Alluvial aquifer	--	--	--	X
4	SC01306207BCB	Alluvial aquifer	--	--	X	--
5	SC01306207BCB2	Alluvial aquifer	X	--	X	--
6	SC01306207BCB3	Alluvial aquifer	--	--	X	--
7	SC01306207BCB4	Lower Denver confining unit	X	X	X	X
8	SC01306207BCA1	Alluvial aquifer	--	--	X	--
9	SC01306207BCA2	Alluvial aquifer	--	--	X	--
10	SC01306230ACC1	Alluvial aquifer	--	--	--	X
11	SC01306230ACC2	Alluvial aquifer	X	--	--	--
12	SC01306230ACC4	Arapahoe aquifer	X	X	--	X
13	SC01406205CAA	Alluvial aquifer	--	--	--	X
14	SC01406205CAA4	Alluvial aquifer	X	--	--	--
15	SC01406205CAA5	Arapahoe aquifer	X	X	--	X
16	SC01406216CDB	Alluvial aquifer	X	--	--	--
17	SC01406216CDB2	Arapahoe aquifer	X	X	--	X
18	SC01406216CCC	Alluvial aquifer	--	--	--	X
19	SC01406220DBC	Alluvial aquifer	--	--	--	X
20	SC01406229BBBA	Alluvial aquifer	X	--	--	--
21	SC01406229BBB3	Laramie-Fox Hills aquifer	X	X	--	X
22	SC01506218ACB	Alluvial aquifer	--	--	--	X
23	SC01506310DCC	Alluvial aquifer	--	--	--	X
24	SC01506313BBB1	Alluvial aquifer	X	--	--	--
25	SC01506313BBB3	Laramie-Fox Hills aquifer	X	X	--	X
26	SC01506324DBA	Alluvial aquifer	X	--	--	--
27	SC01506324DBA2	Laramie-Fox Hills aquifer	X	X	--	X

Figure 5. Locations of selected alluvial and bedrock wells in the upper Black Squirrel Creek Basin--Continued.

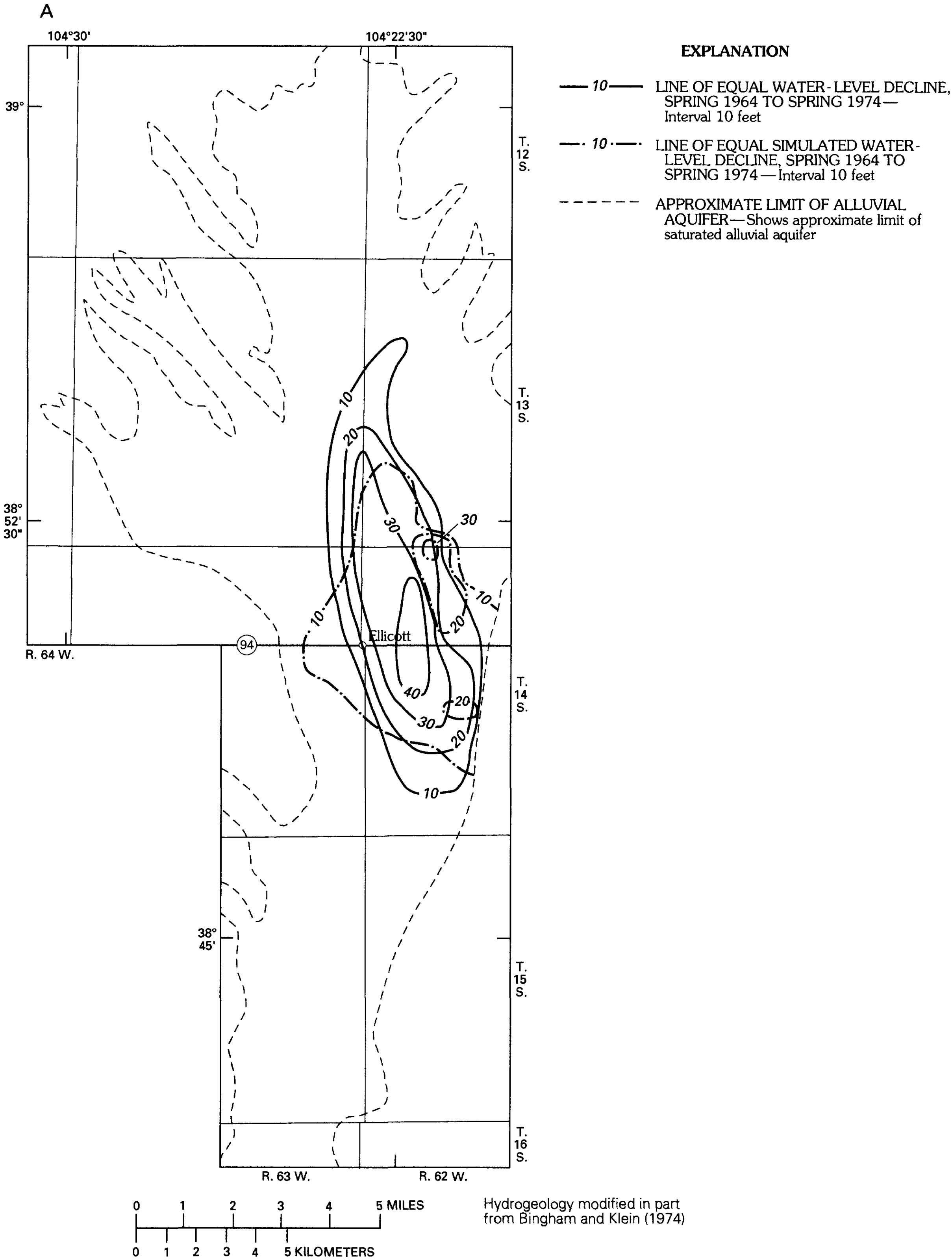


Figure 6. Measured and simulated water-level declines in the alluvial aquifer during (A) 1964–74 and (B) 1974–84.

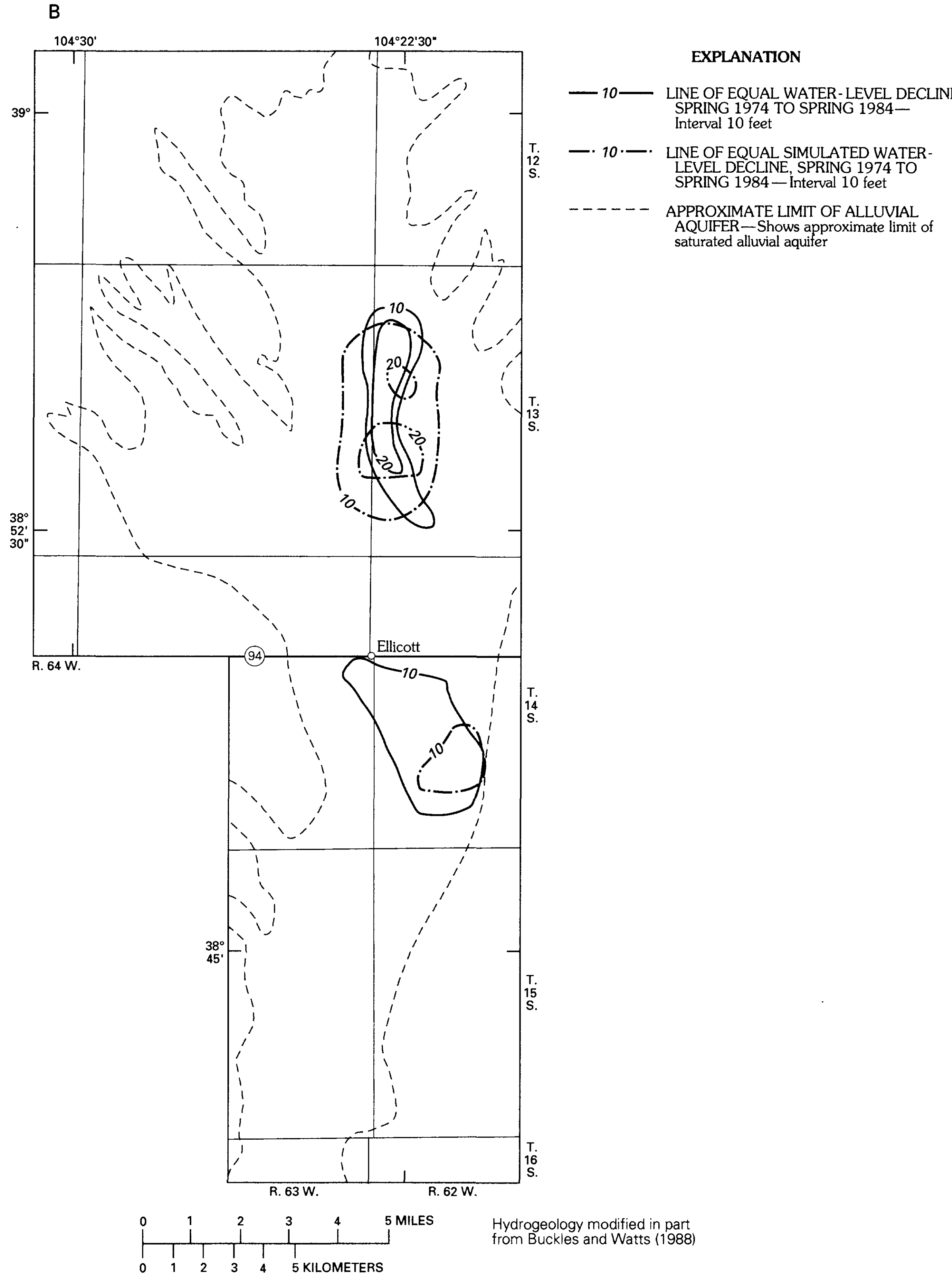


Figure 6. Measured and simulated water-level declines in the alluvial aquifer during (A) 1964–74 and (B) 1974–84 --Continued.

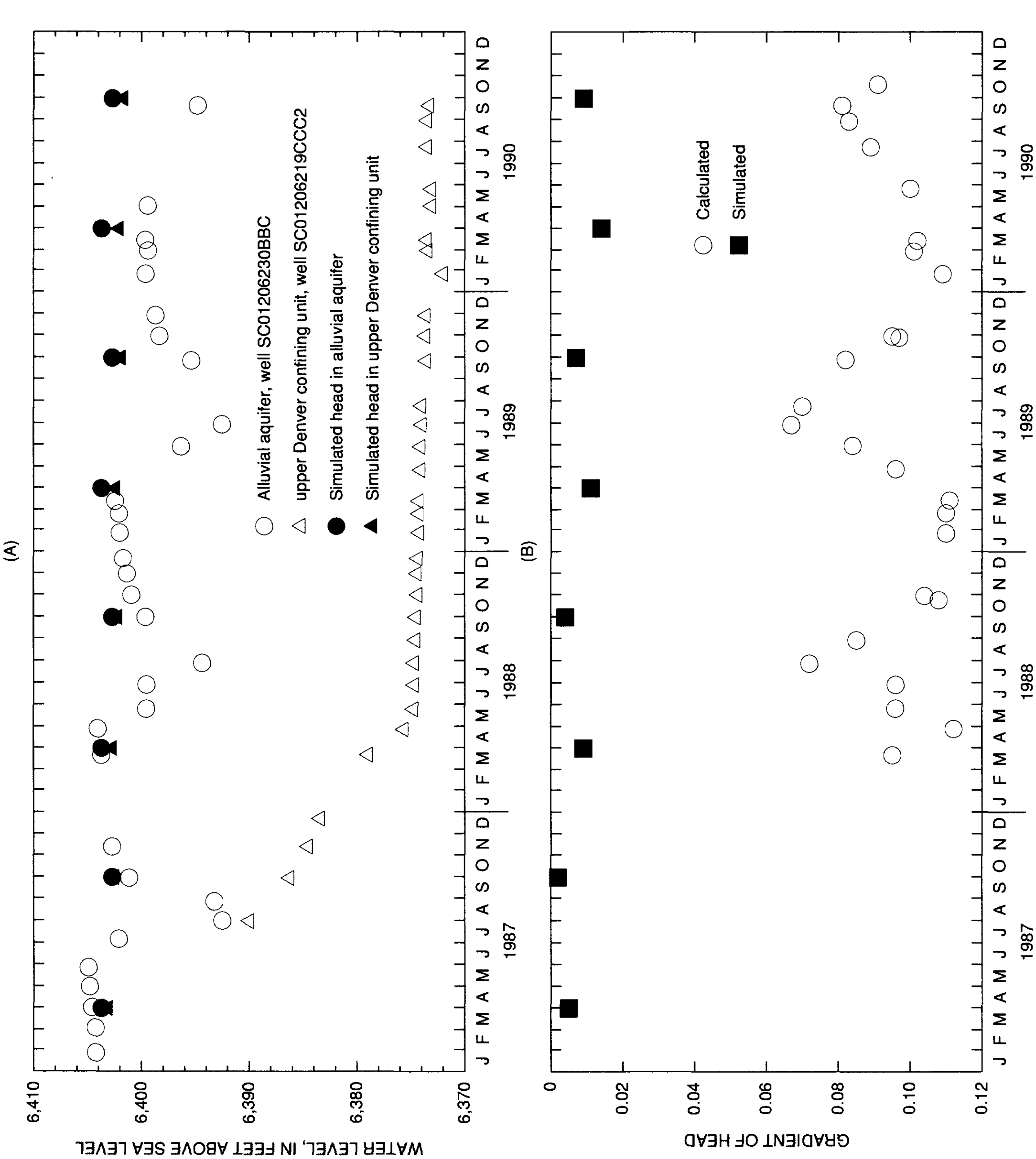


Figure 7. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the upper Denver confining unit at wells SC01206230BBC and SC01206219CCC2 in the upper Black Squirrel Creek Basin.

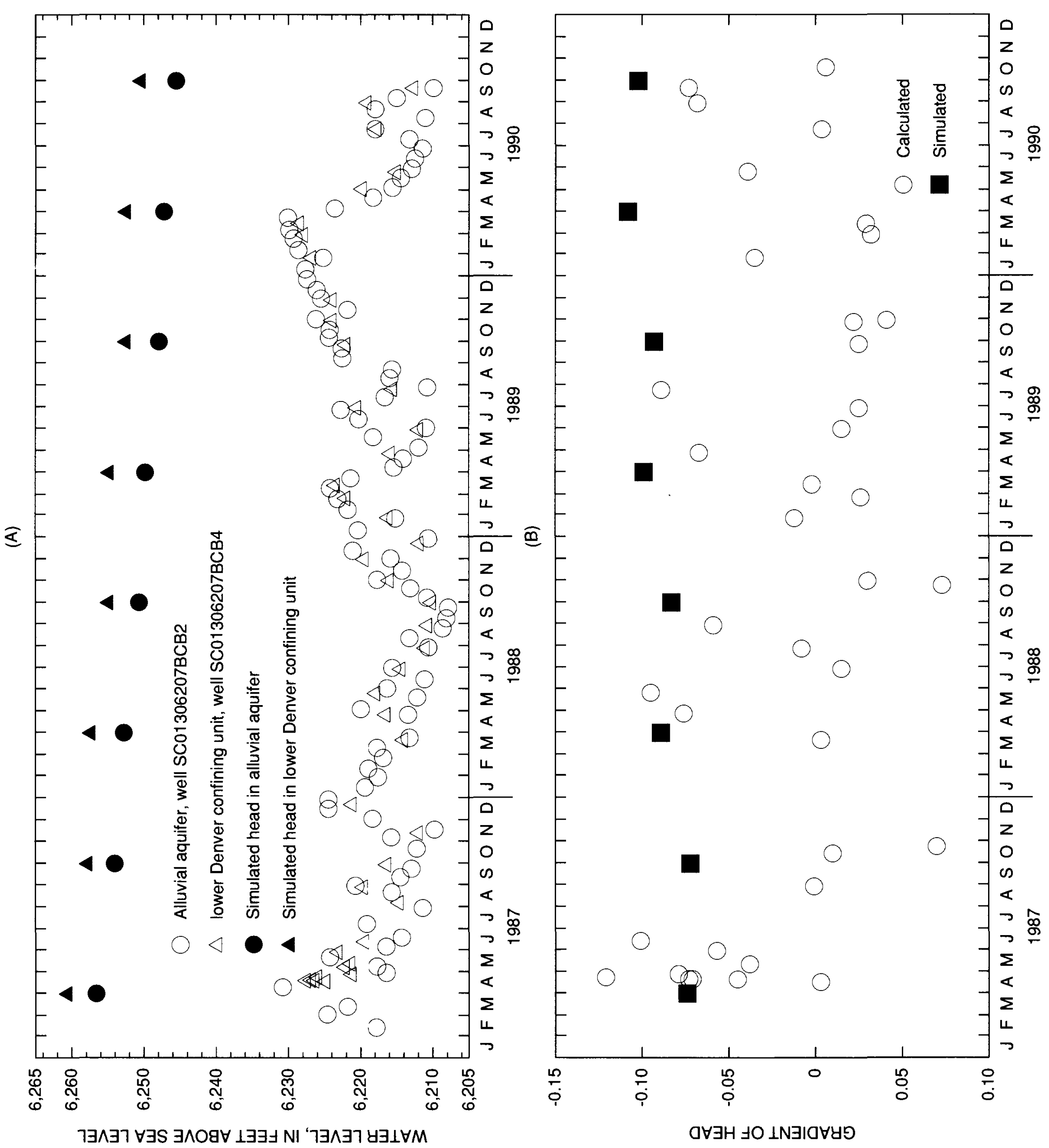


Figure 8. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the lower Denver confining unit at wells SC01306207BCB2 and SC01306207BCB4 in the upper Black Squirrel Creek Basin.

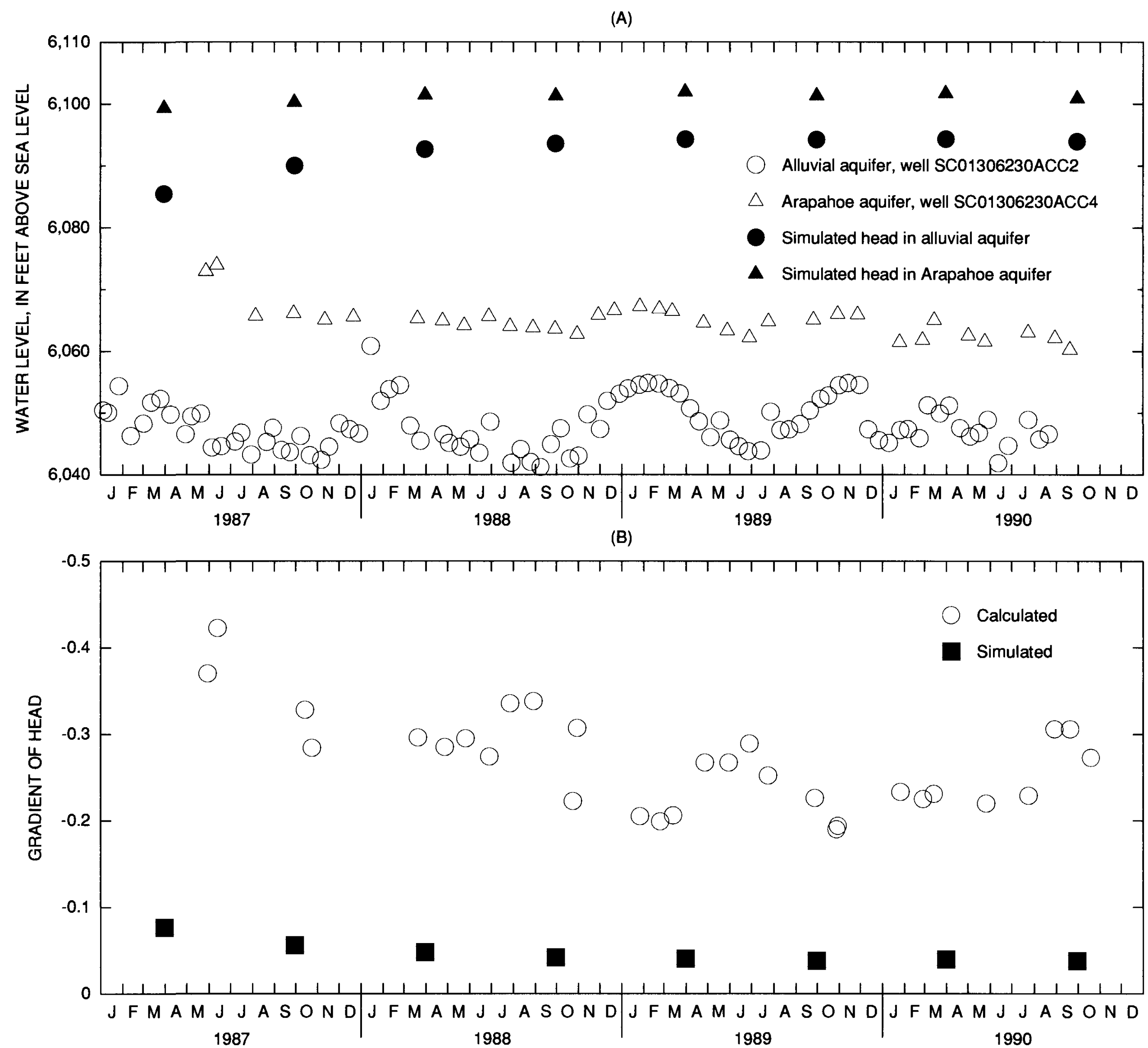


Figure 9. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the Arapahoe aquifer at wells SC01306230ACC2 and SC01306230ACC4 in the upper Black Squirrel Creek Basin.

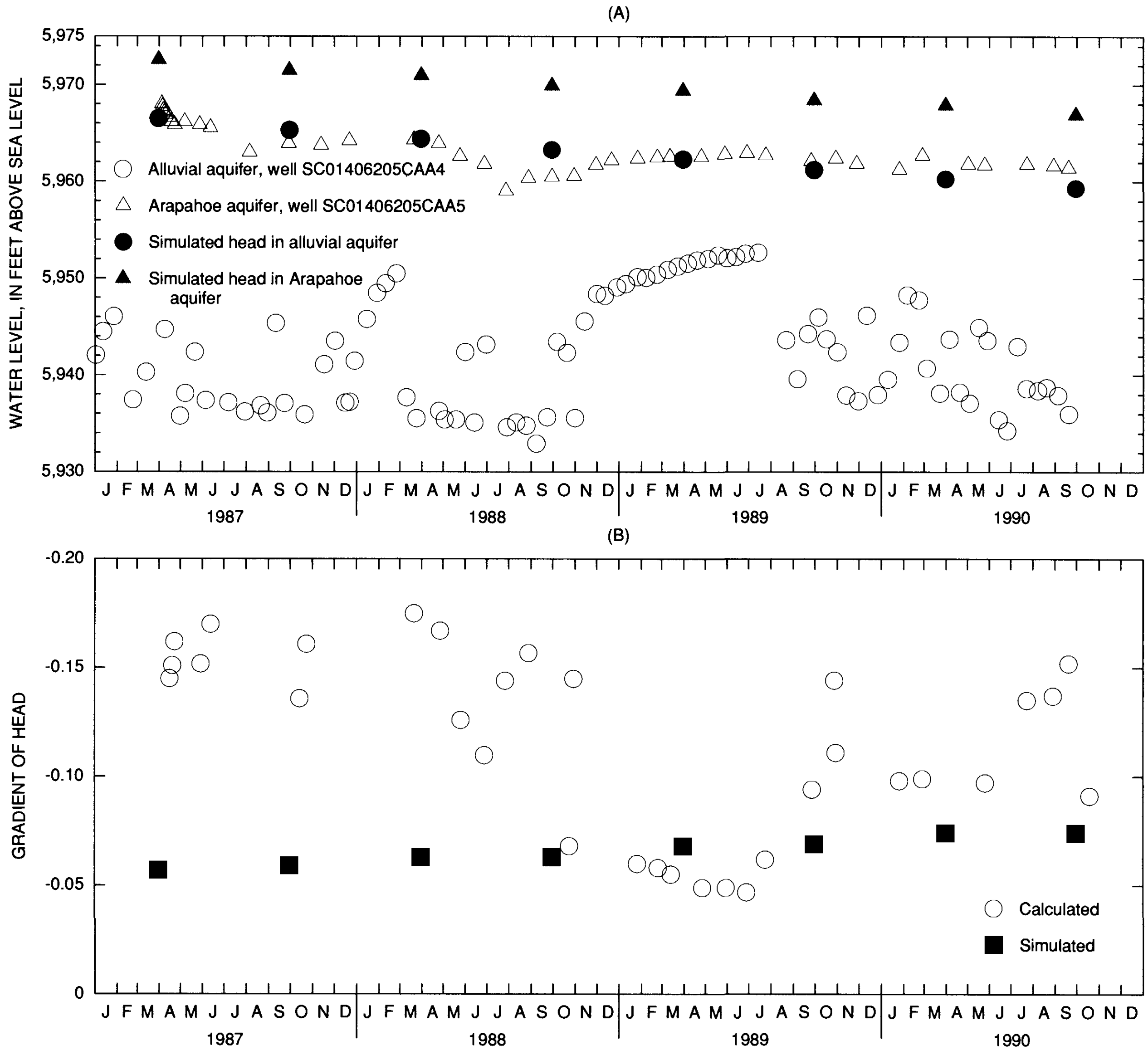


Figure 10. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the Arapahoe aquifer at wells SC01406205CAA4 and SC01406205CAA5 in the upper Black Squirrel Creek Basin.

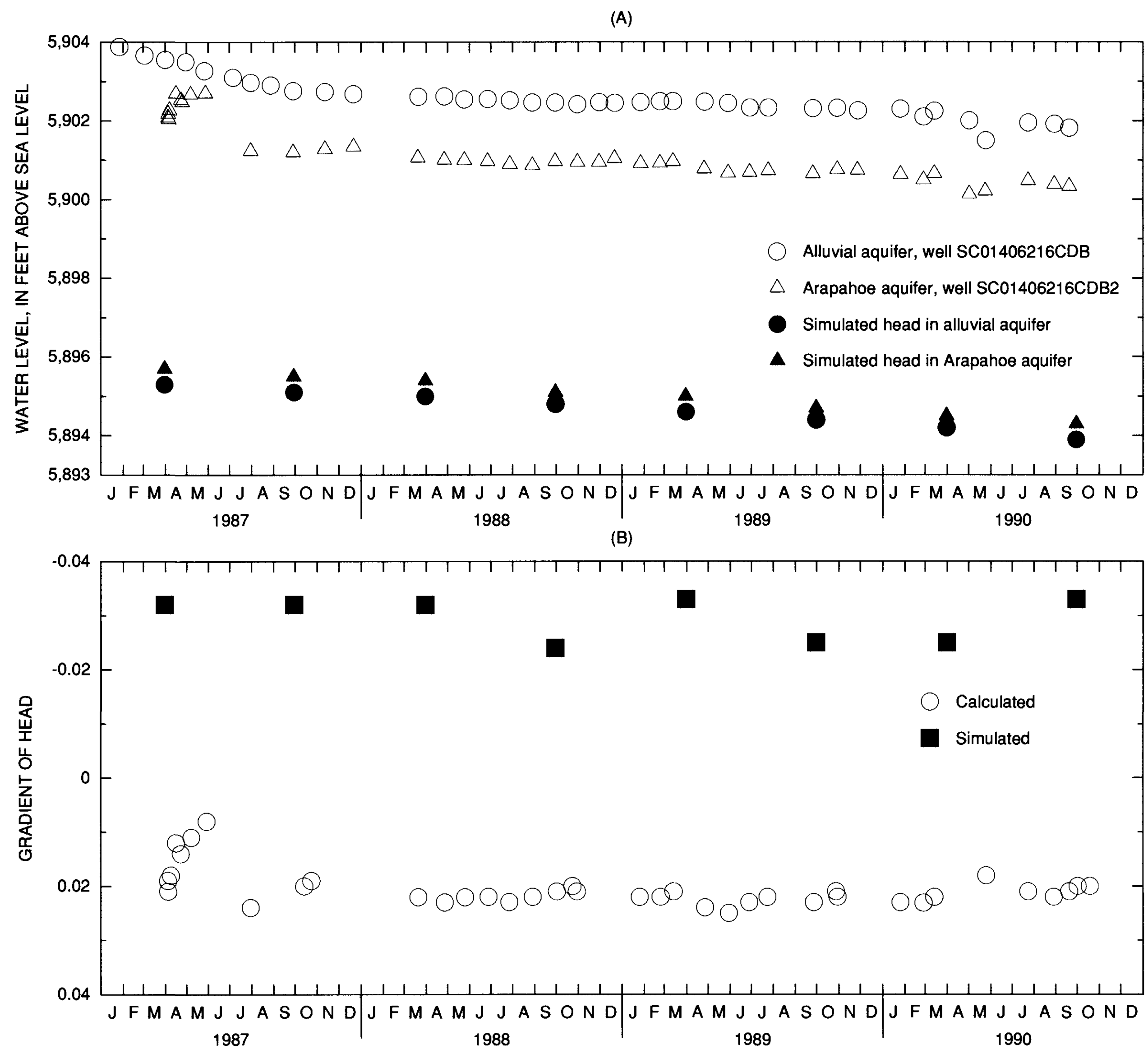


Figure 11. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the Arapahoe aquifer at wells SC01406216CDB and SC01406216CDB2 in the upper Black Squirrel Creek Basin.

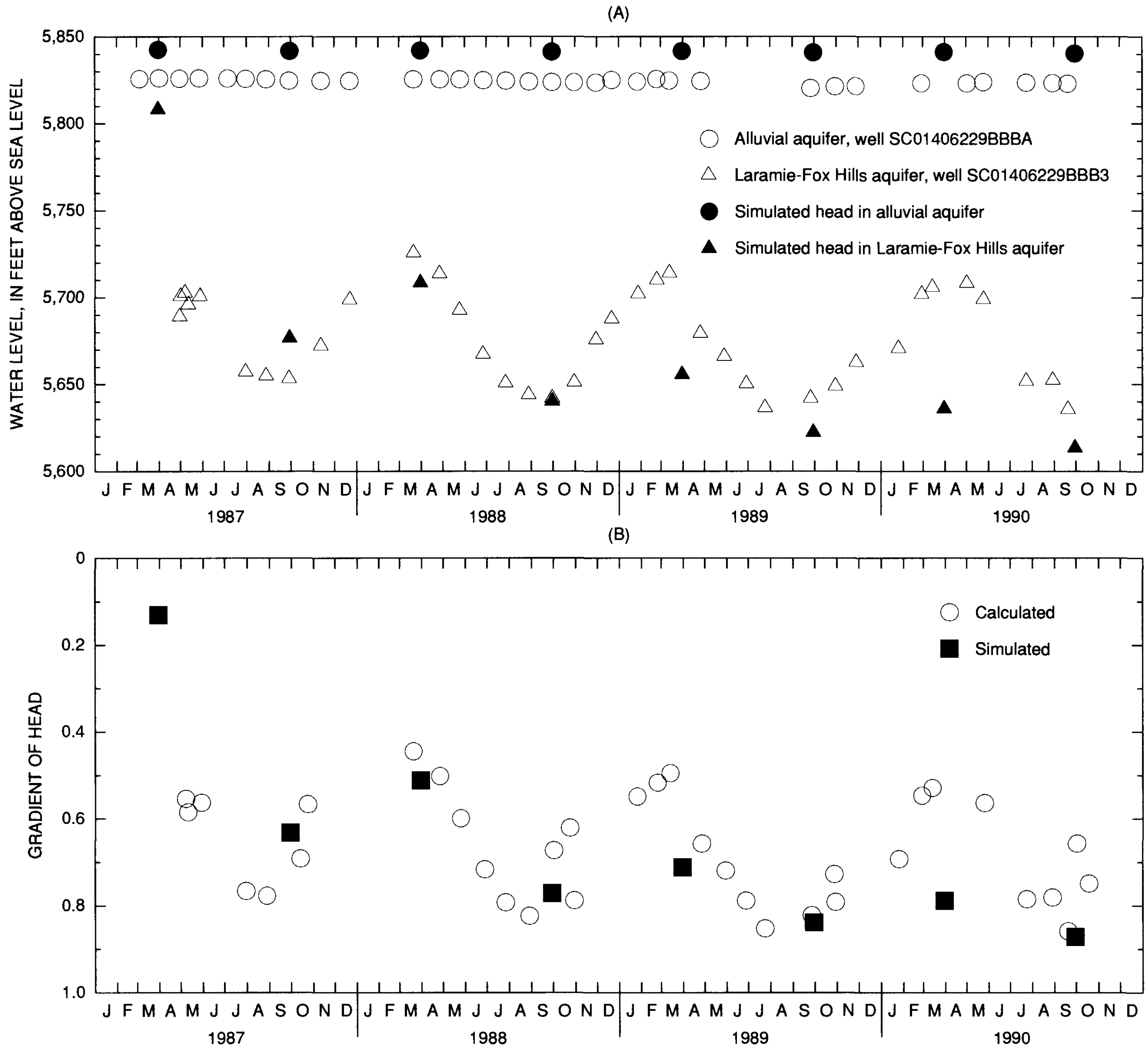


Figure 12. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the Laramie-Fox Hills aquifer at wells SC01406229BBBA and SC01406229BBB3 in the upper Black Squirrel Creek Basin.

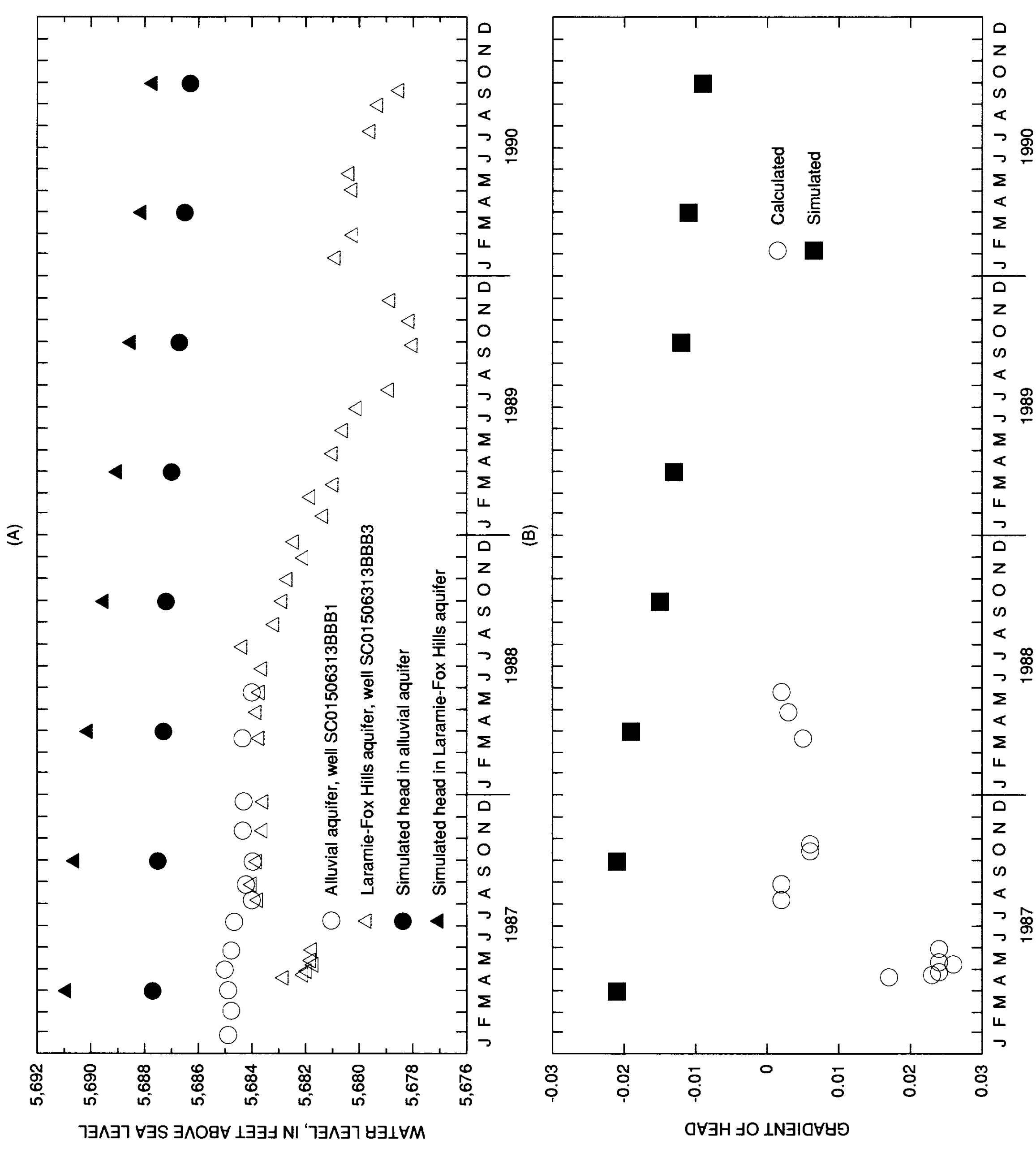


Figure 13. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the Laramie-Fox Hills aquifer at wells SC01506313BBB1 and SC01506313BBB3 in the upper Black Squirrel Creek Basin.

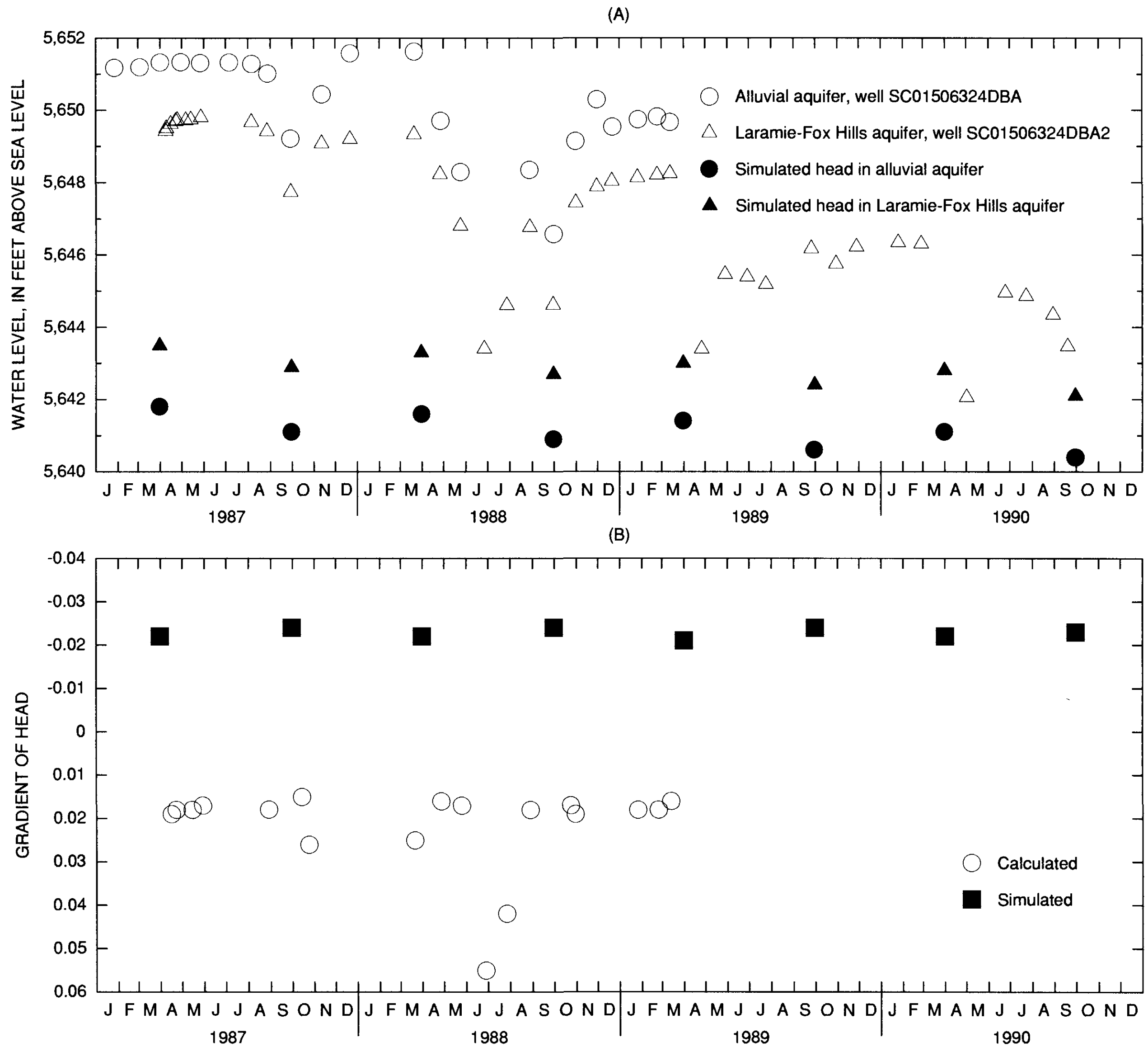


Figure 14. Measured and simulated (A) water levels in and (B) gradient of head between the alluvial aquifer and the Laramie-Fox Hills aquifer at wells SC01506324DBA and SC01506324DBA2 in the upper Black Squirrel Creek Basin.

Water levels in alluvial well SC01506313BBB1 and bedrock (Laramie-Fox Hills aquifer) well SC01506313BBB3 (fig. 13A) can only be compared for measurements made in 1987 and part of 1988 because the alluvial well was plugged during late 1988. Water levels in the two wells were similar in magnitude, which indicate that the alluvial and Laramie-Fox Hills aquifers probably are hydraulically connected in this area.

Water levels in alluvial well SC01506324DBA (an irrigation well) and the bedrock (Laramie-Fox Hills aquifer) well SC01506324DBA2 (fig. 14A) have similar patterns of seasonal fluctuation and long-term trend. In the southern part of the study area, the Laramie Formation was removed by erosion, and the alluvial aquifer directly overlies and probably is hydraulically connected with the Laramie-Fox Hills aquifer. In this area, water levels in the Laramie-Fox Hills aquifer are nearly the same as those in the alluvial aquifer.

Vertical hydraulic gradients and *gradient of head* were calculated for each pair of alluvial and bedrock wells (a well pair) as the difference, Δh, between water levels in the alluvial aquifer, h_a, and in the bedrock aquifer, h_b ($\Delta h = h_a - h_b$), divided by the distance, Δz, between the depth midway between the water table and the base of the alluvial aquifer, z_a, and the depth to the top of the open interval in the bedrock observation well, z_b ($\Delta z = z_b - z_a$). The vertical gradient of head, as used in this report, is negative when h_b is larger than h_a and positive when h_a is larger than h_b. The existence of a vertical gradient of head between hydrogeologic units does not connote vertical flow, it is only a measure of the potential for vertical flow. Under steady-state conditions and in accordance with Darcy's law, the gradient of head, $\Delta h/\Delta z$, is proportional to the ratio of *specific discharge*, q_z, and the vertical hydraulic conductivity, K_z, along the flow path, ($\Delta h/\Delta z = q_z/K_z$). By convention, a negative sign is used in mathematical expression of Darcy's law because hydraulic gradient is defined to be positive in the direction of decreasing head (Lohman and others, 1972, p. 8). The gradient of head and the hydraulic gradient are equal but of opposite sign, so that the sign of the gradient of head is negative when potential for flow is upward and the hydraulic gradient is positive when potential for flow is downward.

The differences between water levels in the alluvial aquifer and in the underlying bedrock aquifer and the values of vertical gradient of head indicate that the potential for flow between the alluvial and bedrock aquifers generally is upward in the northern part of the basin where the alluvial aquifer overlies the Denver and Arapahoe aquifers, and locally is downward in the southern part of the basin, where the alluvial aquifer overlies the Laramie confining unit and Laramie-Fox Hills aquifer. Because the Laramie confining unit directly underlies the alluvial aquifer (fig. 2) in much of the southern part of the basin, downward flow probably occurs only in areas where the Laramie confining unit was removed by erosion. Although regional discharge from the Laramie-Fox Hills to the alluvial aquifer is indicated by the configuration of the Laramie-Fox Hills aquifer's 1978 water-level surface (Robson, 1987, pl. 1), water levels in and vertical gradients of head between well pair SC01506313BB1 and SC01506313BB3 (fig. 13) and well pair SC01506324DBA and SC01506324DBA2 (fig. 14) indicate that locally the Laramie-Fox Hills aquifer is recharged by the alluvial aquifer. Toth (1963) has shown that as local topographic relief increases, local flow systems are more likely to reach the basal boundary of the aquifer system. Assuming that relief on the buried bedrock surface (fig. 2) has a similar effect on development of flow systems, then local, intermediate, and regional flow systems are likely to occur within the area in which the Laramie-Fox Hills aquifer and other bedrock aquifers underlie the alluvial aquifer.

Evaluation of Subsurface-Temperature Profiles for Estimating Specific Discharge Through Leaky Confining Units

Analysis of subsurface-temperature profiles can provide a means of indirectly determining the specific discharge vertically through the bedrock underlying the alluvial aquifer. The simultaneous transport of heat and water through rock can cause curvature of the subsurface-temperature profile (fig. 15). The temperature profile is deflected in the direction of ground-water flow; the magnitude of this deflection is proportional to the rate of flow and the thermal properties of the porous media and the thickness of the leaky interval.

Stallman (1960; 1963) presented a mathematical model describing the simultaneous flow of heat and ground water in three dimensions and suggested that temperature measurements might permit the indirect measurement of ground-water movement. Bredehoeft and Papadopulos (1965) solved Stallman's general equation for the case of steady, one-dimensional (vertical) flow of heat and water through a homogeneous leaky confining unit. Stallman (1967) modified the analytical (type-curve matching) technique of Bredehoeft and Papadopulos (1965) to increase sensitivity when the product of specific discharge and thickness of the leaky confining unit is less than about

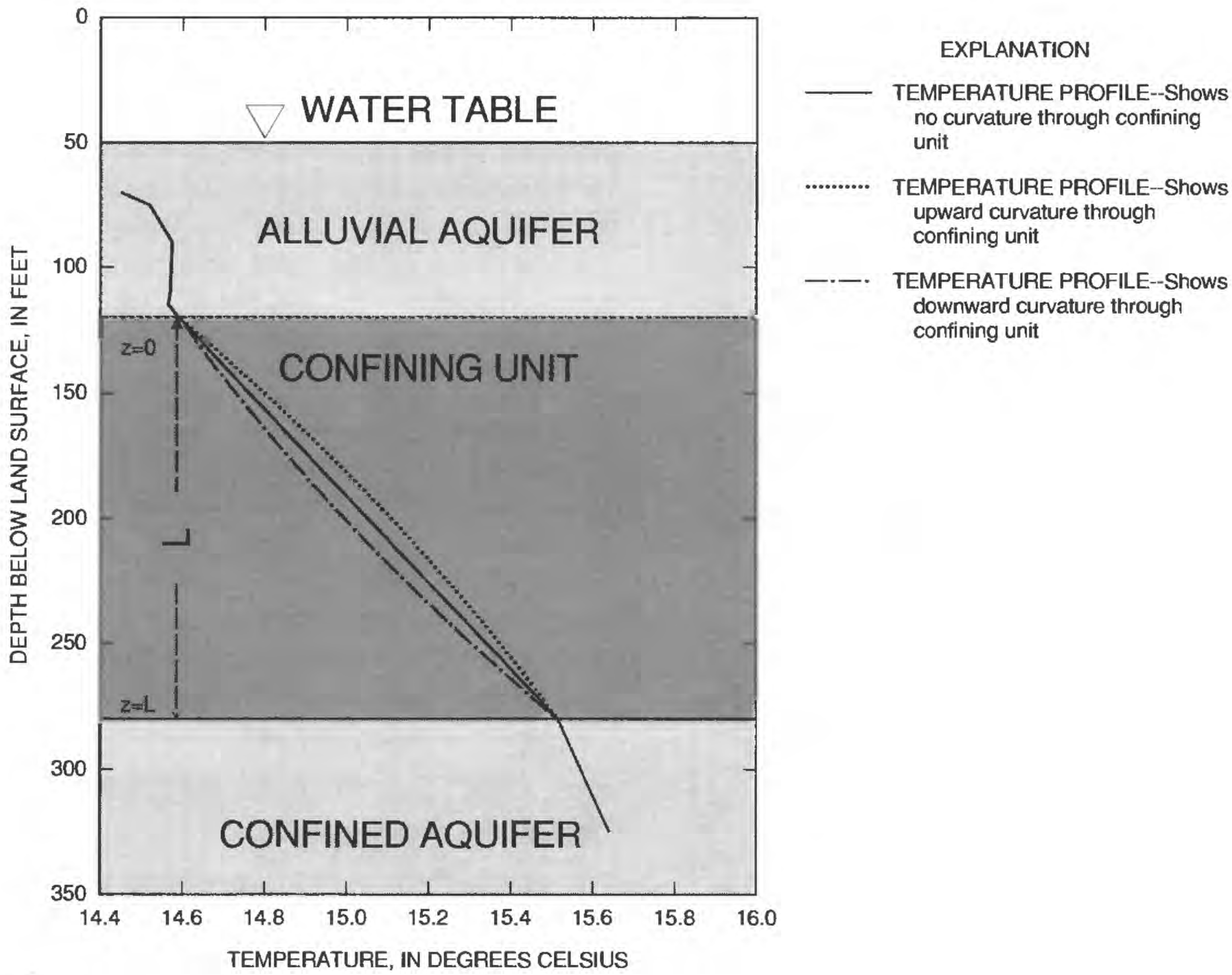


Figure 15. Hypothetical subsurface-temperature profiles.

1×10^{-3} cm^2/s or about 9×10^{-2} ft^2/d. For example, the minimum specific discharge that could be estimated through a confining unit, which is about 10 m (about 30 ft) thick is 1×10^{-6} cm/s (about 3×10^{-3} ft/d, which is about 1 ft/yr); and for a confining unit 100 m (about 300 ft) thick, the minimum specific discharge is about 1×10^{-7} cm/s (about 3×10^{-4} f/d, which is about 0.1 ft/yr).

Assumptions of the mathematical model of Bredehoeft and Papdopulos (1965) are: (1) The leaky confining unit is homogeneous, (2) flow of heat and water through the leaky confining unit is steady and only in the vertical direction, (3) ground-water velocity is defined by Darcy's law, and (4) heat is neither consumed nor generated within the leaky confining unit. From a practical standpoint, the following conditions also should be met: (1) The leaky confining unit must be sufficiently thick so that a measurable temperature gradient exists and the temperature can be measured at a minimum of three points within the unit, (2) temperatures of water in the well and in the confining unit have equilibrated, and (3) vertical flow of water does not occur between perforated zones within the well or in the annular space between the casing and borehole walls. Convective heat transport at the air/water interface also can affect the subsurface-temperature profile measured in a well. Sorey (1971, p. 965) states that, in water-filled, small-diameter (2-inch) plastic well casing "above 30 feet (from land surface), the influence of surface air temperatures produces thermal gradients large enough to cause significant convection in water-filled wells." Sorey also noted that repeated temperature measurements in shallow wells show that temperature changes in the subsurface, as a result of seasonal changes in air temperature, do not extend below 50 ft, and that below this depth, temperatures in the wells were constant in time and presumably in equilibrium with temperatures in the subsurface.

The steady specific discharge (q_z) through an interval in a homogenous leaky confining unit can be estimated from the curvature of a temperature profile if the temperatures are known at a minimum of three points in the interval. Curvature of the temperature profile through an interval in a homogenous confining unit is described by the following equation, which is modified from Bredehoeft and Papadopulos (1965, eq. 4):

$$(T_z - T_o) / (T_L - T_o) = [exp(\beta z/L) - 1] / [exp(\beta) - 1] \quad (1)$$

where

- T_z = temperature (°C) at depth z; $0 \leq z \leq L$;
- z = depth (cm) below top of interval, positive downward;
- T_0 = temperature (°C) at the top of the interval, at z = 0;
- T_L = temperature (°C) at the bottom of the interval, at z = L;
- L = thickness (cm) of the interval; and
- $\beta = q_z c_o \rho_o L / \kappa_T$ (dimensionless), where: (2)
  - q_z = the specific discharge (cm/s);
  - c_o = specific heat of water (cal/g/°C);
  - ρ_o = density of water (g/cm³); and
  - κ_T = the thermal conductivity (cal/cm/s/°C) of the water saturated porous medium.

Rearranging eq. 2 to solve for specific discharge, q_z,

$$q_z = \beta \kappa_T / (c_o \rho_o L) \quad (3)$$

The parameter β is determined graphically by superimposing a plot of the relation of dimensionless depth minus dimensionless temperature to dimensionless depth on a family of type curves (figs. 16C–23C). Dimensionless depth minus dimensionless temperature equals $[z/L - (T_z - T_0) / T_L - T_0]$, and dimensionless depth equals the ratio z/L (Stallman, 1967). Values of estimated specific discharge (q_z) through selected intervals are listed in table 3. The thermal conductivity (k_T) of the interval of thickness L, if not determined from measurements, can be estimated based on the percent quartz content and porosity of the porous medium. The specific heat of water (c_o) is about 1 cal/g/°C, and the density of water (ρ_o) is about 1 g/cm³.

Subsurface-temperature profiles (figs. 16B–23B) were measured in eight bedrock observation wells (fig. 5) during November 1987 using a borehole temperature probe. Temperature/depth data were recorded on a graphic recorder. The temperature probe was lowered into the well and allowed to equilibrate to the ambient water temperature before the temperature was measured. Water temperatures were recorded continuously as the probe was lowered to the bottom of the well at a rate of descent of about 10 ft/min. Data from just below the static water level to about 30 ft below the static water level were not used in the analyses of temperature profiles because of possible effects of air temperature in the upper part of the static water column.

The confining units for which temperature profiles were analyzed consist mainly of shale, shale containing thin sandstones and siltstones, or shaley siltstone and sandstone. The depths to the tops and bottoms of the confining units were determined from the driller's and selected geophysical logs (figs. 16A–23A). Because the thermal-conductivity values of the confining units were not known, they were estimated from a graph of thermal conductivity of water-saturated shale (Robertson, 1988, fig. 11) as a function of estimated percent quartz content and estimated porosity. Although the mineralogy of confining units in the study area has not been described quantitatively, it is assumed that the quartz content of the shales is less than 10 percent. The porosity of the confining units also has not been measured in the study area; typically the porosity of shale is less than 10 percent (Freeze and Cherry, 1979, p. 37). Thermal conductivity of water-saturated shale with quartz content of 0 to 10 percent and porosity of near 0 to 10 percent ranges from about 3.2×10^{-3} to 5.2×10^{-3} cal/cm/s/°C (Robertson, 1988, fig. 11).

Estimated values of specific discharge (q_z) through selected confining units that are penetrated by the eight bedrock wells are listed in table 3. Because the bedrock lithology, as interpreted from the geophysical logs, often consists of several water-yielding zones and confining units, more than one value is shown for some wells in table 3. If the absolute value of β was less than or equal to 0.2, then q_z through the confining unit is too small to be determined by analysis of subsurface-temperature profiles (Sorey, 1971, p. 964), and the absolute value of q_z is less than about 2.8×10^{-3} ft/d (1 ft/yr). For convenience, estimates of q_z, which were calculated using metric units, are listed in units of centimeter per second (cm/s) and feet per day (ft/d) (table 3).

One or more of the assumptions on which the analytical model of Bredehoeft and Papadopulos (1965) are based are not met because of hydrogeologic conditions at the eight bedrock wells. Homogeneous confining units are relatively thin, less than 100 ft thick, or are absent at seven wells (figs. 16A–20A, 22A and 23A), and transient flow conditions are likely to have affected water temperatures at seven wells (figs. 16B–19B and 21B–23B). Horizontal transport of heat also might have affected the temperature profiles of wells SC01506313BBB3 (fig. 22B) and SC01506324DBA2 (fig. 23B), which penetrate the Laramie-Fox Hills aquifer and for which no confining unit greater than 15 ft thick could be identified on the geophysical logs (figs. 22A and 23A).

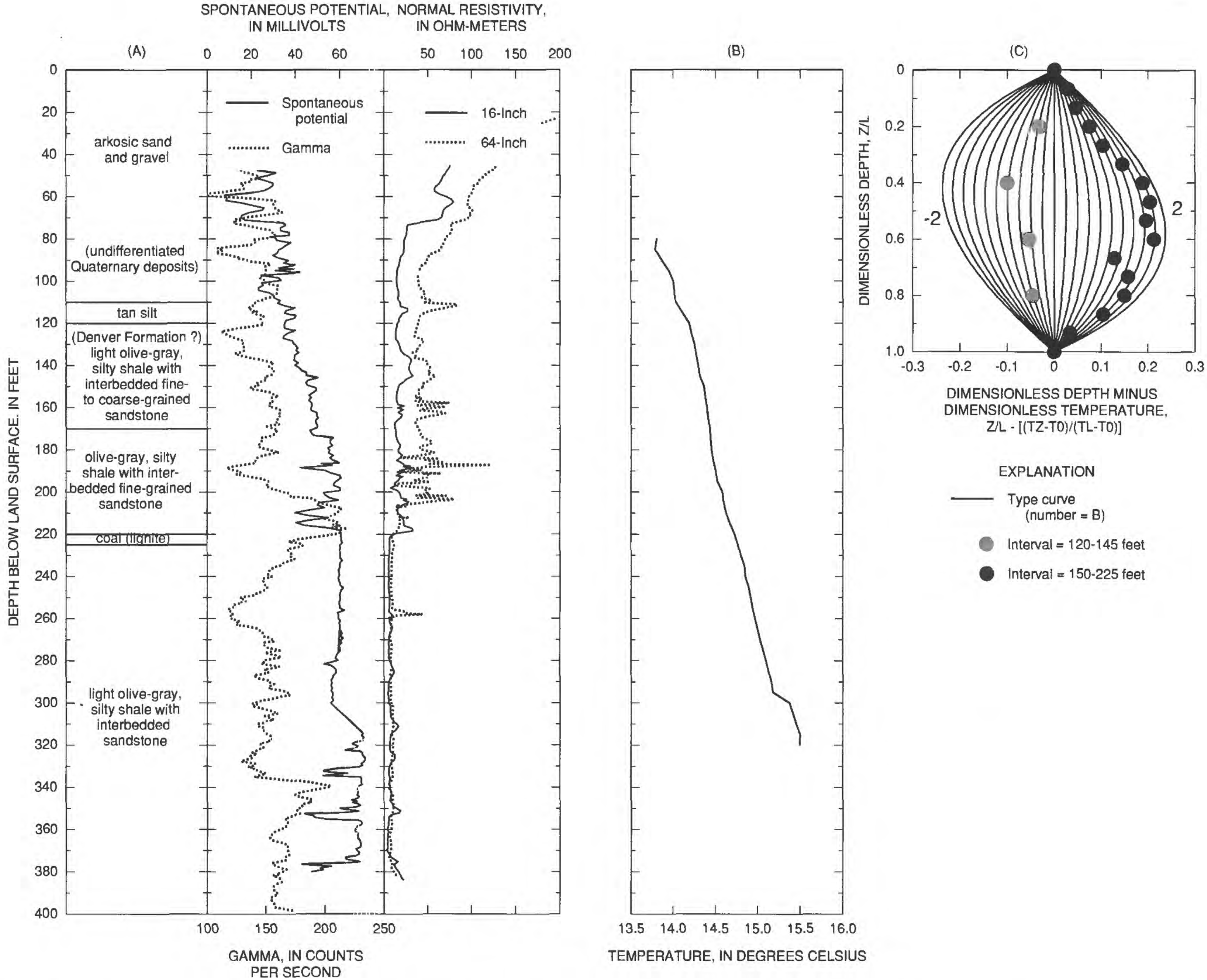


Figure 16. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01206219CCC2 in the upper Black Squirrel Creek Basin.

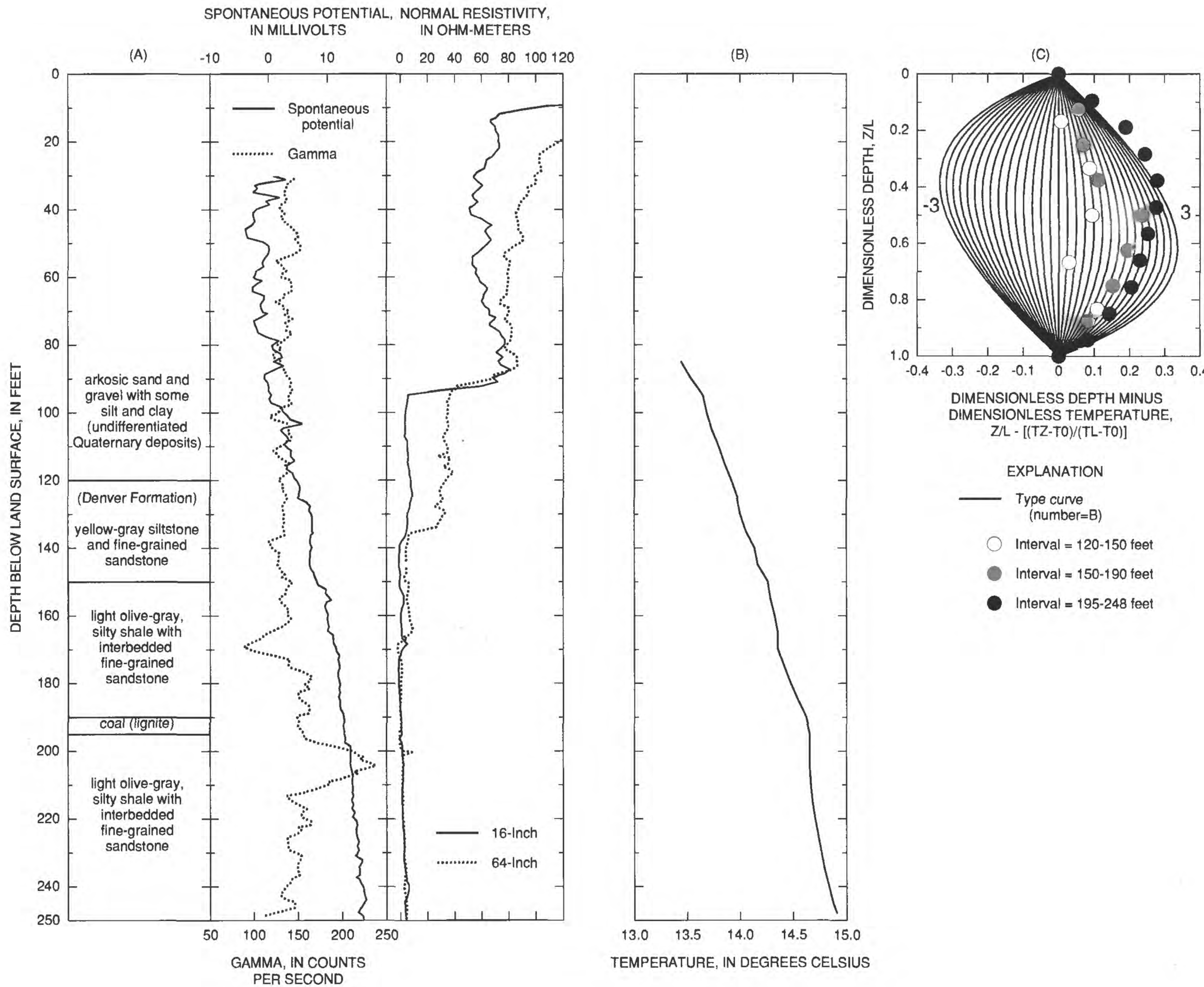


Figure 17. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01306207BCB4 in the upper Black Squirrel Creek Basin.

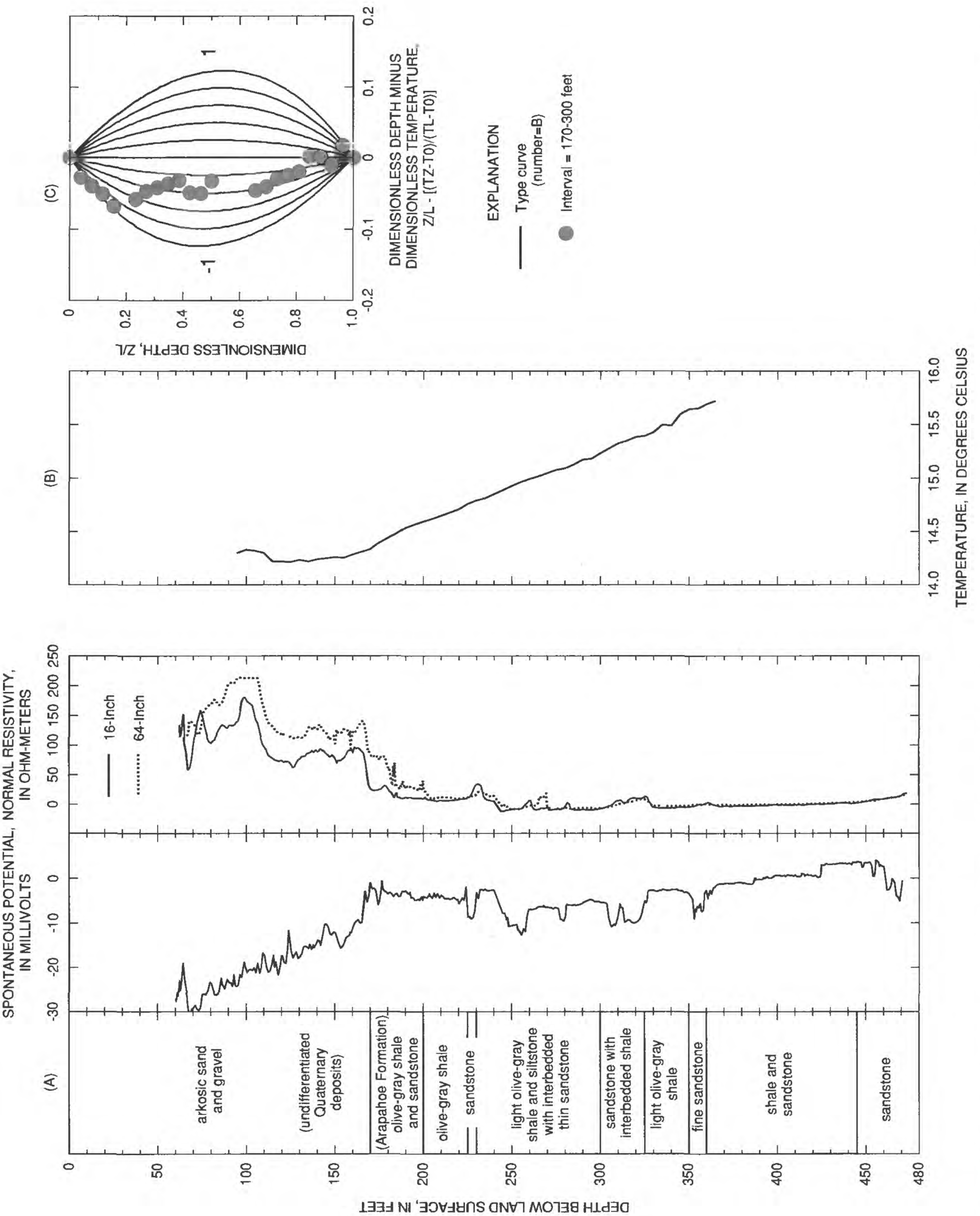


Figure 18. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01306230ACC4 in the upper Black Squirrel Creek Basin.

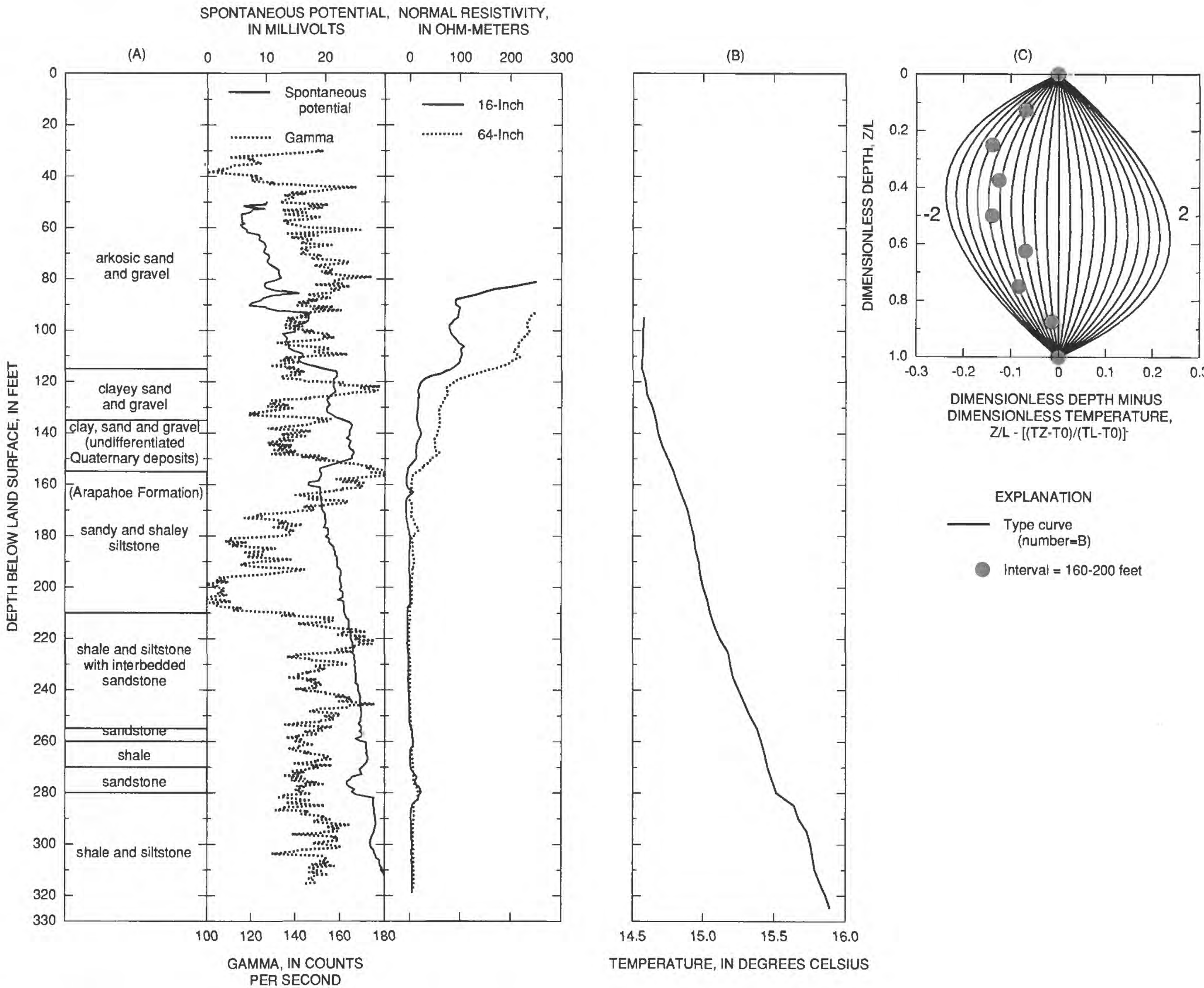


Figure 19. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01406205CAA5 in the upper Black Squirrel Creek Basin.

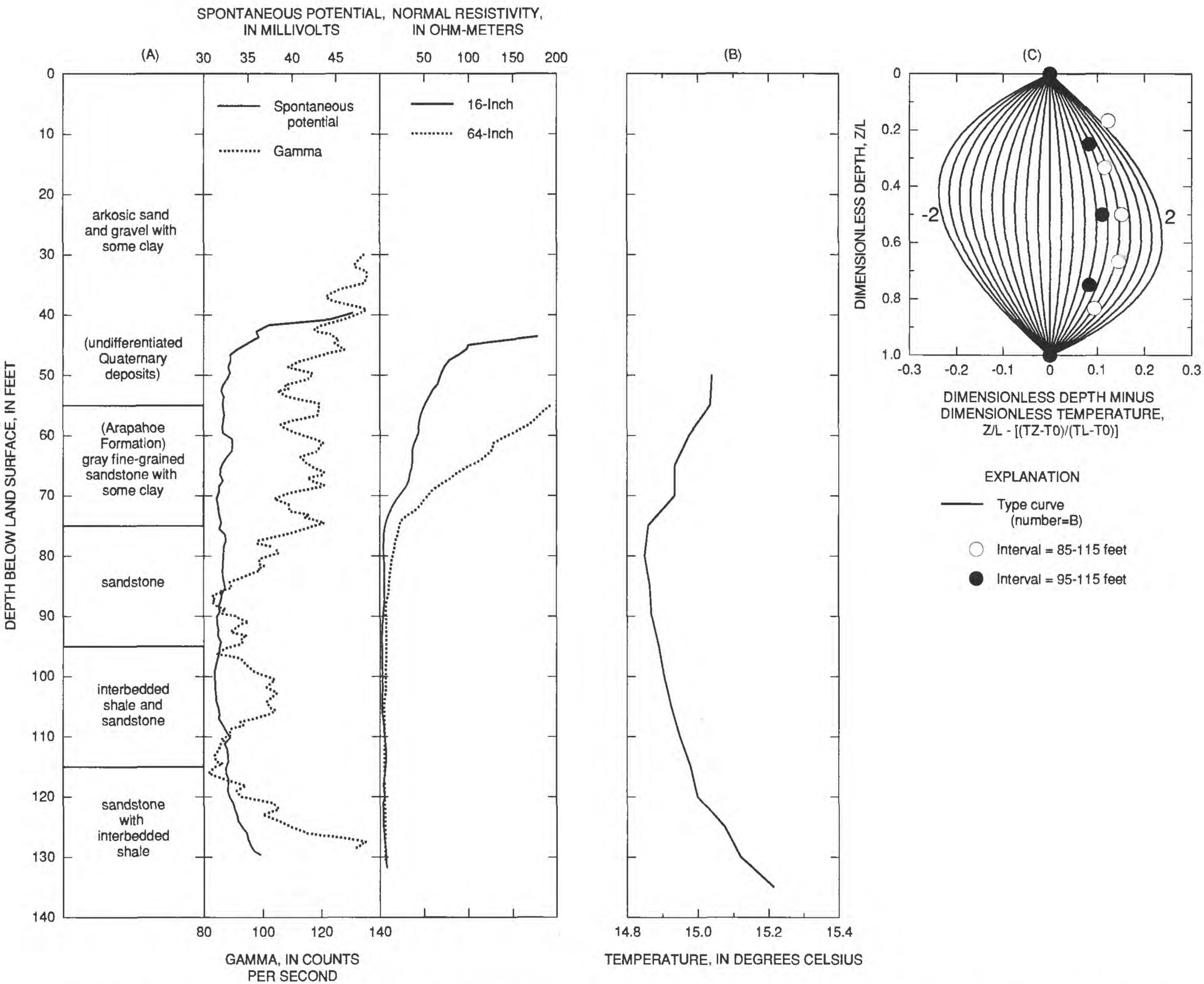


Figure 20. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01406216CDB2 in the upper Black Squirrel Creek Basin.

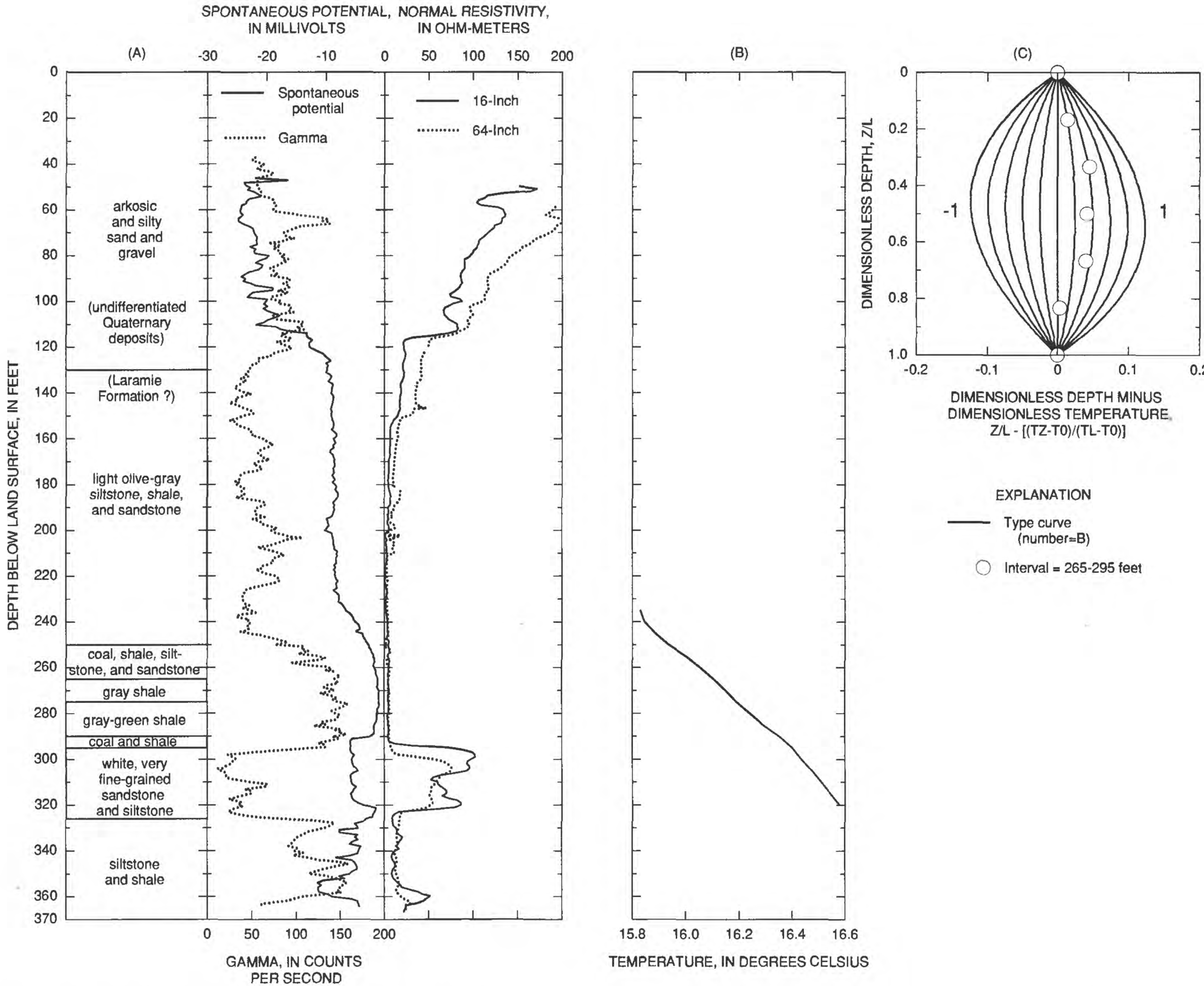


Figure 21. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01406229BBB3 in the upper Black Squirrel Creek Basin.

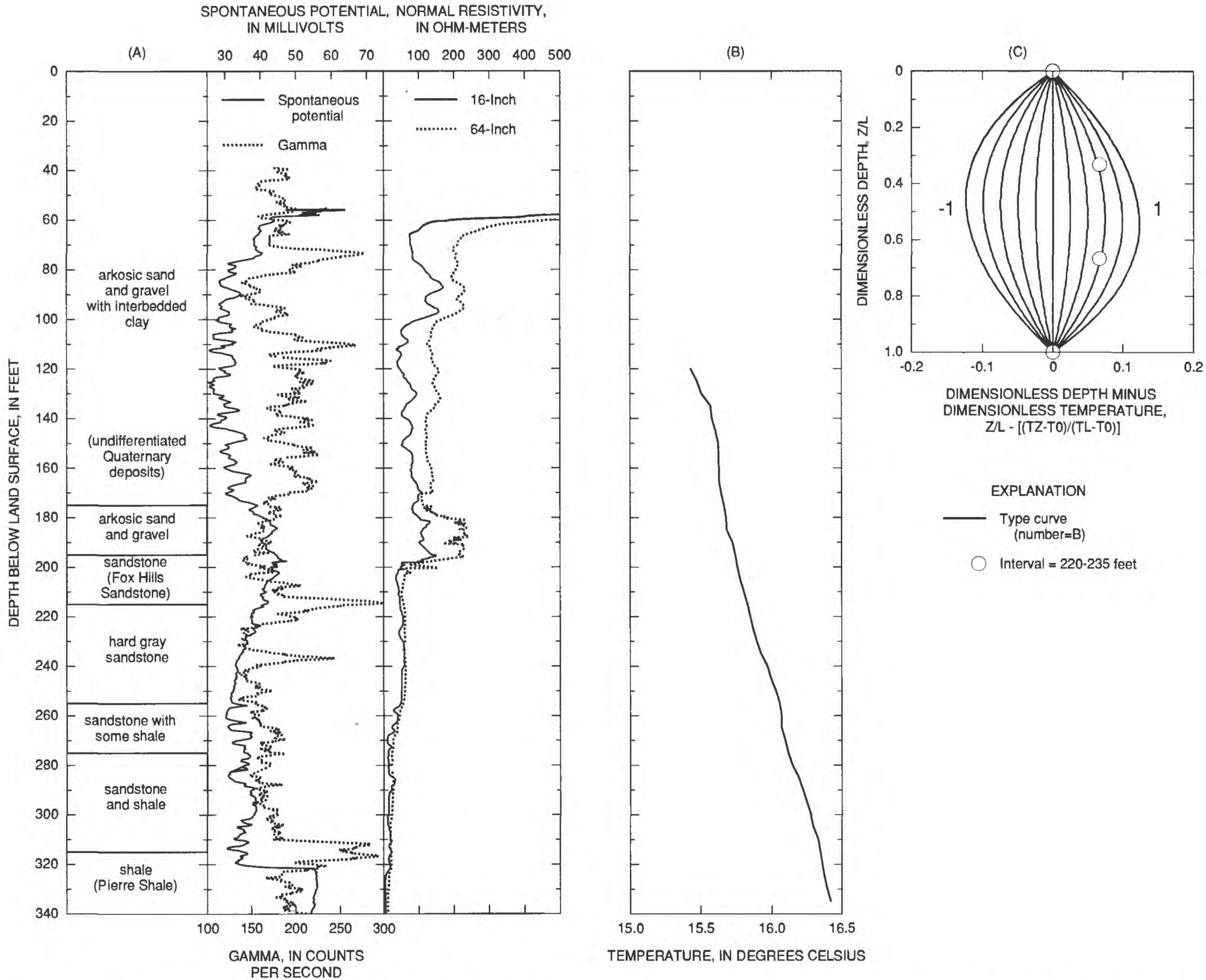


Figure 22. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01506313BBB3 in the upper Black Squirrel Creek Basin.

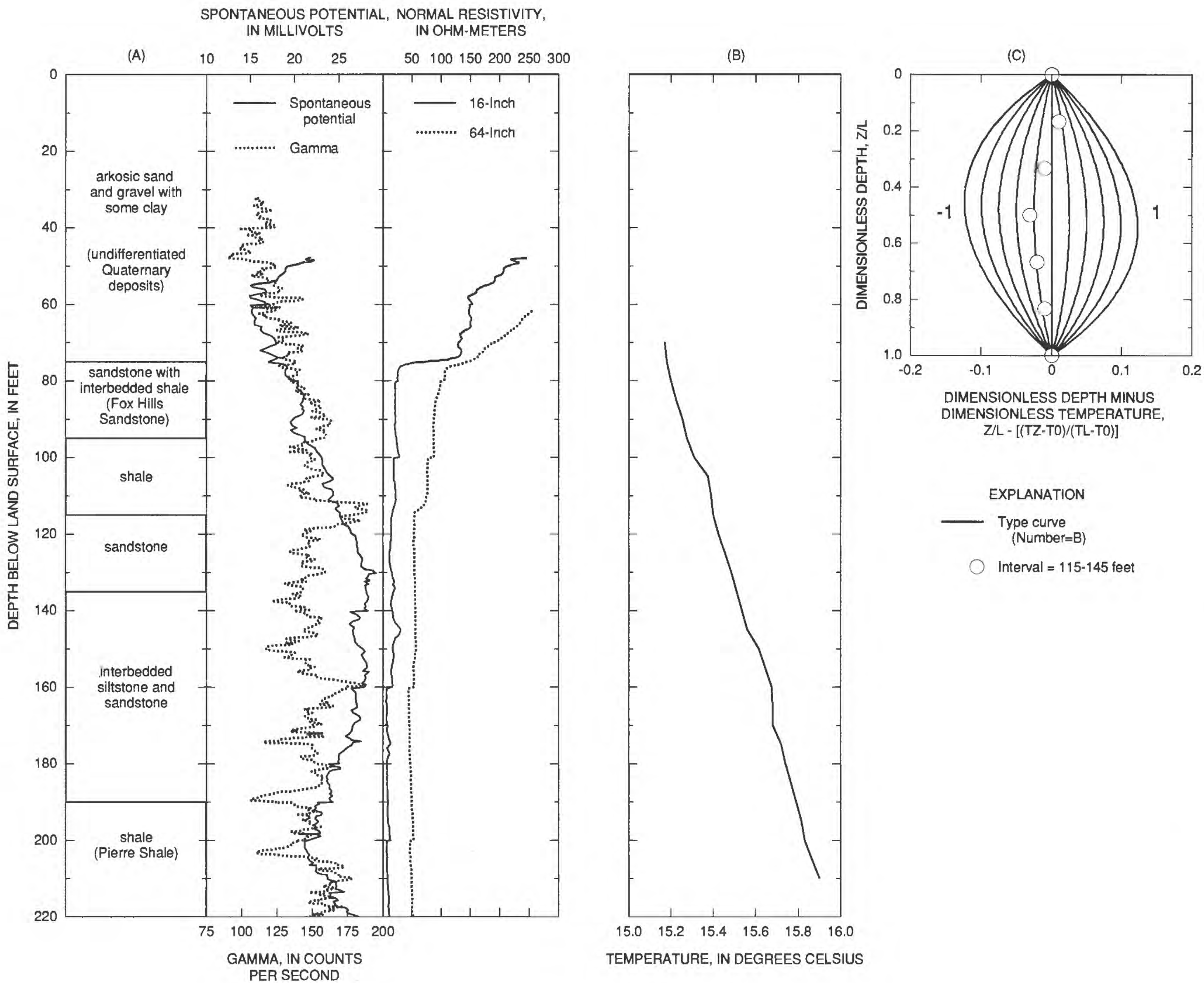


Figure 23. (A) Driller's and selected geophysical logs, (B) subsurface-temperature profile, and (C) relation of dimensionless depth minus dimensionless temperature to dimensionless depth at well SC01506324DBA2 in the upper Black Squirrel Creek Basin.

Table 3. Estimated specific discharge through leaky confining units in the upper Black Squirrel Creek Basin

[c_o, specific heat of water; ρ_o, density of water; k_T, thermal conductivity of saturated porous medium]

Local number	Hydrogeologic unit	Interval number	Depth to top, z_o (feet)	Depth to bottom, z_L (feet)	Approximate thickness, L		Estimated thermal conductivity, k_T (calorie per centimeter per second per degree Celsius)	β[1]	Estimated specific discharge,[2] q_z	
					(feet)	(centimeters)			(centimeters per second)	(feet per day)
SC01206219CCC2	Denver aquifer	1	120	145	25	760	4.2×10^{-3}	−0.6	-3×10^{-6}	−0.01
		2	150	225	75	2,300	4.2×10^{-3}	1.6	3×10^{-6}	.01
SC01306207BCB4	Lower Denver confining unit	1	120	150	30	910	5.2×10^{-3}	0.5	3×10^{-6}	.01
	Lower Denver confining unit	2	150	190	40	1,200	3.2×10^{-3}	1.6	4×10^{-6}	.01
	Lower Denver confining unit	3	195	248	53	1,600	3.2×10^{-3}	2.6	5×10^{-6}	.01
SC01306230ACC4	Arapahoe aquifer	1	170	300	130	4,000	4.2×10^{-3}	−0.4	-4×10^{-7}	−.001
SC01406205CAA5	Arapahoe aquifer	1	160	200	40	1,200	5.2×10^{-3}	−1.2	-6×10^{-6}	−.02
SC01406216CDB2	Arapahoe aquifer	1	85	115	30	910	5.2×10^{-3}	1.2	7×10^{-6}	.02
		2	95	115	20	610	5.2×10^{-3}	.9	8×10^{-6}	.02
SC01406229BBB3	Laramie confining unit	1	265	295	30	910	3.2×10^{-3}	.4	1×10^{-6}	.004
SC01506313BBB3	Laramie-Fox Hills aquifer	1	220	235	15	460	5.2×10^{-3}	.6	7×10^{-6}	.02
SC01506324DBA2	Laramie-Fox Hills aquifer	1	115	145	30	910	5.2×10^{-3}	−0.2	-1×10^{-6}	.003

[1] β is the dimensionless number determined from a match of the field data with a family of theoretical type curves (Bredehoeft and Papadopulos, 1965). If the absolute value of β is less than about 0.2, then the specific discharge is between an upper and lower limit and might be zero.

[2] $q_z = \beta k_T/(c_o \rho_o L)$, where $k_T = 3.2 \times 10^{-3}$ to 5.2×10^{-3} calorie per centimeter per second per degree Celsius (cal/cm/s/°C); $c_o = 1$ calorie per gram per degree Celsius (cal/g/°C); and $\rho_o = 1$ gram per cubic centimeter (g/cm³). Negative values of q_z indicate upward flow, and positive values indicate downward flow.

Analysis of subsurface-temperature profiles to estimate specific discharge through leaky confining units was not found to be appropriate in the study area because: (1) Cyclic and seasonal pumping from the alluvial and the Laramie-Fox Hills aquifers cause large cyclic or seasonal fluctuations in water levels and; therefore, flow is not steady; (2) confining units generally are heterogeneous; and (3) horizontal flow in sandstones within the confining units affects the curvature of the temperature profiles.

Estimating Vertical Hydraulic Properties of Confining Units Using an Aquifer Test

The ratio method, an analytical method developed by Neuman and Witherspoon (1972), was used to determine the vertical hydraulic diffusivity of a 15-ft-thick leaky interval in the lower Denver confining unit at well SC01306207BCB4. The ratio method is based on analysis of the ratio of drawdown in a confining unit to drawdown due to pumping in an adjacent aquifer. Ideally, drawdowns in the confining unit and aquifer are measured simultaneously and in wells at equal radial distances from the control (pumped) well. Ideally, the control well is pumped at a constant rate.

The aquifer test was done during September 1989 and used well SC01306207BCB, Cherokee Metropolitan District well number 6 (CMD–6), as the control well. The well was pumped for 3 days, during which water levels were measured in four observation wells that are completed in the alluvial aquifer (alluvial wells) and one bedrock observation well SC01306207BCB4 that is completed in siltstone and sandstone in the lower Denver confining unit. Locations of the observation wells, relative to the control well, are shown in figure 24. Construction details for the control well and observation wells are listed in table 4.

The alluvial aquifer at the test site consists of about 120 ft of unconsolidated alluvial deposits of Quaternary age (clayey and silty sand and gravel). Static water levels in the alluvial aquifer were about 60 ft below land surface at the beginning of the test; the initial saturated thickness of the alluvial aquifer was about 60 ft. The lower part of the lower Denver confining unit, as interpreted from geophysical logs (fig. 17A), underlies the alluvial aquifer and consists of about 35 ft of siltstone and sandstone, which overlies about 30 ft of silty shale, about 5 ft of coal (lignite), and about 50 ft of silty shale containing interbedded fine-grained sandstone. Depth to the base of the lower Denver confining unit at the test site is estimated to be about 250 ft. The bedrock well, well SC01306207BCB4, is screened in siltstone and sandstone at 135 to 140 ft below land surface. Thickness of the interval between the base of the alluvial aquifer and the open interval of the bedrock well is about 15 ft.

The withdrawal phase of the aquifer test lasted about 3 days, beginning about 8:10 on September 8, 1989. Discharge from well SC01306207BCB was conveyed away from the test site into the distribution system of the Cherokee Metropolitan District. The cumulative volume and instantaneous rate of discharge during the test were measured with an inline flowmeter. The instantaneous rate of discharge varied during the test from about 110,000 ft^3/d (about 570 gal/min) to about 67,000 ft^3/d (about 350 gal/min) (fig. 25). The discharge rate at the start of the test of 110,000 ft^3/d was too large to be maintained for the duration of the test and was decreased to about 77,000 ft^3/d (about 400 gal/min) about 36 minutes after pumping began. During the test, the discharge rate decreased gradually as a result of the reduction in saturated thickness near the control well.

Water levels were measured in the control well and observation wells using calibrated electrical water-level sensing tapes; measurements were recorded to the nearest 0.01 ft. Water levels in the bedrock observation well, SC01306207BCB4, also were recorded on a strip-chart recorder actuated by a float.

The hydraulic properties of the alluvial aquifer were determined using an analytical model for analysis of drawdown in an anisotropic, unconfined aquifer with delayed gravity effects and partially penetrating wells (Neuman, 1975). The vertical hydraulic diffusivity of the 15-ft-thick siltstone and sandstone was determined using the ratio method (Neuman and Witherspoon, 1972). Because the ratio method assumes that the hydraulic diffusivity (horizontal) of the pumped aquifer is known, it was necessary to first determine the hydraulic properties of the alluvial aquifer. The ratio method also requires the simultaneous measurement of drawdowns at the same radial distance from the control well in the confining unit and the pumped aquifer. Because the observation wells in the alluvial aquifer were not at the same radial distance from the control well as the bedrock well, it was necessary to predict drawdown in the alluvial aquifer at the bedrock well. In addition, the alluvial observation wells lie on a line which is approximately perpendicular to a line from the control well to the bedrock observation well, thus, it also was necessary to assume horizontal *isotropy* in the alluvial aquifer.

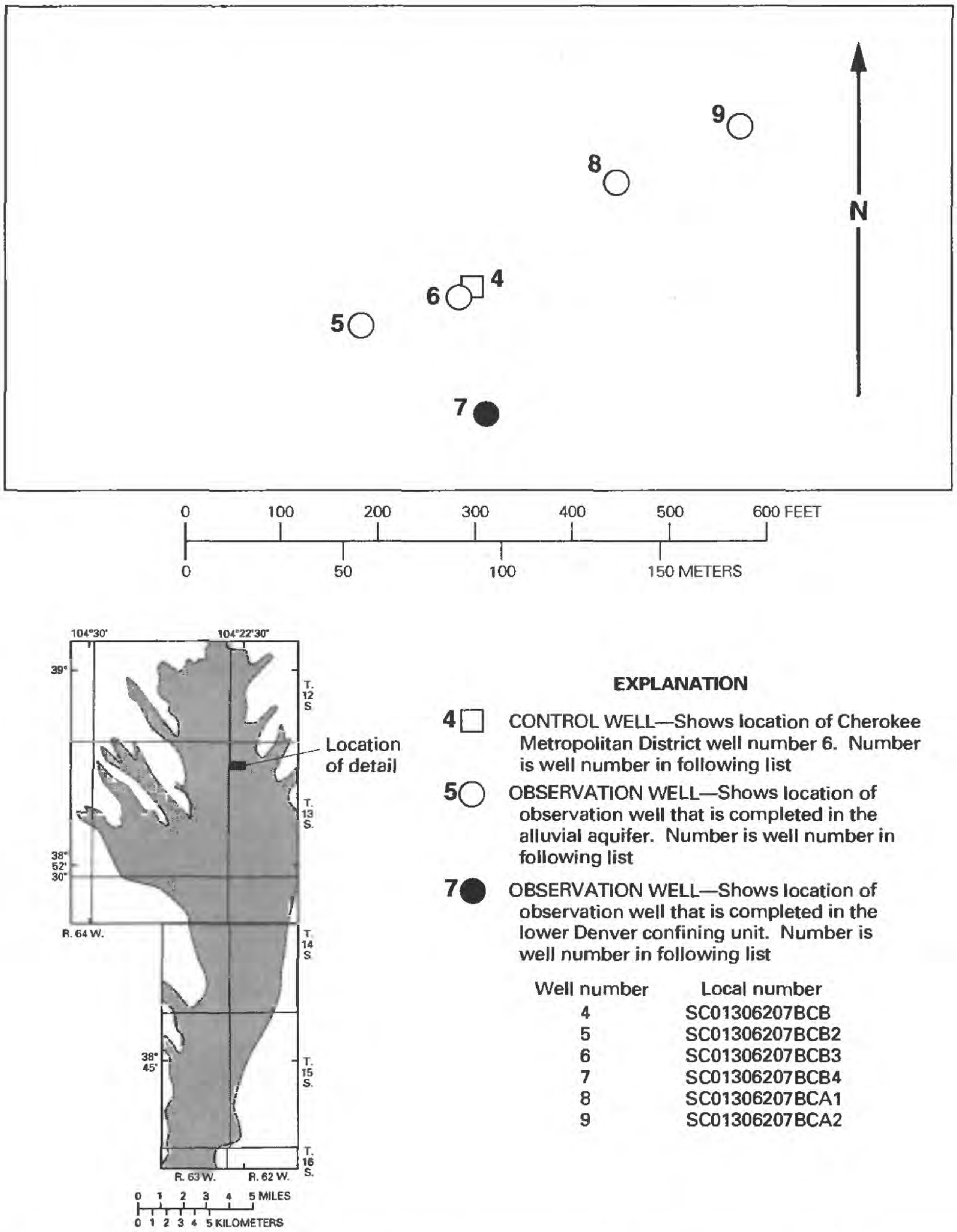


Figure 24. Locations of selected observation wells near well SC01306207BCB, Cherokee Metropolitan District well number 6, in the upper Black Squirrel Creek Basin.

Table 4. Location and construction features and hydrogeologic data for well SC01306207BCB, Cherokee Metropolitan District well number 6, and for selected nearby observation wells in the upper Black Squirrel Creek Basin

Local well number[1]	Well name	Site identification number[2]	Land surface datum (feet)	Well depth (feet)[3]	Depth to				Hydrogeologic unit[4]	Distance from control well (feet)
					Top of screened interval (feet)[3]	Bottom of screened interval (feet)[3]	Top of hydrogeologic unit (feet)	Bottom of hydrogeologic unit (feet)[3]		
SC01306207BCB	CMD 6	385604104230202	6,281.9	130.5	90	130.5	0	126	Alluvial aquifer	0
SC01306207BCA1	CMD 6–1	385606104225601	6,282.5	109R	--	--	0	109	Alluvial aquifer	326
SC01306207BCA2	CMD 6–2	385605104225901	6,283.1	124R	--	--	0	124	Alluvial aquifer	181
SC01306207BCB1	CMD 6–3	385603104230201	6,281.9	120R	--	--	0	120	Alluvial aquifer	17.5
SC01306207BCB2	CMD 6–4	385603104230501	6,283.2	119R	--	--	0	119	Alluvial aquifer	124
SC01306207BCB4	††[5]	385607104230801	6,282.5	246	135	140	120	--	Lower Denver confining unit	133

[1] System of numbering wells is explained at the back of the report.
[2] Site identification number is the U.S. Geological Survey site identification.
[3] R indicates reported depth; -- indicates depth not reported.
[4] Hydrogeologic unit in which well is screened.
[5] ††, no well name identified.

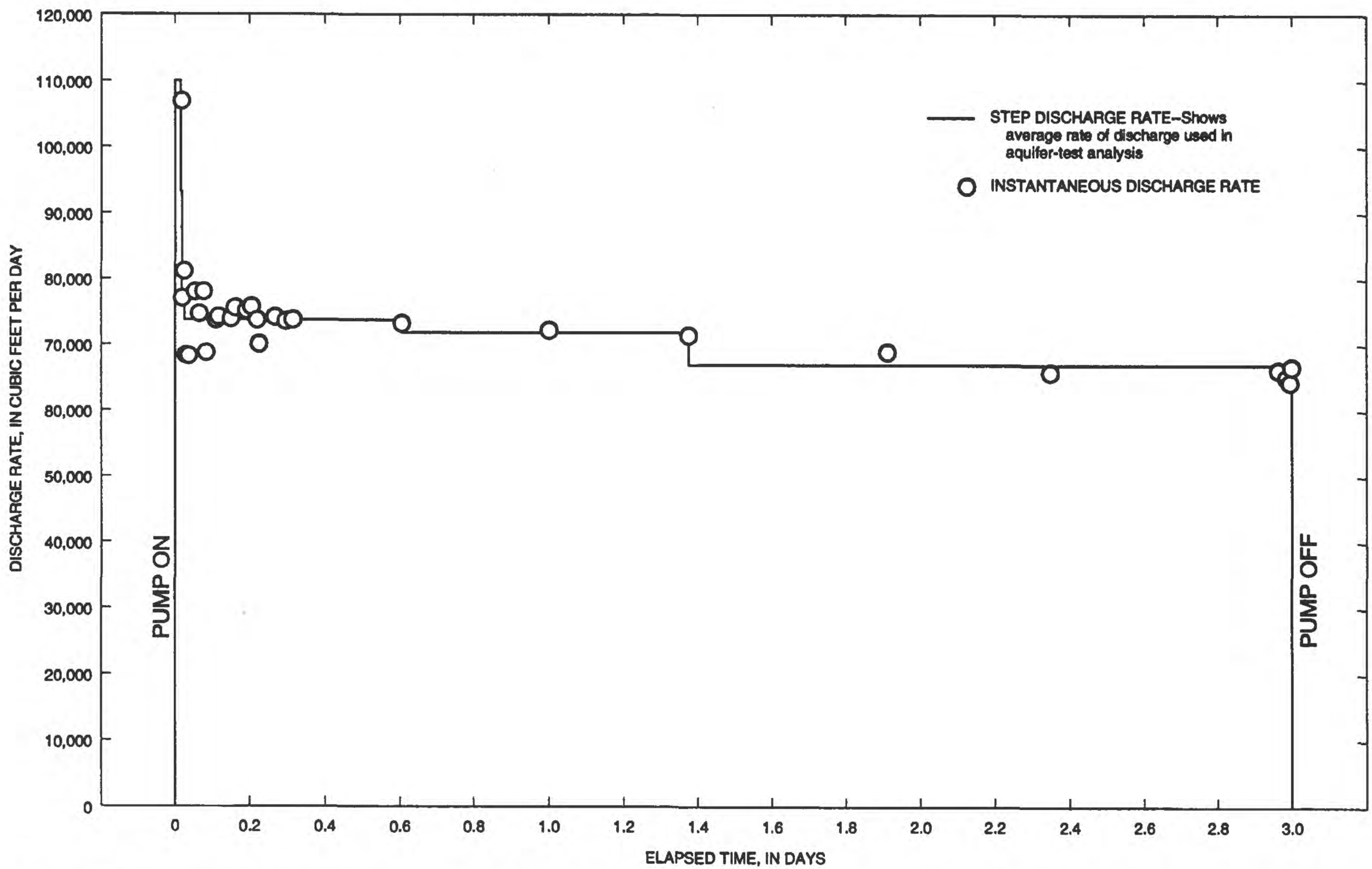


Figure 25. Instantaneous discharge rate of well SC01306207BCB, Cherokee Metropolitan District well number 6, in the upper Black Squirrel Creek Basin during September 8–10, 1989, aquifer test.

Type curves were developed for the four alluvial observation wells based on assumptions of: (1) Unconfined and vertically anisotropic conditions, (2) delayed-gravity response, (3) fully penetrating observation wells, and (4) a partially penetrating control well. Construction features of the alluvial observation wells, other than depth, were not known and it was assumed that they were fully penetrating. A computer program, originally developed by Neuman (1975), was modified to account for variable discharge conditions, using the principle of superposition (Reilly and others, 1987), and was used to generate theoretical type curves for alluvial observation wells.

Although the principle of superposition is strictly valid only for linear systems and the analytical equation for drawdown in an unconfined aquifer is nonlinear, the principle of superposition is approximately correct when drawdown is small compared to the initial saturated thickness. Because drawdown is a function of the square of the radial distance from the control well, the effects of nonlinearity decrease exponentially as radial distance increases. For practical purposes, the analytical equation can be considered to be linear for early time, soon after pumping starts, at relatively large radial distances from the control well.

An iterative approach was used in determining the hydraulic properties for a theoretical type curve that would match measured drawdown in the alluvial observation wells. Initial values of the horizontal and vertical hydraulic conductivity and of the early-time storage coefficient of the alluvial aquifer, which were based on analysis of early-time drawdown and estimated specific yield of the alluvial aquifer, were adjusted until the theoretical type curve (fig. 26, lines) was developed for each observation well that approximately fit the time-drawdown data (fig. 26, symbols) for the well.

The hydraulic properties of the alluvial aquifer, determined from the aquifer test, were: Horizontal hydraulic conductivity, $K_{x,y} = 48$ ft/d; vertical hydraulic conductivity, $K_z = 3$ ft/d; early-time storage coefficient, $S_e = 0.003$; and an estimated specific yield, $S_y = 0.15$ (table 2). *Transmissivity* (T) of the alluvial aquifer, assuming a saturated thickness of 60 ft, is about 2,900 ft^2/d near well SC01306207BCB, and specific storage of the alluvial aquifer (S_s) is about 5×10^{-5} per ft. Hydraulic diffusivity of the alluvial aquifer, $K_{x,y}/S_s$, is about 9.6×10^5 ft^2/d during early time.

The hydraulic properties of the alluvial aquifer (table 2) were used with the model of Neuman (1975) to predict drawdown in the pumped aquifer for a hypothetical well that is completed near the base of the alluvial aquifer at a radius of 133 ft from the control well (fig. 26). Because the alluvial aquifer is anisotropic (the ratio of horizontal to vertical hydraulic conductivity is 16:1 [48:3, table 2]), drawdown was computed for a hypothetical partially penetrating well that was perforated near the base of the alluvial aquifer. Drawdown near the base of the alluvial aquifer would be larger than drawdown in a fully penetrating observation well.

Drawdown (s′) measured in the bedrock observation well was corrected for long-term water-level trend (recovery) and for barometric effects and is shown in figure 26. The value of drawdown (s) in the alluvial aquifer was determined for selected values of time (t) from the steep part of the theoretical time-drawdown curve near well SC01306207BCB4 (fig. 26). The ratio s′/s for selected values of s′, measured near the start of the test, are listed in table 5 and are shown superimposed on the family of type curves in figure 27 (modified from Neuman and Witherspoon, 1972). Early-time data are preferred for use in the ratio method because the type curves are steepest and, therefore, more definitive during early time (Neuman and Witherspoon, 1972).

A value of dimensionless time (t_D) is calculated for each value of s′/s, as:

$$t_D = \frac{K_{x,y}t}{S_s r^2} or \frac{Tt}{S_e r^2} \qquad (4)$$

where

$K_{x,y}$ = horizontal hydraulic conductivity of the pumped aquifer, $K_{x,y} = 48$ ft/d;

t = elapsed time since pumping began;

S_s = specific storage of the pumped aquifer, $S_s = 5 \times 10^{-5}$ per ft;

r = radial distance from control well to point of observation, r = 133 ft;

T = transmissivity of the pumped aquifer, T = 2,900 ft^2/d; and

S_e = early-time storage coefficient of the pumped aquifer, $S_e = 0.003$.

For example, when t = 0.04028 d and at r = 133 ft, t_D = (48 ft/d × 0.04028 d)/(5 × 10^{-5} per ft × (133 ft)^2 = 2.19, s' = 0.192 ft, and s = 3.89 ft; at t_D = 2.19 and s′/s = 0.049, and t'_D = 0.18 (fig. 27). The dimensionless time, t'_D, for the confining interval is:

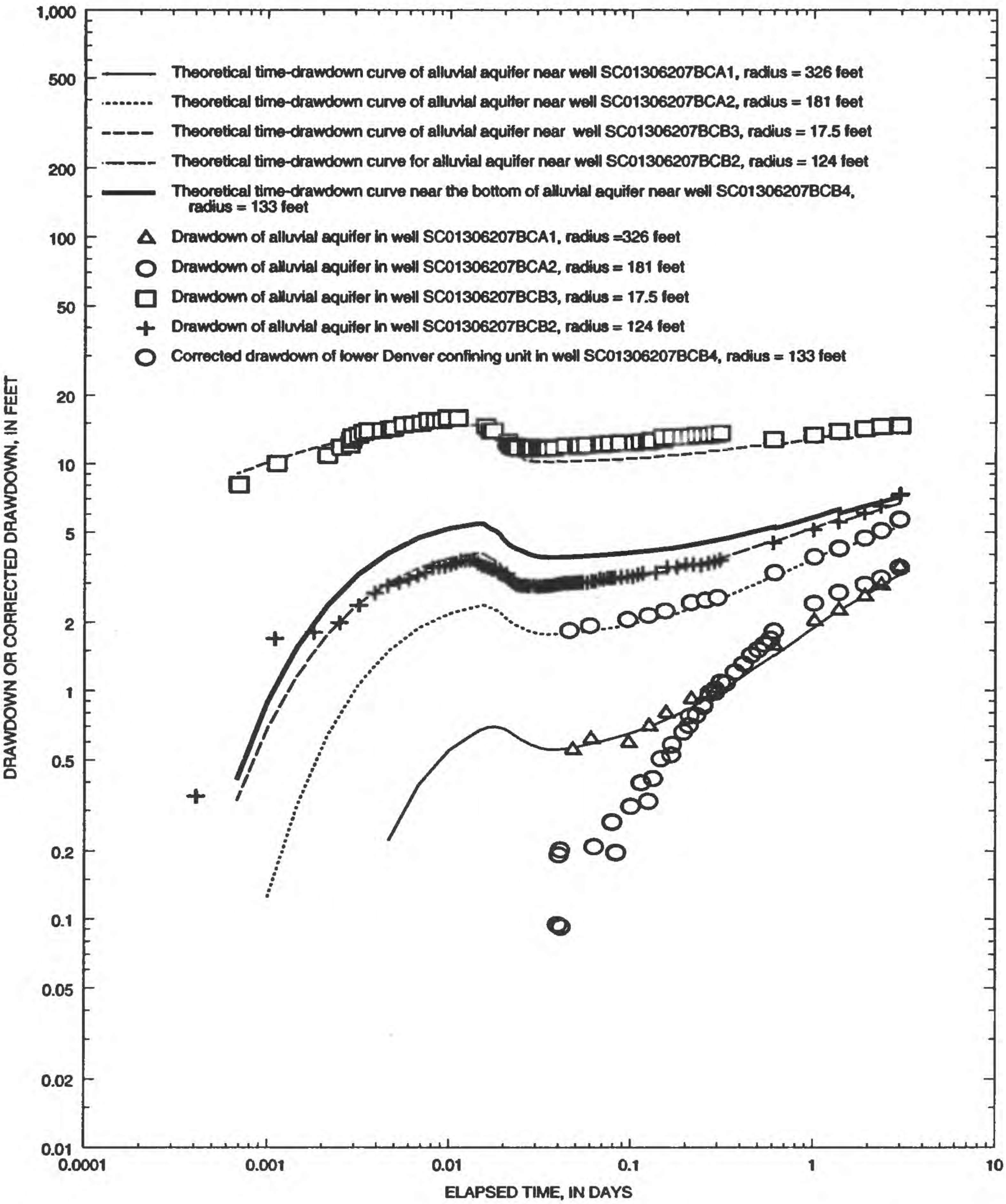


Figure 26. Relation of theoretical time-drawdown curves and measured drawdown to time in the four alluvial observation wells, theoretical drawdown in a hypothetical alluvial well at a radial distance of 133 feet, and relation of corrected drawdown to time in the lower Denver confining unit in well SC01306207BCB4 in the upper Black Squirrel Creek Basin.

Table 5. Vertical hydraulic diffusivity of a 15-foot-thick siltstone and sandstone in the Denver aquifer that underlies the alluvial aquifer at well SC01306207BCB4 in the upper Black Squirrel Creek Basin

Elapsed time since pumping began, t (day)	Measured drawdown in confining unit, s' (foot)	[1]Hypothetical drawdown near the base of alluvial aquifer, s (feet)	Drawdown ratio, s'/s	[2]Dimensionless time for the pumped aquifer, t_D	[3]Dimensionless time for the confining interval, t'_D	[4]Vertical hydraulic diffusivity of confining interval, K_z'/S_s' (square foot per day)
0.03958	0.094	3.89	0.024	2.15	0.15	8.5×10^2
.04028	.192	3.89	.049	2.19	.18	1.0×10^3
.04097	.202	3.89	.052	2.22	.19	1.0×10^3

[1]Hypothetical drawdown near the base of the alluvial aquifer was computed using analytical model of Neuman (1975).

[2]Dimensionless time, $t_D = K_{x,y}t/S_s r^2$, where K = 48 feet per day, $S_s = 5 \times 10^{-5}$ per foot, t = elapsed time in days, and r = 133 feet; or $t_D = Tt/S_e r^2$, where T = 2,900 square feet per day and $S_e = 0.003$.

[3]Dimensionless time, t'_D, determined graphically from type curve (Neuman and Witherspoon, 1972).

[4]Vertical hydraulic diffusivity of the confining unit, $K_z'/S_s' = t'_D z^2/t$, where K_z' = the vertical hydraulic conductivity and S_s' = the specific storage of the leaky confining interval; t'_D = the dimensionless time; z = thickness of the confining interval from the contact with the pumped aquifer to the top of the screened interval (z = 15 feet); and t = the elapsed time since pumping began.

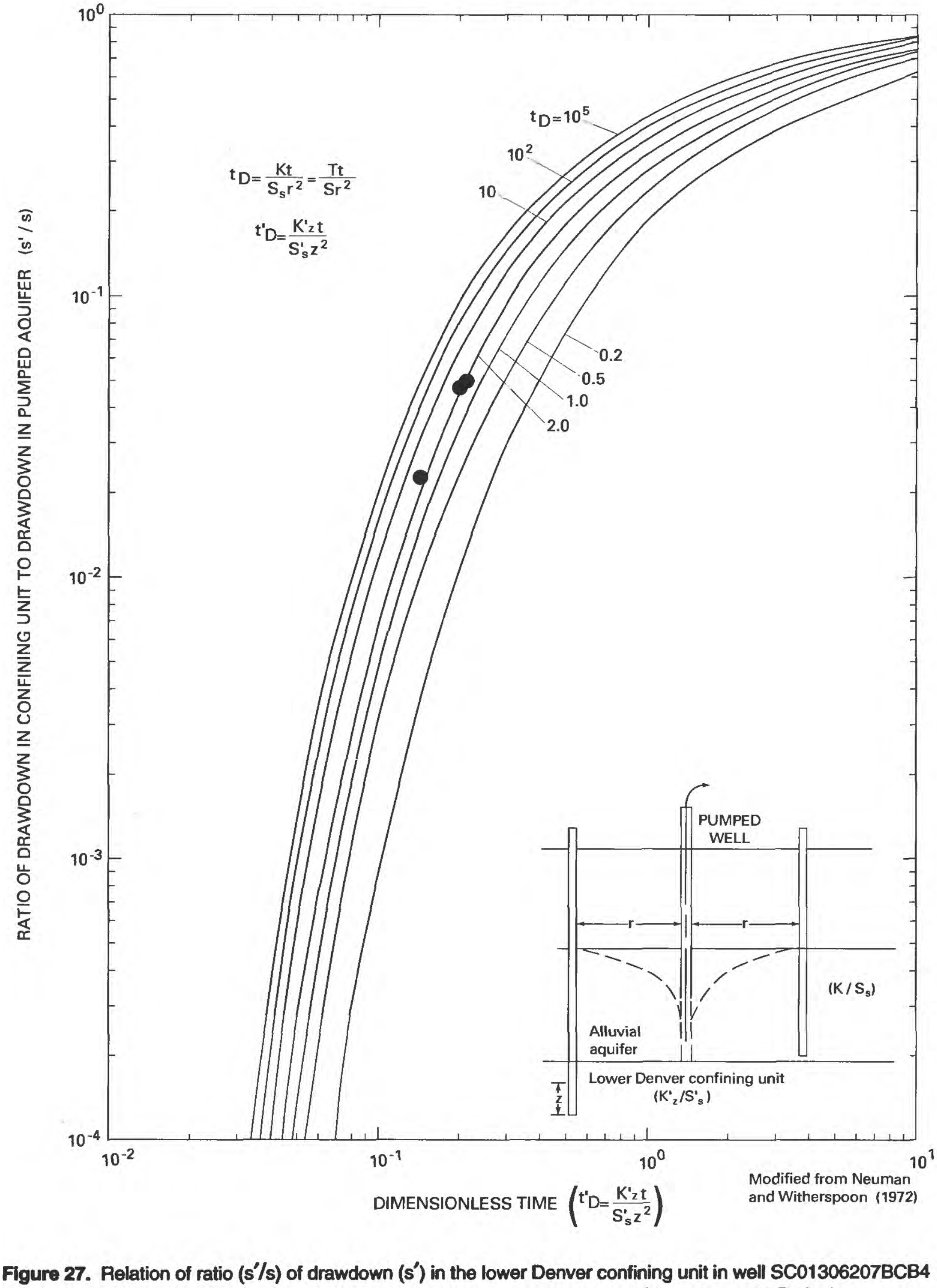


Figure 27. Relation of ratio (s′/s) of drawdown (s′) in the lower Denver confining unit in well SC01306207BCB4 with hypothetical drawdown (s) near the base of the alluvial aquifer near well SC01306207BCB4 in the upper Black Squirrel Creek Basin to dimensionless time (t'_D), superimposed on the family of type curves of dimensionless time (t_D).

$$t'_D = \frac{K_z't}{S_s'z^2} \qquad (5)$$

where

K'_z = vertical hydraulic conductivity of confining interval;

z = thickness of confining interval; and

S_s' = specific storage of the confining interval.

8×10^{-4} ft/d (1,000 ft^2/d $\times$ 8×10^{-7} per ft = 8×10^{-4} ft/d).

Rearranging eq. 5, the vertical hydraulic diffusivity of the confining unit (K_z'/S_s') is computed as:

$$K_z'/S_s' = t'_D z^2/t \qquad (6)$$

For example, at t = 0.04028 d, t'_D = 0.18, and z = 15 ft, K_z'/S'_s is approximately 1,000 ft^2/d (0.18 $\times$ 225 ft^2 0.04028d = 1,005 ft^2/d).

Specific storage of the confining unit (S'_s) was estimated independently, from the barometric efficiency of the well (0.53) and estimated porosity (0.3) of the perforated interval, to be about 8×10^{-7} per ft. The estimate of porosity of 0.3 was based on porosity of similar rocks (Robson, 1987, table 1). The vertical hydraulic conductivity of the confining interval (K_z') at well SC01306207BCB4 was estimated as the product of vertical hydraulic diffusivity and specific storage at 8×10^{-4} ft/d (1,000 ft^2/d $\times$ 8×10^{-7} per ft = 8×10^{-4} ft/d).

A vertical hydraulic conductivity (K_z') for the lower Denver confining unit of 8×10^{-4} ft/d is about 60 times larger than estimated $K_z' = 1.3 \times 10^{-5}$ ft/d for the lower Denver confining unit (Banta, 1989). The value of K_z' for the lower Denver confining unit was determined by the ratio method for a 15-ft-thick interval, which consists of siltstone and sandstone, whereas Banta's estimate is for the 200- to 300-ft-thick lower Denver confining unit, which is mainly shale. Based on the lithology of the confining unit, the values of K_z' and S_s' determined at well SC01306207BCB4 likely are representative of values for bedrock aquifers, not the bedrock confining units. The aquifer test results indicate that the ratio method can be used to estimate the vertical hydraulic diffusivity of leaky confining units in the study area.

Water Quality—An Indicator of Flow of Water Between the Alluvial and Bedrock Aquifers

Selected physical and chemical characteristics of water from the alluvial aquifer generally are quite different from those of water from the underlying bedrock aquifers. Where similarities in water quality in the alluvial and bedrock aquifers occur, the similarities may indicate the direction of flow between the alluvial aquifer and underlying bedrock aquifers. Values of selected physical properties and concentrations of selected dissolved constituents in water from seven alluvial wells and from the eight bedrock wells are listed in table 6 (locations of wells listed in table 6 are shown in figure 5).

Water from the alluvial aquifer in the study area generally is classified as either a calcium sodium-mixed anion or a calcium sodium bicarbonate type of water (fig. 28). The concentrations of dissolved calcium plus magnesium approximately equal concentrations of dissolved sodium plus potassium in water samples from the alluvial aquifer (fig. 28). The samples from the alluvial wells SC01506218ACB (site 22 in fig. 28) and SC01506310DCC (site 23 in fig. 28), wells which are located in an area in which the alluvial aquifer is hydraulically connected with the Laramie-Fox Hills aquifer, contain a larger proportion of carbonate and bicarbonate (alkalinity) than most samples from the alluvial aquifer. Water from the Denver, Arapahoe, and the Laramie-Fox Hills aquifers in the study area is classified as either sodium bicarbonate or sodium-mixed anion type of water (fig. 28). The relatively large proportion of calcium plus magnesium in water from well SC01506313BBB3 (site 25 in fig. 28), a well completed in the Laramie-Fox Hills aquifer, is similar to the proportion of calcium plus magnesium in water from the alluvial aquifer, indicating flow from the alluvial aquifer to the Laramie-Fox Hills aquifer near well SC01506313BBB3. Water from most of the bedrock wells sampled during this study contain at least 88 percent sodium plus potassium ions; water from well SC01506313BBB3 contained only about 65 percent sodium plus potassium ions.

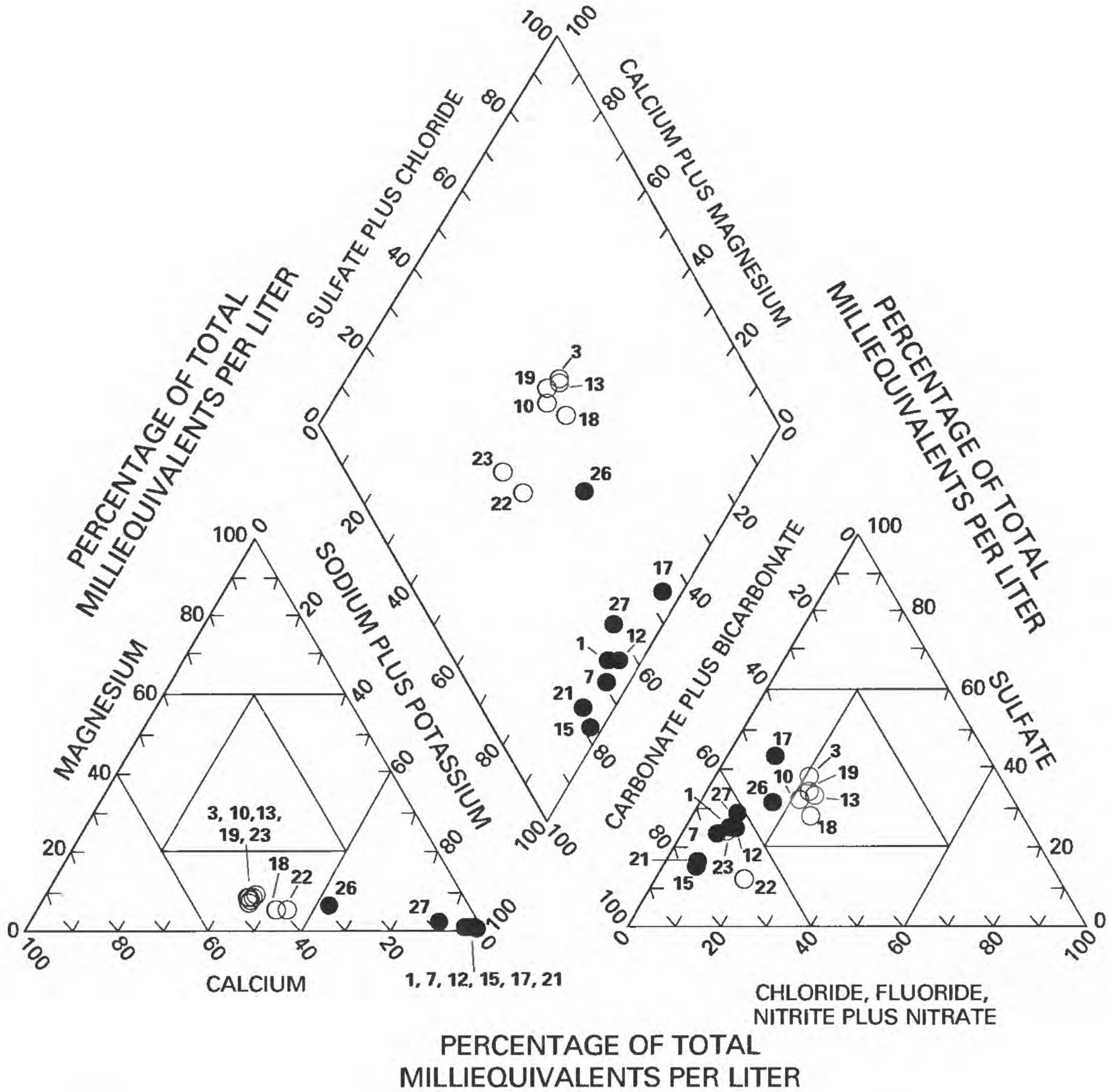
Concentrations of nitrite plus nitrate, as nitrogen, in water from the alluvial aquifer vary considerably, ranging from less than 0.10 to as much as 72 mg/L (Buckles and Watts, 1988, p. 23–25). Concentrations of nitrite plus nitrate, as nitrogen, in ground water that are greater than about 5 to 10 mg/L, generally indicate anthropogenic effects from surface sources of nitrogen-enriched water; for example, septic or animal waste or fertilizer (Hem, 1985, p. 125). Concentrations of nitrite plus nitrate, as nitrogen, in ground water from

Table 6. Selected physical and chemical characteristics of water from the alluvial and bedrock hydrogeologic units in the upper Black Squirrel Creek Basin

[°C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter; <, less than]

Site number in figure 5	Local well number[1]	Hydrogeologic unit	Date sampled	Temperature, water (°C)	pH (standard units)	Specific conductance (µS/cm)	Solids, sum of constituents, dissolved (mg/L)	Calcium, dissolved (mg/L)	Magnesium, dissolved (mg/L)	Potassium, dissolved (mg/L)	Sodium, dissolved (mg/L)	Alkalinity, laboratory (mg/L as $CaCO_3$)	Chloride, dissolved (mg/L)	Fluoride, dissolved (mg/L)	Sulfate, dissolved (mg/L)	Nitrogen, nitrite plus nitrate, dissolved (mg/L)
1	SC01206219CCC2	Upper Denver confining unit	06–29–87	12.5	9.6	885	652	6.3	1.6	1.7	240	323	19	7.6	130	1.2
3	SC01206336ACC	Alluvial aquifer	08–08–84	11.5	6.3	400	259	35	4.1	2.4	39	79	10	.4	65	6.3
7	SC01306207BCB4	Lower Denver confining unit	06–26–87	13.5	9.3	445	337	3.7	.6	1.2	130	178	9.7	1.5	62	.28
10	SC01306230ACC1	Alluvial aquifer	08–07–86	13.0	7.3	358	328	35	<4.4	2.4	40	97	12	.4	60	<6.8
12	SC01306230ACC4	Arapahoe aquifer	06–26–87	14.0	9.3	520	402	5.0	1.3	1.6	140	203	12	3.6	79	.60
13	SC01406205CAA	Alluvial aquifer	08–07–84	13.5	6.7	410	266	36	4.0	2.3	41	82	14	.4	58	7.0
15	SC01406205CAA5	Arapahoe aquifer	06–24–87	14.0	9.3	476	285	1.7	.2	.8	110	197	6.3	4.3	34	<0.10
17	SC01406216CDB2	Arapahoe aquifer	06–23–87	13.5	9.5	554	339	3.9	.3	1.2	120	123	16	2.9	110	<0.10
18	SC01406216CCC	Alluvial aquifer	08–10–84	13.0	7.5	870	546	73	7.2	2.8	100	197	48	.7	140	8.1
19	SC01406220DBC	Alluvial aquifer	08–12–86	13.0	7.3	535	284	47	<5.2	2.6	50	125	24	.3	72	<10
21	SC01406229BBB3	Laramie-Fox Hills aquifer	06–17–87	15.0	9.1	473	290	4.6	.4	2.2	100	205	12	1.4	40	<0.10
22	SC01506218ACB	Alluvial aquifer	08–08–84	13.5	7.1	525	326	44	4.7	2.4	67	176	13	1.0	61	1.2
23	SC01506310DCC	Alluvial aquifer	08–07–84	14.5	7.2	280	200	30	2.8	1.9	32	104	7.0	.4	17	5.5
25	SC01506313BBB3	Laramie-Fox Hills aquifer	07–02–87	13.5	7.9	444	338	32	3.5	2.8	79	131	15	.9	75	4.9
27	SC01506324DBA2	Laramie-Fox Hills aquifer	06–30–87	13.5	8.3	398	265	8.3	1.3	3.0	83	129	7.3	.8	60	2.0

[1]The system of numbering wells is explained in "System of Numbering Wells" at the back of the report.



EXPLANATION

18 ○ ALLUVIAL AQUIFER—Shows cation-anion ratios for a water sample from the alluvial aquifer. Number near circle represents site number in table on facing page

21 ● BEDROCK AQUIFER OR CONFINING UNIT—Shows cation-anion ratios for a water sample from a bedrock aquifer or confining unit. Number near circle represents site number in table on facing page

Figure 28. Relative proportions of dissolved cations and anions in water from the alluvial aquifer and bedrock hydrogeologic units in the upper Black Squirrel Creek Basin.

Site number	Local number	Hydrogeologic unit
1	SC01206219CCC2	Upper Denver confining unit
3	SC01206236ACC	Alluvial aquifer
7	SC01306207BCB4	Lower Denver confining unit
10	SC01306230ACC1	Alluvial aquifer
12	SC01306230ACC4	Arapahoe aquifer
13	SC01406205CAA	Alluvial aquifer
15	SC01406205CAA5	Arapahoe aquifer
18	SC01406216CCC	Alluvial aquifer
17	SC01406216CDB2	Arapahoe aquifer
19	SC01406220DBC	Alluvial aquifer
21	SC01406229BBB3	Laramie-Fox Hills aquifer
22	SC01506218ACB	Alluvial aquifer
23	SC01506310DCC	Alluvial aquifer
25	SC01506313BBB3	Laramie-Fox Hills aquifer
27	SC01506324DBA2	Laramie-Fox Hills aquifer

Figure 28. Relative proportions of dissolved cations and anions in water from the alluvial aquifer and bedrock hydrogeologic units in the upper Black Squirrel Creek Basin--Continued.

the Denver and Arapahoe aquifers in the study area range from less than 0.10 to 1.2 mg/L and in water from the Laramie-Fox Hills aquifer from less than 0.10 to 4.9 mg/L (table 6). The relatively large concentration of nitrite plus nitrate, as nitrogen (4.9 mg/L), in water from the Laramie-Fox Hills aquifer at well SC01506313BBB3, indicates flow from the alluvial aquifer to the Laramie-Fox Hills aquifer because water in the alluvial aquifer contains relatively large concentrations of nitrite plus nitrate, as nitrogen, in this area (Buckles and Watts, 1988, fig. 14).

Similarities between physical and chemical characteristics of water from the Laramie-Fox Hills aquifer in the southern part of the study area, with physical and chemical characteristics of water from the alluvial aquifer, indicate that the Laramie-Fox Hills and alluvial aquifers locally are hydraulically connected. Locally, in upland areas west of Black Squirrel Creek, downward flow of water from the alluvial aquifer to the Laramie-Fox Hills aquifer may occur where the Laramie confining unit does not separate the aquifers.

Lack of similarity between physical and chemical characteristics of water from the Denver and Arapahoe aquifers with characteristics of water from the alluvial aquifer indicates that either the alluvial and bedrock aquifers are not hydraulically connected or that upward flow from the bedrock aquifers to the alluvial aquifer is too small to substantially alter the characteristics of the water in the alluvial aquifer. In general, the physical and chemical characteristics of water from the alluvial aquifer do not indicate upward flow from the bedrock aquifers. Dilution can mask the effects of transport and mixing of dissolved ions in water from the bedrock aquifers on concentrations of dissolved ions in water in the alluvial aquifer.

SIMULATION OF FLOW BETWEEN THE ALLUVIAL AND BEDROCK AQUIFERS IN THE UPPER BLACK SQUIRREL CREEK BASIN

A numerical model of ground-water flow is a mathematical tool used for further understanding of the ground-water system and can be used to evaluate the hydrologic effects that can result from changes in recharge and discharge conditions. A model of steady-state conditions is used to evaluate hydrologic conditions of the natural system prior to the development and use of water resources. A model of steady-state conditions provides estimates of long-term average recharge and discharge and relates the hydraulic characteristics and geometric configuration of the aquifer system and recharge and discharge to pristine water levels. A model of transient-state conditions is used to evaluate the time-dependent response of the natural system to the development and use of water resources or to predict the system's response to future changes in recharge and discharge conditions. Because of the large number of variables involved in modeling ground-water flow and because some of the variables are poorly defined by data, a unique solution of the system of equations cannot be guaranteed. However, a plausible solution to the system of equations generally can be determined if care is taken to match the response of the real system with calculated response of the numerical model.

Conceptual Model

A conceptual model is a qualitative description of the hydrologic system on which the numerical model is based. The saturated unconsolidated deposits (alluvial aquifer) unconformably overlie a sequence of permeable and relatively impermeable rocks in the upper Black Squirrel Creek Basin east of Colorado Springs (fig. 29). The bedrock aquifers (table 1), which include the Dawson, Denver, Arapahoe, and Laramie-Fox Hills aquifers, are separated by the upper and lower Denver and Laramie confining units and overlie the relatively impermeable Pierre Shale confining unit (table 1).

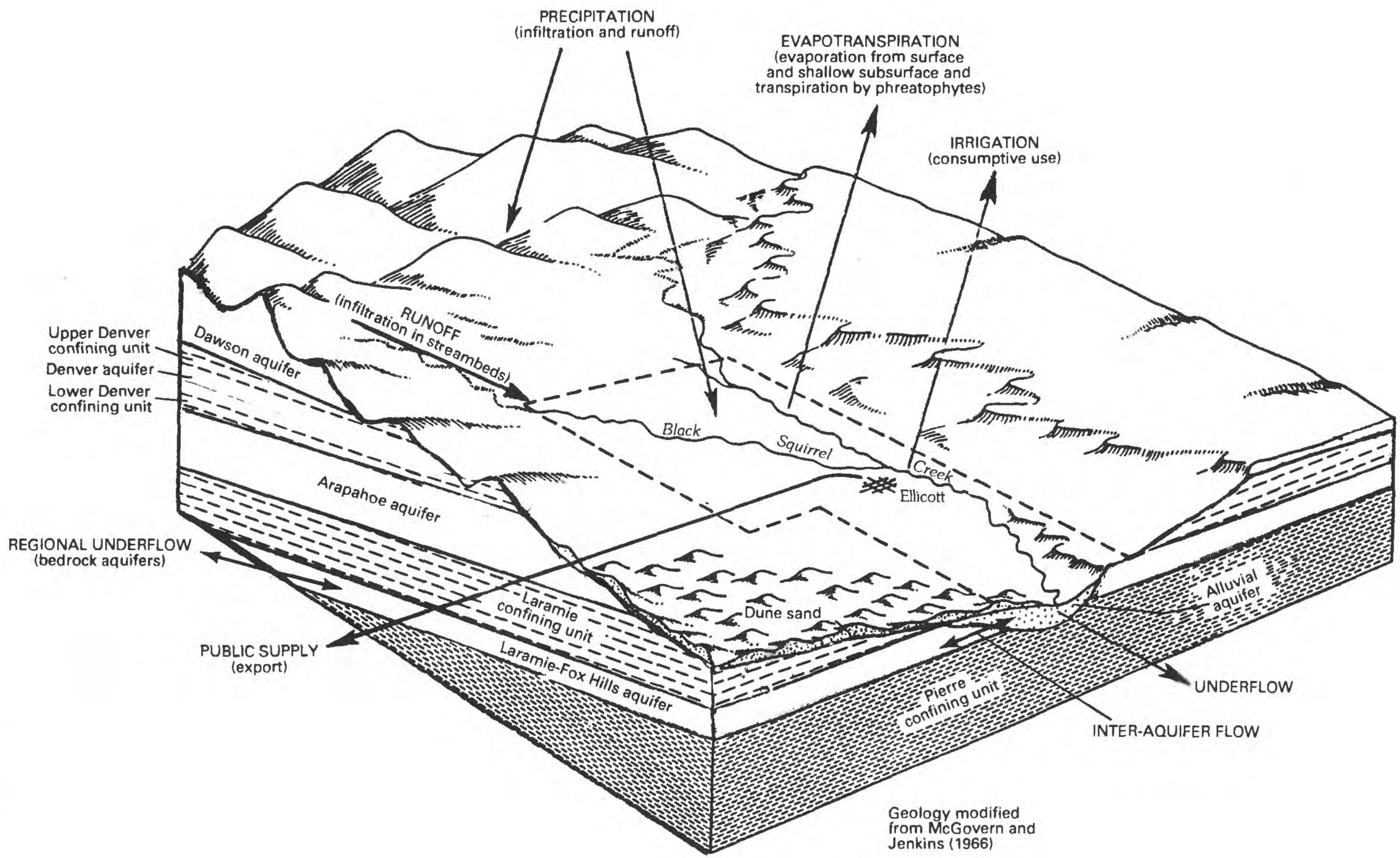


Figure 29. Conceptual model of the alluvial/bedrock aquifer system in the upper Black Squirrel Creek Basin.

Locally, the alluvial aquifer is hydraulically connected with the underlying bedrock aquifers. Flow of water in the alluvial aquifer is toward the south, and, locally, in the bedrock aquifers, is toward the upper Black Squirrel Creek Basin. During 1949–90, withdrawals from the alluvial aquifer, for irrigation and municipal supplies, resulted in substantial water-level declines in the alluvial aquifer. These water-level declines altered steady-state hydraulic gradients and specific discharge between the alluvial and bedrock aquifers.

A numerical model of three-dimensional ground-water flow, which is based on the conceptual model, was developed to quantify the flow between the alluvial aquifer and underlying bedrock aquifers and confining units in the upper Black Squirrel Creek Basin and to predict the potential effects of anticipated increases in withdrawals from the bedrock aquifers in El Paso County on water levels and the ground-water budget of the alluvial aquifer. The numerical model described in this report essentially couples the numerical model of two-dimensional flow in the alluvial aquifer (Buckles and Watts, 1988) with a numerical model of three-dimensional flow in the bedrock aquifers (Banta, 1989). Although based on these previous models, the model described in this report differs from them in several aspects and can be considered a refinement of both models.

In this model, recharge to the alluvial aquifer from infiltration of precipitation and surface water (runoff) is simulated as areal recharge. Discharge from the alluvial aquifer by evapotranspiration is simulated at a maximum rate of 2.5 ft/yr for steady-state, projection, and transient-state simulations (October 1949–March 1964) and at 5 ft per irrigation season (April–September) for the transient simulations (April 1964–September 1990). The maximum rate of evapotranspiration is simulated when water levels are at or above land surface; evapotranspiration is not simulated when water levels in the alluvial aquifer are 5 ft or more below land surface. Discharge from the alluvial aquifer to streams and springs is simulated as discharge to drains so that only outflow to streams and springs is computed. Underflow in the alluvial aquifer, across the southern limit of the study (model) area, is simulated using general-head boundaries. The head (water level) at the general-head boundaries was specified as the estimated water level 1.5 mi south (downgradient) of the model boundary. *Conductance* of the general-head boundaries was calculated as the product of the transmissivity of the alluvial aquifer and *cell* width (0.5 mi) divided by the distance (1.5 mi or 7,920 ft) to the downgradient *general-head boundary.* Flow between the alluvial aquifer and underlying bedrock is simulated as *head-dependent flow.* Recharge to the bedrock aquifer system from infiltration of precipitation in areas of outcrop is simulated at average rates, as defined previously by Banta (1989). Flow between the bedrock and surface-water bodies and saturated unconsolidated deposits along water courses, outside of the local model area, is simulated as flow to or from *constant-head boundaries.* Discharge from the bedrock to springs, unconsolidated deposits along water courses, and streams in upland areas, is simulated as discharge to constant-head boundaries, outside of the upper Black Squirrel Creek Basin and as discharge to drains in the upper Black Squirrel Creek Basin. Discharge from the bedrock by wells is simulated at constant rates, as previously estimated by Banta (1989).

The numerical model of two-dimensional flow in the alluvial aquifer of Buckles and Watts (1988) did not simulate flow between the alluvial aquifer and underlying bedrock. Buckles and Watts assumed that vertical flow between the alluvial aquifer and underlying bedrock was included as an unknown component in a lumped areal recharge term. Therefore, the effects of withdrawals from the bedrock aquifers on the alluvial aquifer could not be evaluated with the model of Buckles and Watts.

The numerical model of three-dimensional flow in the bedrock aquifers, used by Banta (1989), simulated conditions in the bedrock aquifers but did not explicitly model conditions in overlying alluvial aquifers. The alluvial aquifer in the upper Black Squirrel Creek Basin, as well as alluvial aquifers in other parts of the Denver Basin, were considered to be constant-head boundaries of the bedrock aquifer system. Banta (1989) used four model layers and a quasi three-dimensional approach to model flow in the bedrock aquifer system; this approach did not consider storage in confining units.

The numerical model of three-dimensional flow in the alluvial/bedrock aquifer system, which is described in this report, was developed in two stages: First, the model was used to simulate predevelopment (pre-1949) steady-state water levels in the alluvial aquifer of the upper Black Squirrel Creek Basin, then the model was used to match historical (1949–90) transient-state conditions and water levels. The model of transient-state conditions was later used to predict the potential effects of hypothetical increases in withdrawals from the bedrock aquifers in El Paso County on water levels and the ground-water budget of the alluvial aquifer for a 50-yr projection.

Model Description

The modular finite-difference model, MODFLOW, version 1638 (McDonald and Harbaugh, 1988), was used to simulate three-dimensional flow of water in the alluvial/bedrock aquifer system. A finite-difference approach is used in MODFLOW to solve a partial-differential equation that describes the three-dimensional flow of water of constant density through porous media (McDonald and Harbaugh, 1988, eq. 1, chap. 2, p. 1). This partial-differential equation describes flow of water under non-equilibrium (transient) conditions in porous media; the porous media can be heterogeneous and anisotropic. In the model, the principal axes of the hydraulic-conductivity tensor are assumed to be orthogonal to (aligned with) the model coordinate system. The finite-difference method approximates the continuous system defined by the flow equation at fixed points in both space and time.

Starting water levels, the geometry, hydraulic properties, recharge and discharge rates, and boundary conditions of each hydrogeologic unit are described at nodes, which represent the centers of rectangular cells of porous media. The values specified at each node are assumed to be uniform (homogeneous) within the cell. Anisotropy in the horizontal plane is specified at the interface between laterally adjacent cells. However, horizontal anisotropy was not specified for cells in this model. Anisotropy in the vertical direction is specified as a property of the interface between vertically adjacent cells. The layered *heterogeneity* of the alluvial/bedrock aquifer system causes the system to be anisotropic in the vertical direction; therefore, the aquifer system was modeled by subdividing the system into layers. Cells within each layer are simulated as homogeneous and isotropic volumes of porous media; therefore, in this model, anisotropy is not simulated within a cell, only between vertically adjacent cells.

Model Grid and Layers

The regional model grid (fig. 30A) for this study includes the same area modeled by Banta (1989, p. 2) and also includes within it, the area modeled by Buckles and Watts (1988, fig. 15). The grid spacing used by Banta (1989, pl. 2) was modified to correspond with the 0.5-mi grid spacing used by Buckles and Watts (1988, fig. 15). The model grid consists of 99 rows of 54 columns of cells (fig. 30A). The minimum grid spacing is 0.5 mi in the upper Black Squirrel Creek Basin, hereinafter referred to as the local model area (fig. 30B), and maximum grid spacing is 7 mi. For purposes of comparison of water levels and water budgets with results from the previous model (Buckles and Watts, 1988), results of this model (simulated water levels and water budgets) are shown only for the area inclusive of rows 52–99 and columns 25–44, the local model area (fig. 30B). The regional model boundaries of Banta (1989) were used so that artificial boundaries would be located as distantly from the local model area as possible. The local model area model approximates that part of the upper Black Squirrel Creek Basin that is underlain by the alluvial aquifer. The local model area is slightly larger than the area modeled by Buckles and Watts because the southern limit of the alluvial aquifer (layer 1) was extended 1.5 mi further south than the southern limit of the alluvial aquifer, as modeled by Buckles and Watts (1988). The model grid contains a total of 42,768 cells, but calculations are done by the model for less than half of these cells by the model.

The alluvial/bedrock aquifer system was subdivided into eight layers, with each layer representing one of the major hydrogeologic units listed in table 1. Layers 1, 2, 4, 6, and 8 represent aquifers, and layers 3, 5, and 7 are confining units (table 1). The basal confining unit, the Pierre confining unit, is a regionally persistent and a relatively thick and impermeable confining unit and is assumed to be a no-flow boundary at the base of the alluvial/bedrock aquifer system.

Layers may be specified, in MODFLOW, as either: (1) Unconfined; (2) confined; (3) confined/unconfined having transmissivity, which is constant; or (4) confined/unconfined having transmissivity, which may vary if simulated water levels fall below the top of the hydrogeologic unit. The alluvial aquifer (layer 1) was assumed to be unconfined. The Dawson aquifer (layer 2), the Denver aquifer (layer 4), the Arapahoe aquifer (layer 6), and the Laramie-Fox Hills aquifer (layer 8) were assumed to be confined or unconfined, depending on the position of simulated water levels relative to the top of the aquifer. Therefore, the transmissivity of layers 2, 4, 6, and 8 was allowed to vary as a function of saturated thickness. The upper Denver confining unit (layer 3), the lower Denver confining unit (layer 5), and the Laramie confining unit (layer 7) were assumed to be confined units with constant transmissivity.

Because the thickness of hydrogeologic units generally varies spatially, the thickness of the corresponding layer in the model also varies and results in a "deformed finite-difference mesh" (McDonald and Harbaugh, 1988, chap. 2, p. 30). A deformed finite-difference mesh has cell faces that might not be rectangular and cells in which the major axes of hydraulic conductivity might not be aligned with the model axes. Layers that conform to hydro-

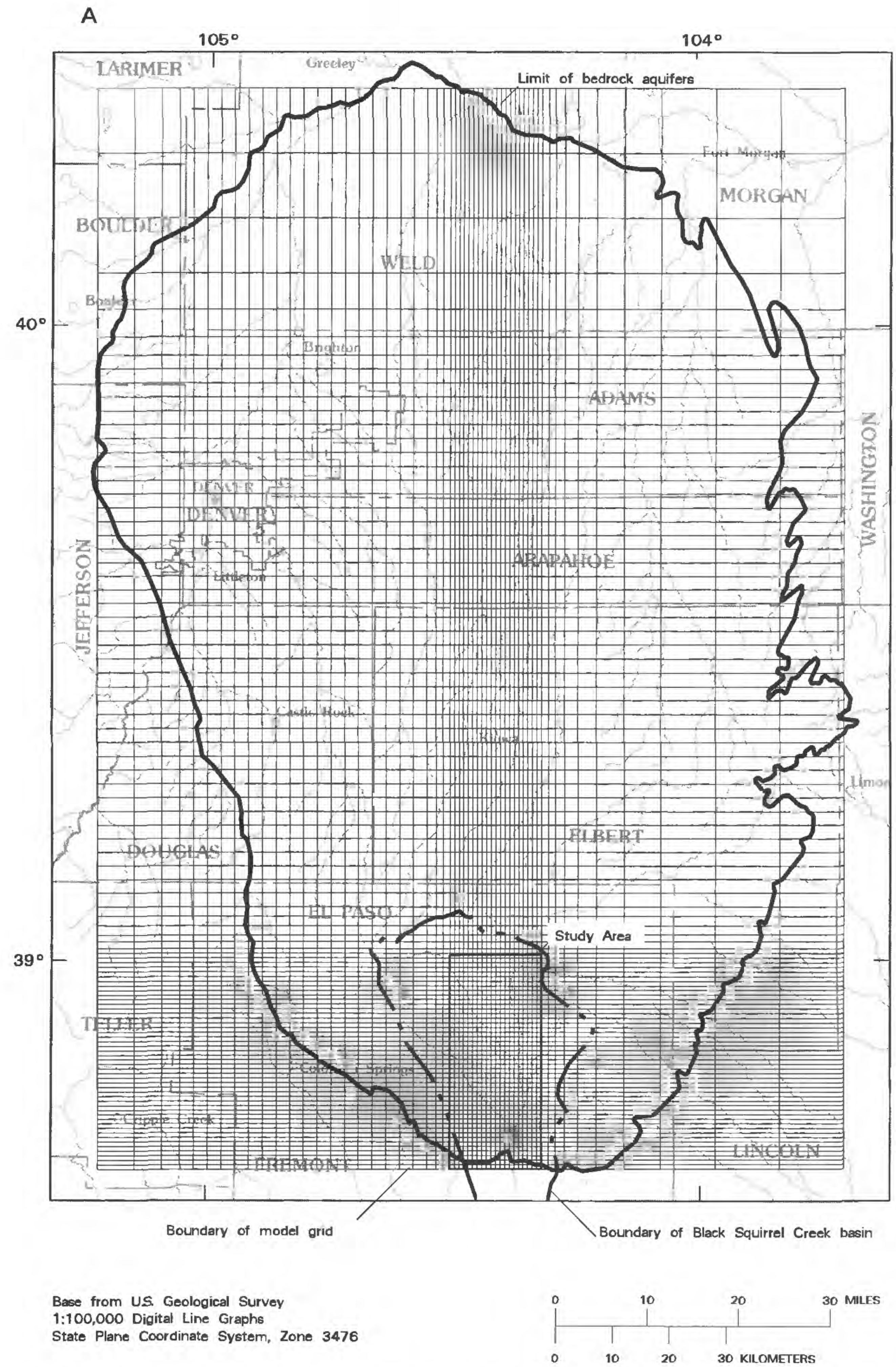


Figure 30. (A) Finite-difference grid for the alluvial/bedrock aquifer system and (B) detail of the finite-difference grid for the local model area.

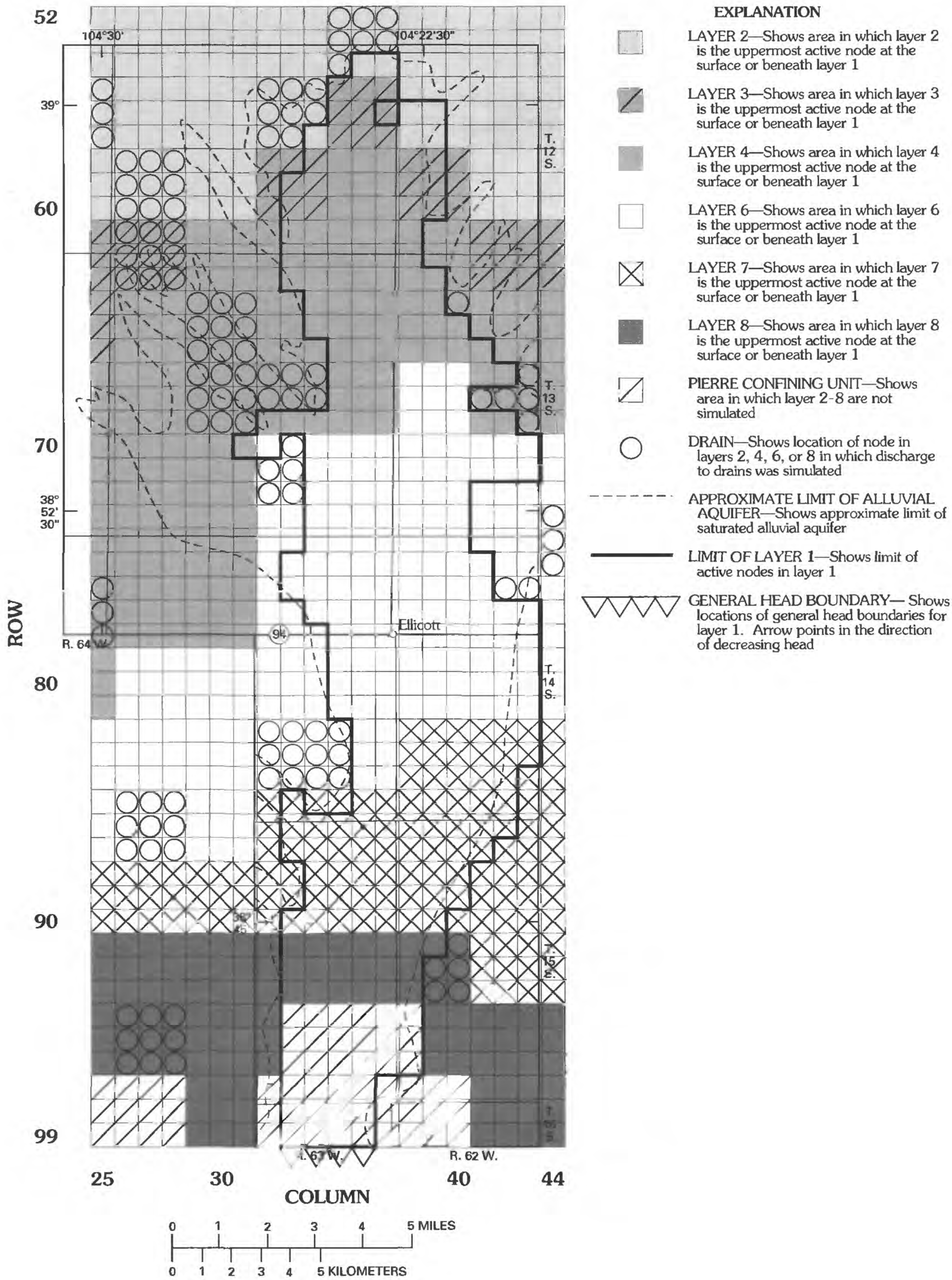


Figure 30. (A) Finite-difference grid for the alluvial/bedrock aquifer system and (B) detail of the finite-difference grid for the local model area--Continued.

Table 7. Sources of data for the numerical model of ground-water flow in the upper Black Squirrel Creek Basin

[Source of data: 1m, modified from Buckles and Watts (1988); 2m, modified from Banta (1989); 3m, modified from Robson and Romero (1981b); +, this study; --, not used; 1, Buckles and Watts (1988); 2, Banta (1989); t10, table 10; t11, table 11]

Model input	Layer number in numerical model							
	1	2	3	4	5	6	7	6
External geometry								
Altitude of top	1m	2m	2m	2m	3m	2m	2m	2m
Altitude of bottom	1m	2m	3m	2m	2m	2m	2m	2m
Starting water levels	1m	2m	2m	2m	2m	2m	2m	2m
Hydraulic properties								
Specific storage	--	2m	t11	2m	t11	2m	t11	2m
Specific yield	1	2	--	2	--	2	--	2
Hydraulic conductivity, horizontal	1	2	t10	2	t10	2	t10	2
Hydraulic conductivity, vertical	+	+	2	+	2	+	2	+
Recharge rates	1	2	--	2	--	2	--	2
Evapotranspiration								
Altitude of surface	1m	--	--	--	--	--	--	--
Extinction depth	1	--	--	--	--	--	--	--
Rate	1	--	--	--	--	--	--	--
Well discharge rates	1	2m	-	2m	--	2m	--	2m
Drains								
Altitude	1	2m	--	2m	--	2m	--	2m
Conductance	1	2m	--	2m	--	2m	--	2m
Boundaries								
Constant head	--	2m	--	2m	--	2m	--	2m
General head	+	--	--	--	--	--	--	--
No flow, lateral	1m	2	+	2	+	2	+	2
No flow, vertical	1m	--	--	--	--	--	--	2

geologic units, however, are more likely to satisfy the assumptions of *homogeneity* and isotropy than are layers that have uniform thickness but do not conform to stratigraphic irregularities (McDonald and Harbaugh, 1988, chap. 2, p. 29–31).

Simulation of vertical flow within a hydrogeologic unit requires that the unit be represented by more than one layer in the model. However, the primary objective of this study was to evaluate flow between the alluvial and bedrock aquifers; therefore, hydrogeologic units were modeled as single layers. The use of multiple layers to represent a hydrogeologic unit would be warranted if the objective of the study were to simulate the vertical distribution of water levels within the hydrogeologic unit.

Model Input

Sources and types of data used as input for the alluvial aquifer (layer 1) are summarized in table 7. The external geometry, starting water levels, hydraulic properties, areal recharge rates, evapotranspiration factors, well-discharge rates, drain factors, and no-flow boundaries for the alluvial aquifer (layer 1) were modified from input data for the numerical model of two-dimensional flow in the alluvial aquifer (Buckles and Watts, 1988). Modifications to input data for the alluvial aquifer consisted of: (1) Extension of the southern boundary 1.5 mi (3 rows) to the south; (2) altitudes of the bottom, starting water levels, and evapotranspiration (land) surface in the extended area; (3) extension of the lateral no-flow boundary to encompass the extended area; and (4) the addition of a permeable lower boundary in areas in which the bedrock hydrogeologic units underlie the alluvial aquifer. The altitudes of the bottom and starting water levels in the extended area were determined from maps (Romero, 1992), and the altitude of the land surface in the extended area was determined from topographic maps. In addition, the southern boundary of the alluvial aquifer was modified from the constant-flow boundary of Buckles and Watts (1988) to a general-head boundary.

Sources and types of data used as input for the bedrock aquifers (layers 2, 4, 6, and 8) also are summarized in table 7. The external geometry, starting water levels, hydraulic properties, areal recharge rates, well-discharge rates, drain factors, and locations of constant-head and no-flow boundaries for the bedrock aquifers (layers 2, 4, 6, and 8) were modified from input for the numerical model of three-dimensional flow in the Denver ground-water basin (Banta, 1989). Modifications to input data for the bedrock aquifers consisted of: (1) Adjustment of matrices of top and bottom altitudes to conform to the model grid, (2) correction of altitudes of the top and bottom of the aquifers in areas in which they exceeded the altitude of the bottom of the alluvial aquifer, (3) conversion of storage coefficient matrices to a specific-storage value and a thickness for each cell, and (4) adjustment of areal recharge rates and well-discharge rates to the new model grid. In the local model area, the constant-head boundaries used by Banta (1989, pl. 2) were converted to drain nodes.

Sources and types of data used as input to the model for the bedrock confining units (layers 3, 5, and 7) also are summarized in table 7. The altitudes of the tops of the bedrock confining units were assumed to equal: (1) The altitude of the bottom of the overlying bedrock aquifer; (2) the altitude of the land surface, in areas where overlying hydrogeologic units are missing; or (3) the altitude of the base of the alluvial aquifer. Similarly, the altitude of the bottom of the bedrock confining units was assumed to equal the altitude of the top of the underlying bedrock aquifer. Horizontal hydraulic conductivity of the bedrock confining units were assigned values that were 1 percent of the vertical hydraulic conductivity values that were estimated by Banta (1989, p. 25, table 2) during calibration of his model. Small values of horizontal hydraulic conductivity were used to constrain horizontal flow in the confining units. Recharge, evapotranspiration, well discharge, and drains were not simulated for the bedrock confining units. No-flow boundaries were assumed at the lateral extents of the confining units.

Vertical leakance at the bases of the alluvial aquifer and of the upper six bedrock hydrogeologic units were calculated, using equation 51 of McDonald and Harbaugh (1988, ch. 5, p. 13). The vertical leakance (Vcont) at the surface between vertically adjacent cells is calculated as:

$$Vcont_{i,j,k+\frac{1}{2}} = \frac{1}{\dfrac{(\Delta V_{i,j,k})/2}{K_{zi,j,k}} + \dfrac{(\Delta V_{i,j,k+1})/2}{K_{zi,j,k+1}}} \qquad (7)$$

where

$Vcont_{i,j,k+\frac{1}{2}}$ is the vertical leakance between nodes $_{i,j,k}$ and $_{i,j,k+1}$;

$\Delta V_{i,j,k}$ is the thickness of the upper cell;

$\Delta V_{i,j,k+1}$ is the thickness of the lower cell;

$K_{zi,j,k}$ is the vertical hydraulic conductivity of the upper cell; and

$K_{zi,j,k+1}$ is the vertical hydraulic conductivity of the lower cell.

Vertical leakance is not calculated at the base of cells in layer 8 because it is assumed that this surface is impermeable.

Steady-State (Pre-1949) Conditions

The model was calibrated to simulate steady-state (predevelopment) conditions that existed prior to large withdrawals of water from the alluvial and bedrock aquifers in the study area, prior to 1949. Although withdrawals from the bedrock aquifers in the Denver area began in the 1880's (Robson, 1987, p. 23), water levels in the upper Black Squirrel Creek Basin probably were not affected substantially until the 1960's, when withdrawals from the alluvial aquifer began to cause large water-level declines in the alluvial aquifer and, therefore, changes in vertical hydraulic gradients between the alluvial aquifer and underlying bedrock hydrogeologic units.

The model was calibrated to steady-state water levels in the alluvial aquifer (layer 1) by trial-and-error reductions to simulated recharge rates to the alluvial aquifer and by trial-and-error modification to values of the vertical hydraulic conductivity of the alluvial and bedrock aquifers (layers 1, 2, 4, 6, and 8). The vertical hydraulic-conductivity values of the bedrock confining units (layers 3, 5, and 7) were assumed to be known and were not changed during calibration. The model was considered calibrated when the trial-and-error modifications to model input did not substantially decrease the mean square error (M.S.E.) of simulated heads of layers 1, 2, 4, 6, and 8, with starting heads of the respective layers in the study area. Steady-state water-level data for the bedrock confining units (layers 3, 5, and 7) were unknown and, therefore, M.S.E. was not calculated for layers 3, 5, and 7. Simulated steady-state water levels for layers 3, 5, and 7 cannot be verified with historical water-level data. The M.S.E. was calculated as:

$$\text{M.S.E.} = \overline{\Delta H}^2 + S^2 \qquad (8)$$

where

$\overline{\Delta H}^2 = \frac{1}{n}\sum_{i=1}^{n} \Delta Hi^2$;

$\Delta H_i = H_{ci} - H_{si}$;

H_{ci} = model computed water level (head) at node i;

H_{si} = starting head at node i;

n = the number of nodes with both H_{ci} and H_{si};

$\Delta H = \frac{1}{n}\sum_{i=1}^{n} \Delta Hi$; and

$S^2 = 1/(n-1)\sum_{i=1}^{n} (\Delta H_i - \overline{\Delta H})^2$

The M.S.E., when minimized, produces an estimate with small bias and small variance, which means that the mean error ($\Delta\overline{H}$) is near zero and the variance of the errors (S^2) about $\Delta\overline{H}$ is small (Wonnacott and Wonnacott, 1984, p. 197–198).

The simulated steady-state water levels are shown in figure 31, as contours of the depth-to-water surface, for comparison with reported pre-1956 depth-to-water values. Computed depth to water was calculated by subtraction of computed water levels from the altitude of the land surface, which was used as the evapotranspiration surface in the model. Errors in estimating average land-surface altitude at cells probably are less than 10 to 20 ft, depending on the contour interval of the map and topographic relief of the land surface in the cell. The model of steady-state conditions was accepted as calibrated when the computed depth-to-water surface of layer 1 was less than or equal to reported pre-1956 depth-to-water values and also was below land surface.

The distribution of steady-state recharge to the alluvial aquifer (fig. 4) used in this model is similar to that used by Buckles and Watts (1988). However, total areal recharge simulated in this model is about 93 percent of the total areal recharge simulated by Buckles and Watts (1988, table 5). The difference in total recharge is a result of the simulation of vertical flow between the alluvial aquifer and underlying bedrock. The steady-state ground-water budget simulated for the alluvial aquifer (table 8) is slightly different than the steady-state ground-water budget presented by Buckles and Watts (1988, table 5) because this ground-water budget separates vertical flow between the alluvial aquifer and bedrock aquifers and confining units from areal recharge. Total simulated steady-state inflow to and outflow from the alluvial aquifer are about 1.1 percent larger than the simulated inflow and outflow rates reported by Buckles and Watts (1988).

This model is a refinement of the previous model of flow in the alluvial aquifer by Buckles and Watts because it quantifies separately areal recharge from the surface and flow between the alluvial aquifer and underlying bedrock. The previous model of the alluvial aquifer could not be used to quantify the historical or potential effects of water use from the bedrock aquifers on water levels in the alluvial aquifer. Buckles and Watts reported that recharge to the alluvial aquifer was about 12.7 ft^3/s (9,200 acre-ft/yr) and included recharge from infiltration of precipitation and surface water and discharge from the bedrock. This model further refines the estimate by quantifying separately the components of recharge from above (11.97 ft^3/s) and from below (0.87 ft^3/s). Infiltration of precipitation and surface water is the primary source of recharge to the aquifer, about 93 percent of total recharge; upward flow from the bedrock is a minor source of recharge to the aquifer, about 7 percent of total recharge. The model also indicates that during steady-state conditions, prior to October 1949, underflow across the southern limit of the study area, about 6.26 ft^3/s, and evapotranspiration, about 5.93 ft^3/s, were the primary mechanisms of discharge from the alluvial aquifer.

The combined steady-state budget for the bedrock hydrogeologic units, layers 2–8 (table 9), cannot be compared directly with the simulated steady-state discharge to the Black Squirrel Creek Basin that was reported by Banta (1989, table 4) because of differences in boundary conditions and consideration of local flow components. Banta simulated discharge to the Black Squirrel Creek Basin from bedrock aquifers of about 1.71 ft^3/s, whereas the ground-water budget for the bedrock hydrogeologic units in the local model area includes both inflow and outflow components. Simulated net discharge from the bedrock to the local model area includes: Discharge to drains (springs and streams) of 0.41 ft^3/s and net discharge to the alluvial aquifer of 0.82 ft^3/s (upward flow minus downward flow; 0.87 – 0.05 = 0.82) was 1.23 ft^3/s (0.41 + 0.82 = 1.23). Net simulated inflow to the bedrock hydrogeologic units in the local model area (table 9) includes: Net underflow (inflow minus

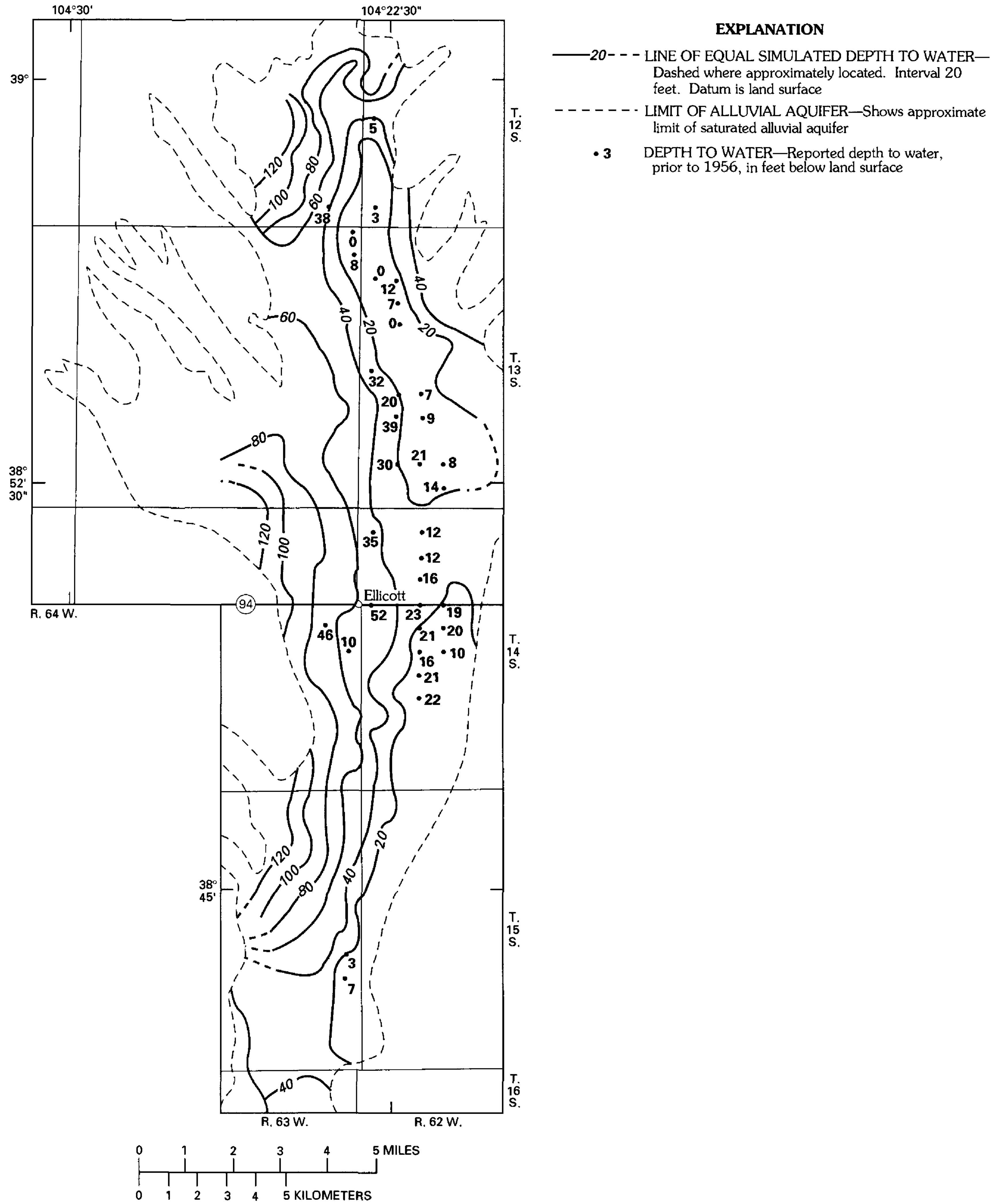


Figure 31. Simulated steady-state depth-to-water surface and reported pre-1956 depth-to-water values for the alluvial aquifer (layer 1).

Table 8. Simulated steady-state (predevelopment) ground-water budget for the alluvial aquifer in the upper Black Squirrel Creek Basin

[Note: Instantaneous rates are rounded to the nearest 0.01 cubic foot per second, and annual volumes are rounded to the nearest 10 acre-feet per year]

Water-budget component	Instantaneous rate (cubic feet per second)	Annual volume (acre-feet per year)
	Inflows	
[1]Areal recharge	11.97	8,670
Upward flow from bedrock	.87	630
Total inflow	12.84	9,300
	Outflows	
Evapotranspiration	5.93	4,290
Underflow	6.26	4,540
[2]Drains	.60	430
Downward flow to bedrock	.05	40
Total outflows	12.84	9,300

[1] Areal recharge includes deep percolation of precipitation and infiltration of surface water from ephemeral streams.

[2] Outflow to drains represents discharge of ground water to streambeds and springs.

Table 9. Simulated steady-state (predevelopment) ground-water budget for the bedrock hydrogeologic units in the upper Black Squirrel Creek Basin

[Note: Instantaneous rates are rounded to the nearest 0.01 cubic foot per second, and annual volumes are rounded to the nearest 10 acre-feet per year]

Water-budget component	Instantaneous rate (cubic feet per second)	Annual volume (acre-feet per year)
	Inflows	
Areal recharge	0.67	490
Underflow	1.12	810
Downward flow from alluvial aquifer	.05	40
Total inflow	1.84	1,340
	Outflows	
[1]Drains	0.41	300
Underflow	.56	410
Upward flow to alluvial aquifer	.87	630
Total outflow	1.84	1,340

[1] Drains represent discharge to springs and streams in areas of the upper Black Squirrel Creek Basin that are not underlain by the alluvial aquifer.

outflow; 1.12 – 0.56 = 0.56) of 0.56 ft^3/s and areal recharge to the bedrock of 0.67 ft^3/s. Areal recharge to the bedrock in the discharge area of the upper Black Squirrel Creek Basin was not simulated by Banta's model. Simulated areal recharge to the bedrock in the discharge area of the upper Black Squirrel Creek Basin was 0.67 ft^3/s, thus, simulated discharge from the bedrock aquifers in Banta's model corrected for local recharge was 1.04 ft^3/s (1.71 – 0.67 = 1.04). The difference, 0.19 ft^3/s (1.23 – 1.04 = 0.19), between Banta's estimate of regional bedrock discharge to the upper Black Squirrel Creek Basin, corrected for local recharge to the bedrock, of 1.04 ft^3/s and the discharge estimated by this model (1.23 ft^3/s) primarily results from differences in boundary conditions and consideration of local flow paths in this model. Because this model simulates the alluvial aquifer as a layer and does not include constant-head nodes in the local model area, the simulated steady-state ground-water budget for the bedrock (table 9) might be a more realistic estimate of regional ground-water discharge to the basin than previous estimates.

Values of vertical hydraulic conductivity for each of the hydrogeologic units (layer 1–8) used in the model of steady-state conditions are listed in table 10. These values are the average values of vertical hydraulic conductivity of each layer and might not be equivalent to values determined by aquifer testing of thin lithologic units with the hydrogeologic units (table 2). The bedrock hydrogeologic units, layers 2–8, locally, are heterogeneous and contain both water-yielding and confining intervals. Site-specific values might differ substantially from average regional values. The equivalent vertical hydraulic conductivity of a layered heterogeneous system predominantly is controlled by the small values of vertical hydraulic conductivity of included confining intervals (Freeze and Cherry, 1979). The equivalent vertical hydraulic conductivity is approximated by the harmonic mean of hydraulic-conductivity values of the layers and is given by Freeze and Cherry (1979, eq. 2.31) as:

$$\bar{K}_z = \frac{\sum_{i=1}^{n} b_i}{\sum_{i=1}^{n} (b_i/K_{zi})} \qquad (9)$$

where

K_z = harmonic mean of vertical hydraulic conductivity of a layered system;

b_i = thickness of layer i; and

K_{zi} = vertical hydraulic conductivity of layer i.

For example, in a two-layer system in which $b_1 = b_2$ = 10 ft, K_{z1} = 100 ft/d, and $K_{z2} = 1 \times 10^{-5}$ ft/d, $\overline{K}_z$ = (10 + 10) ft/(10 ft/100 ft/d + 10 ft/1 × 10^{-5} ft/d) = 2×10^{-5} ft/d. It is apparent from this example that the vertical hydraulic conductivity of a heterogeneous layered system can be approximated by the smaller values of vertical hydraulic conductivity of included confining intervals. Because the bedrock aquifers generally consist of conglomerate sandstone and siltstone, which contain interbedded shale, the vertical hydraulic conductivity of the interbedded shale could have a profound effect on $\overline{K}_z$ for the bedrock aquifers.

Table 10. Vertical hydraulic-conductivity values used in the model of the alluvial/bedrock aquifer system and vertical hydraulic conductivity that was determined by an aquifer test in the upper Black Squirrel Creek Basin

[The vertical hydraulic-conductivity values of layers 3, 5, and 7 were modified from Banta (1989, table 2); --, not determined]

Layer number in numerical model	Hydrostratigraphic unit	Vertical hydraulic conductivity used in the model (feet per day)	Vertical hydraulic conductivity determined by aquifer test (feet per day)
1	Alluvial aquifer	4	3
2	Dawson aquifer	2.0×10^{-4}	--
3	Upper Denver confining unit	4.1×10^{-5}	--
4	Denver aquifer	3.0×10^{-4}	8×10^{-4}
5	Lower Denver confining unit	1.3×10^{-5}	--
6	Arapahoe aquifer	4.0×10^{-4}	--
7	Laramie confining unit	6.2×10^{-7}	--
8	Laramie-Fox Hills aquifer	1.1×10^{-3}	--

Flow between layers in the model is simulated only in the vertical direction, whereas flow across the unconformable contact between the alluvium and bedrock likely includes horizontal and vertical components of flow. Consequently, the values of vertical hydraulic conductivity used in the model (table 10) might be larger than the value of vertical hydraulic conductivity that would be determined by aquifer tests of the bedrock hydrogeologic units. Because of the dominant effect of small values of vertical hydraulic conductivity of the confining units in calculation of vertical conductance between layers, values of vertical hydraulic conductivity of the aquifers for layers 1, 2, 4, 6, and 8 are important primarily in areas in which bedrock confining units do not separate bedrock aquifers from the alluvial aquifer (fig 30b).

Transient-State (October 1949–September 1990) Conditions

Simulated water levels from the model for steady-state conditions were used as starting water levels for simulation of October 1949–September 1990 transient-state conditions. Transient-state conditions are assumed to have been initiated in the alluvial aquifer of the upper Black Squirrel Creek Basin after 1949, when withdrawals of water from the alluvial aquifer for irrigation and municipal supplies increased. Withdrawals from the alluvial aquifer between 1964 and 1984 resulted in water-level declines of as much as 44 ft (Bingham and Klein, 1974; Buckles and Watts, 1988). Average depletion of storage, during 1964–84, in the alluvial aquifer was estimated to be about 5,000 acre-ft/yr. During 1949–90, substantial water-level declines, as large as 50 ft, in the alluvial aquifer resulted from withdrawals from the aquifer for irrigation and municipal supplies. Although the bedrock aquifers in the study area were relatively undeveloped prior to the 1980's, vertical hydraulic gradients between the alluvial aquifer and bedrock aquifers changed as a result of water-level declines in the alluvial aquifer and are assumed to have affected water levels in the bedrock aquifers. The model of transient-state conditions was used to simulate water-level declines in the alluvial aquifer during October 1949 through September 1990 by dividing the transient-state period into 57 stress periods (fig. 32). The period during which withdrawals of ground water from the alluvial aquifer increased rapidly, October 1949 through March 1964, was divided into four stress periods of 5-, 4-, 3-, and 2.5-yr duration. Stress periods 5 through 57 each simulate a 0.5-yr period to more closely approximate seasonal variations in withdrawals

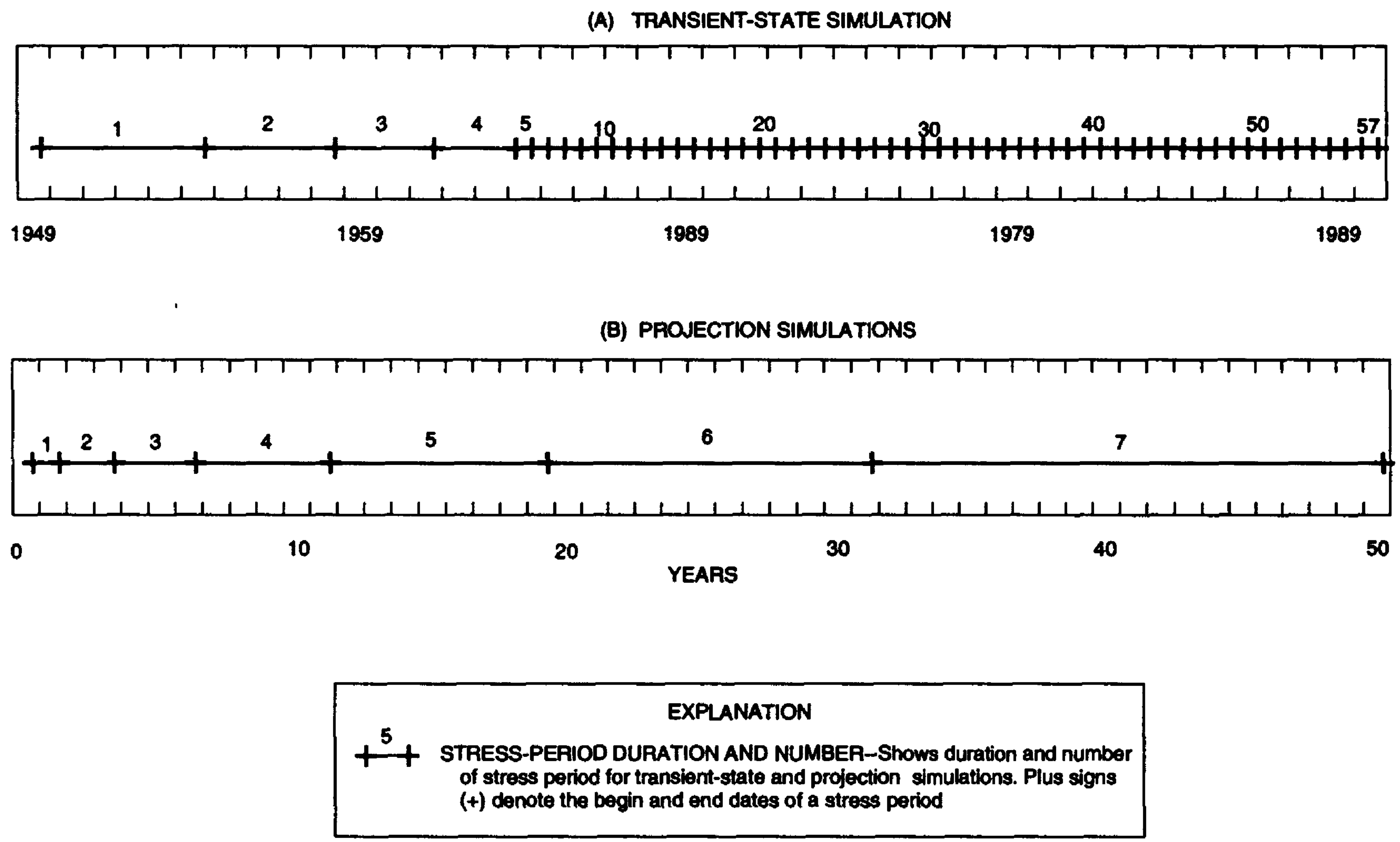


Figure 32. Stress-period durations for (A) the transient-state simulation, October 1949–September 1990, and (B) for the 50-yr projection simulations.

from the alluvial aquifer. Withdrawals from the alluvial aquifer for irrigation supply generally occur during April through September, the irrigation season, and withdrawals for municipal supply occur throughout the year. Withdrawals from the bedrock aquifers were assumed to be constant during the transient simulation and, thus, do not closely simulate the seasonal pattern of irrigation withdrawals of some wells.

The model was calibrated primarily to simulate transient-state 1964–74 and 1974–84 water-level declines in the alluvial aquifer, by trial-and-error adjustment of storage properties of the bedrock hydrogeologic units (layers 2 through 8), and secondarily to match model-computed water levels and vertical gradient of head values with measured water levels and computed vertical gradient-of-head values for the eight pair wells. Because systematic measurement of water levels in the bedrock aquifers of the study area did not begin until June 1987, only model-computed heads for stress periods 49–57 are used for comparison with measured water-level and gradient-of-head values.

Because of similarities in the lithology of the bedrock aquifers, specific storage (S_s) of layers 2, 4, 6, and 8 was assumed to be the same value; likewise, S_s of the bedrock confining units (layers 3, 5, and 7) was assumed to be the same value. Specific-storage values simulated in the model of transient-state conditions are listed in table 11. The S_s of the bedrock aquifers (layers 2, 4, 6, and 8) used in this model was 1×10^{-6} per ft and is equivalent to one-half of the specific-storage value used by Robson (1987) and Banta (1989). However, total storage of the eight layers in this model roughly equals total storage in the four-layer models of Robson and Banta.

Hydrographs (figs. 7A–14A) of model-computed water levels are superimposed on hydrographs of measured water levels in eight alluvial bedrock well pairs. Model-computed water levels simulate average water levels in the layers at the end of 6-month-long stress periods, whereas measured water levels represent the average water level in the open interval of the observation wells, which were measured weekly, biweekly, or monthly. Therefore, the measured water levels fluctuate more than model-computed water levels. Differences between model-computed water levels and measured water levels also occur because model-computed water levels are computed as the average water level at the center of the cell, and measured water levels might be affected by vertical flow within the hydrogeologic unit. Although the model-computed water levels generally do not match the measured water levels closely, long-term water-level trends for model-computed water levels are approximately parallel with trends in measured long-term water levels for some

Table 11. Storage properties used in the model of the alluvial/bedrock aquifer system in the upper Black Squirrel Creek Basin

[NA, not applicable for the hydrogeologic unit as simulated]

Layer number in numerical model	Hydrogeologic unit	Specific storage (per foot)	Average specific yield (percent)
1	Alluvial aquifer	NA	[1]18
2	Dawson aquifer	1×10^{-6}	[2]18
3	Upper Denver confining unit	2×10^{-7}	NA
4	Denver aquifer	1×10^{-6}	[2]14
5	Lower Denver confining unit	2×10^{-7}	NA
6	Arapahoe aquifer	1×10^{-6}	[2]18
7	Laramie confining unit	2×10^{-7}	NA
8	Laramie-Fox Hills aquifer	1×10^{-6}	[2]20

[1]Source: Buckles and Watts (1988).
[2]Source: Robson (1987) and Banta (1989).

wells (figs. 7 and 11). Model-computed gradient of head and measured gradient of head are shown in figures 7B–14B. Absolute values of model-computed and measured gradients of head, greater than about 0.1 (for example, fig. 12B), indicate that there is no substantial hydraulic connection between the alluvial aquifer and underlying bedrock. Absolute values of model-computed and measured gradients of head that are almost zero (for example, fig. 8), indicate that bedrock underlying the alluvial aquifer is very leaky, and water levels in the bedrock will be equivalent to water levels in the alluvial aquifer. During each stress period of the transient-state simulation, wells were simulated to discharge at constant rates; in reality, most irrigation and municipal wells in the study area are pumped cyclically. Initially, wells are pumped until drawdown becomes excessive and the well is shut down; while the well is not being pumped, the water levels recover. This cyclic drawdown and recovery pattern is shown in the water-level hydrographs for observation wells near Cherokee Metropolitan District municipal supply wells (figs. 7–9). Short-term water-level fluctuations are not reproduced in the hydrographs of model-computed water levels because the duration of the stress periods is greater than that of the pumping cycles. The simulated transient-state groundwater budget for the alluvial aquifer (layer 1) for October 1949–September 1990 (stress periods 1–57) is shown in figure 33. Simulated net flow between

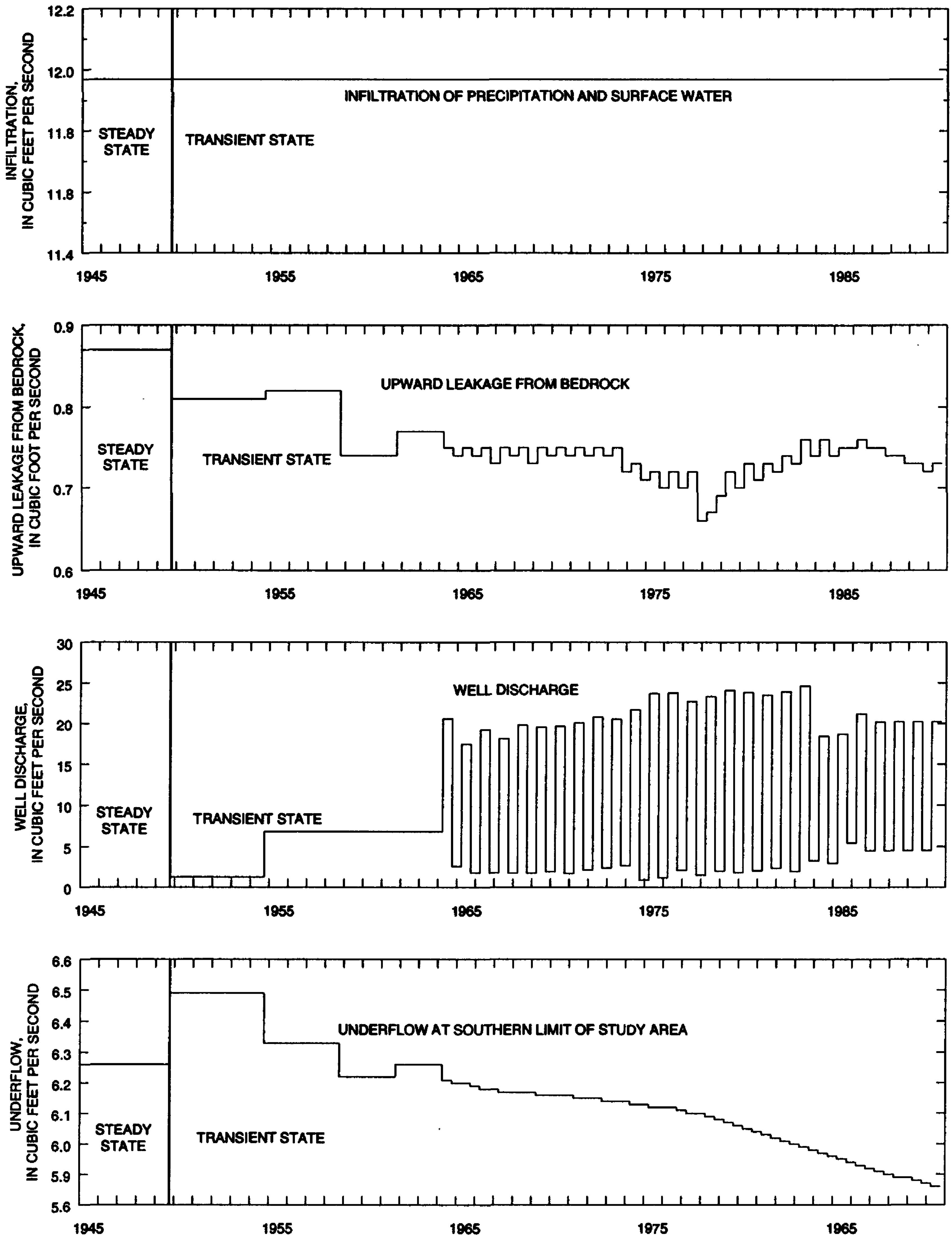


Figure 33. Simulated ground-water budget of the alluvial aquifer in the upper Black Squirrel Creek Basin for steady-state (pre-1949) conditions and for transient-state (October 1949–September 1990) conditions.

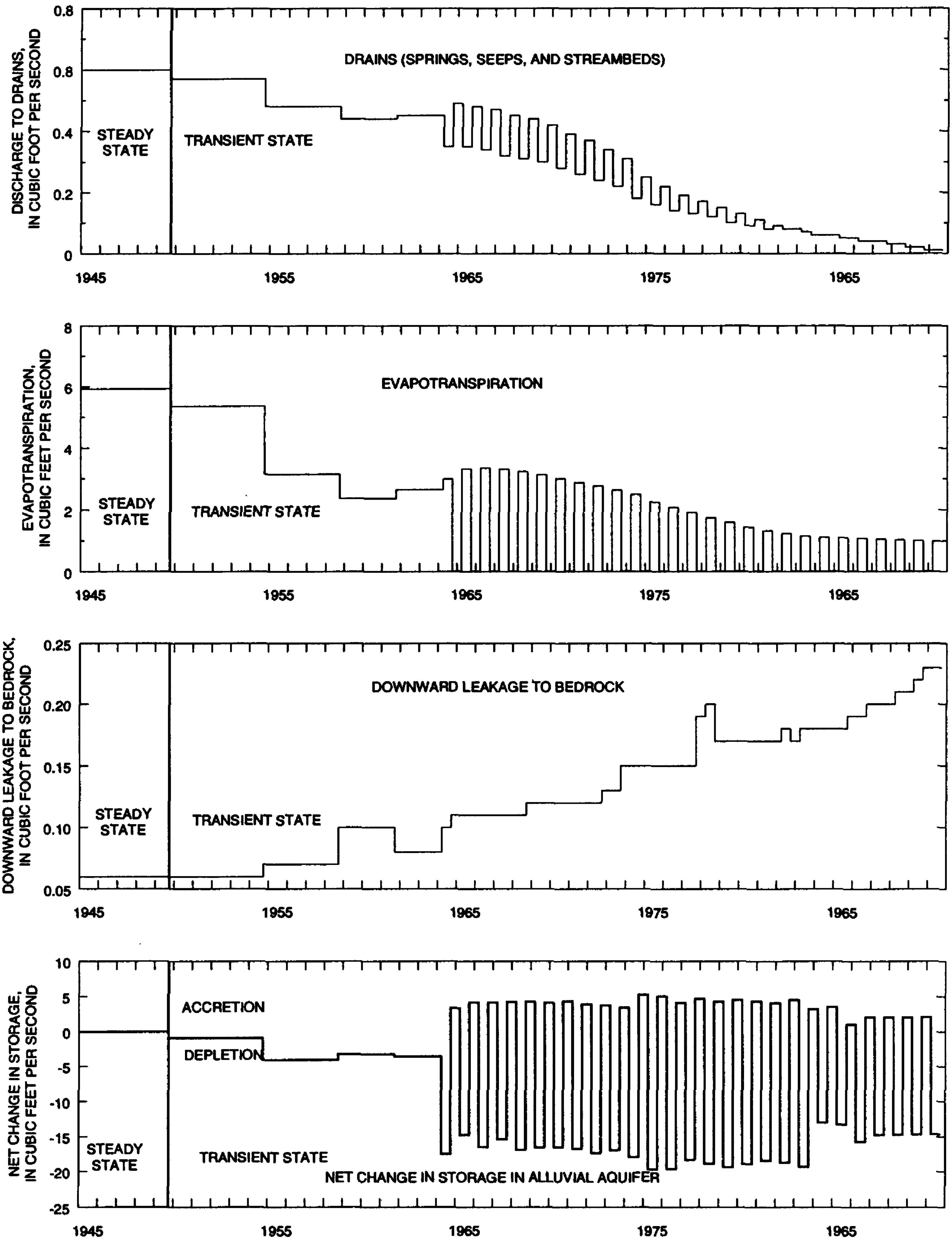


Figure 33. Simulated ground-water budget of the alluvial aquifer in the upper Black Squirrel Creek Basin for steady-state (pre-1949) conditions and for transient-state (October 1949–September 1990) conditions--Continued.

the alluvial aquifer and underlying bedrock is the difference between upward leakage to and downward leakage from the alluvial aquifer. For example, simulated net vertical flow for stress period 1 is 0.81 ft^3/s (0.87 – 0.06 = 0.81). The simulated net flow to the alluvial aquifer from the underlying bedrock decreased from 0.82 ft^3/s (0.87 – 0.05 = 0.82) for predevelopment conditions to 0.50 ft^3/s (0.73 – 0.23 = 0.50) during April–September 1990 (fig. 33).

The areal distribution of simulated vertical flow between the alluvial aquifer and underlying bedrock at the end of the 1990 irrigation season (April–September 1990) is shown in figure 34. Specific discharge is a negative value where flow is upward to the alluvial aquifer and is a positive value where flow is downward to the bedrock. Because simulated values of vertical flow are relatively small, less than the lower limit of the temperature-profile method of 3.2×10^{-8} ft/s for confining units less than several hundred feet thick (Bredehoeft and Papadopulos, 1965), it is unlikely that specific discharge could be estimated accurately from analyses of subsurface-temperature profiles in most parts of the study area. The pattern of simulated water-level decline in the alluvial aquifer from the start of stress period 5 (April 1964) to the end of stress period 24 (March 1974, fig. 6A) approximates the pattern of water-level decline reported by Bingham and Klein (1974, pl. 1). Similarly, the pattern of simulated water-level decline from the start of stress period 25 (April 1974) to the end of stress period 44 (March 1984, fig. 6B) is similar to the pattern of water-level decline reported by Buckles and Watts (1988, fig. 11). The average simulated decline in storage in the alluvial aquifer during April 1964–March 1974 was about 6.3 ft^3/s and during April 1974–March 1984 was about 7.3 ft^3/s (fig. 33). The total simulated depletion of the alluvial aquifer was about 190,000 acre-ft from October 1949–September 1990.

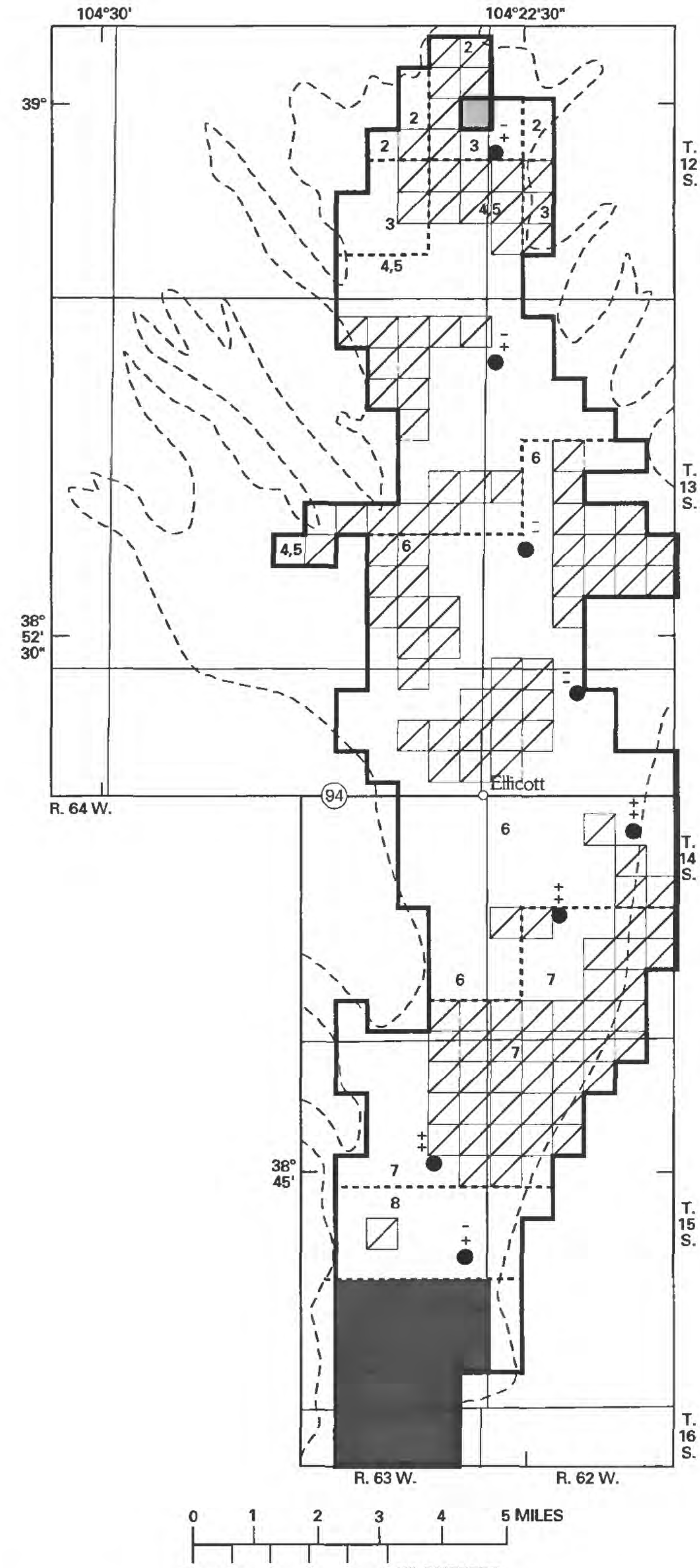
Limitations of the Model

The numerical model of ground-water flow is an approximation of a complex hydrogeologic system and, as such, is limited by the availability of accurate data to define the system. The model described in this report primarily was developed to: (1) Evaluate flow between the alluvial and bedrock aquifers, and (2) simulate the potential hydrogeologic effects to the alluvial aquifer of anticipated increased withdrawals from the bedrock aquifers in El Paso County. The model was calibrated to simulate historical (October 1949–September 1990) conditions, primarily in the alluvial aquifer. The simulation of historic conditions in the bedrock hydrogeologic units is not as realistic as the simulation of historic conditions in the alluvial aquifer.

The use of only seven layers to represent the heterogeneous bedrock hydrogeologic units makes direct comparison of simulated water levels with measured water levels in the bedrock hydrogeologic units difficult. Generally, measured water levels in the bedrock are from partially penetrating wells, whereas simulated water levels represent the average water level in an assumed homogeneous isotropic layer. How well the model is calibrated is difficult to assess because there is little historical hydrogeologic data for the bedrock hydrogeologic units in the study area. Local effects of withdrawals on the water levels for small areas (areas less than 160 acres in extent) cannot be simulated with the model described in this report. The model described in this report also cannot be used to simulate the vertical distribution of water levels within a hydrogeologic unit. The intended use of this model is evaluating regional, long-term changes in water levels in and the ground-water budgets of the alluvial aquifer that result from changes in recharge and discharge conditions in the bedrock aquifers.

Sensitivity Tests

Sensitivity tests of the model were done to define a range of potential error that could result from errors in values of vertical hydraulic conductivity (K_z) for the alluvial and bedrock aquifers and in values of specific storage (S_s) for the bedrock aquifers and confining units. The mean error and M.S.E. of simulated water levels for the alluvial and bedrock aquifers for selected sensitivity tests are listed in table 12. The mean error of simulated water levels for the model of steady-state conditions was calculated for the differences (errors) between the initial starting water levels and the simulated steady-state water levels. Therefore, the mean error for the model of steady-state conditions includes errors in both starting and simulated water levels. The mean error in simulated water levels for the sensitivity tests was calculated for the difference between simulated water levels for the model of steady-state conditions and the simulated water levels for the sensitivity tests. The M.S.E. was calculated, as described previously in the report "Steady-State (Pre-1949) Conditions." The mean error and M.S.E. for the sensitivity tests of the model of steady-state conditions quantify the amount of water-level change that results



EXPLANATION

UPWARD FLOW—Shows area in which simulated flow is upward into alluvial aquifer. Cumulative rates of upward flow from bedrock hydrogeologic units are in table on facing page

DOWNWARD FLOW—Shows area in which simulated flow is downward out of alluvial aquifer. Cumulative rates of downward flow to bedrock hydrogeologic units are in table on facing page

NO FLOW IN ALLUVIAL AQUIFER—Shows area within simulated limits of alluvial aquifer in which alluvial aquifer is simulated as unsaturated

NO FLOW IN BEDROCK HYDROGEOLOGIC UNITS—Shows area within simulated limits of alluvial aquifer in which Pierre confining unit underlies alluvial aquifer

APPROXIMATE LIMIT OF ALLUVIAL AQUIFER—Shows approximate limit of saturated alluvial aquifer

SIMULATED LIMIT OF ALLUVIAL AQUIFER—Shows limit of alluvial aquifer as simulated in model

7 / 8 SIMULATED LIMIT OF BEDROCK HYDROGEOLOGIC UNIT—Shows simulated limit of bedrock hydrogeologic units at base of alluvial aquifer. Numbers indicate layer number of uppermost bedrock hydrogeologic unit at base of the alluvial aquifer

VERTICAL FLOW ESTIMATES—Shows approximate location of well for which specific discharge across base of alluvial aquifer was estimated from subsurface temperature profile and for which gradient of head was calculated. Upper symbol (-/+) indicates direction of flow from temperature profile; minus sign indicates upward flow, plus sign indicates downward flow. Lower symbol (-/+) indicates direction of flow from gradient of head; minus sign indicates upward flow, plus sign indicates downward flow

Figure 34. Simulated discharge between the alluvial aquifer and underlying bedrock at the end of the 1990 irrigation season (September 1990).

Total simulated flow across the bottom of layer 1 base of the alluvial aquifer at the end of stress period 57, September 1990.

Layer number	Upward flow (cubic foot per second)	Downward flow (cubic foot per second)	Net vertical flow (cubic foot per second)
2	–0.006	0.000	–0.006
3	–0.002	.005	.003
4, 5	–0.311	.103	–0.208
6	–0.296	.098	–0.198
7	–0.009	.027	.018
8	–0.108	.002	–0.106
Totals	–0.732	0.235	–0.497

Figure 34. Simulated discharge between the alluvial aquifer and underlying bedrock at the end of the 1990 irrigation season (September 1990)--Continued.

from changing the vertical hydraulic-conductivity value for the test. The mean error and M.S.E. listed in table 12 were computed for nodes in layer 1 (the alluvial aquifer) and for nodes in layers 2, 4, 6, and 8 (the bedrock aquifers) in the local model area (rows 52–99, columns 25–44).

Simulated water levels for the alluvial aquifer were relatively insensitive to the values of vertical hydraulic conductivity tested (table 12). An order-of-magnitude decrease in the value of K_z for layer 1 and one and two order of magnitude increases in the value of K_z of the bedrock aquifers did not substantially affect mean error and M.S.E. of simulated water levels in the alluvial aquifer. Simulated water levels in the bedrock aquifers, particularly the Arapahoe and Laramie-Fox Hills, were affected by changes in the values of vertical hydraulic conductivity. Because there is little data for the bedrock aquifers in the study area, the values of K_z can not be evaluated rigorously. The vertical hydraulic conductivity of a 15-ft-thick siltstone and sandstone in the lower part of the Denver aquifer at well SC01306207BCB4 was estimated to be about 8×10^{-4} ft/d, which is within the range of values used for the bedrock aquifer, 2×10^{-4} to 1×10^{-3} ft/d (table 10).

Sensitivity tests of the model of transient-state conditions were done for selected values of specific storage for the bedrock aquifers (layers 2, 4, 6, and 8) and for the bedrock confining units (layers 3, 5, and 7). However, these tests were inconclusive because there were insufficient measurements of water-level change in the bedrock aquifers during this period for comparison with simulated water-level change.

Potential Effects of Increased Withdrawals From the Bedrock Aquifers on Water Levels and Flow Between the Alluvial and Bedrock Aquifers

After calibration of the model to simulate historic water-level change in the alluvial aquifers, the model was used to evaluate the potential effects of anticipated increases in withdrawals by wells from the bedrock aquifers in El Paso County on water levels in the alluvial aquifer. The 50-yr projection period is subdivided into seven stress periods with durations of 1, 2, 3, 5, 8, 12, and 19 yrs (fig. 32). Three projection scenarios were simulated: (1) A baseline projection that simulated the effects of withdrawal from the alluvial and bedrock aquifers (layers 1, 2, 4, 6), approximately at estimated October 1989–September 1990 rates; (2) a 1-percent-depletion projection that simulated the effects of withdrawal from the alluvial aquifer (layer 1) at estimated October 1989–September 1990 rates and increased withdrawal from the bedrock aquifers in El Paso County at rates equal to 1 percent of their total storage within El Paso County; and (3) a 0.33-percent-depletion projection that simulated the effects of continued withdrawals from the alluvial aquifer (layer 1) at estimated October 1989–September 1990 rates and increased withdrawal from the bedrock aquifer at rates equal to 0.33 percent of storage within El Paso County. In the latter two projection scenarios, maximum simulated annual withdrawal from a cell in a bedrock aquifer (layers 2, 4, 6, or 8) equaled 1 percent or 0.33 percent of the available storage in the cell. Available storage was calculated as the product of area, aquifer (saturated) thickness, and average specific yield (table 2). In some instances, estimated October 1989–September 1990 withdrawal rates from the bedrock aquifers (layers 2, 4, 6, and 8) were already being simulated at rates larger than the 1-percent- or 0.33-percent-depletion rate, and then the larger rate was used for the projection simulation.

Although the simulated rates of withdrawal from the bedrock aquifers are much greater than likely would occur, they are based on the maximum rates that could be allowed under current regulations. It also is improbable that the rates of withdrawal from the bedrock aquifers would increase from current (1990) estimated rates to the maximum allowable rates of withdrawal in a short period. Withdrawals from the bedrock aquifers probably would increase gradually, with the bedrock aquifer nearest land surface being developed first. Therefore, the 1-percent and 0.33-percent-depletion scenarios can be considered to be worst-case scenarios.

Table 12. Mean error and mean square error in simulated heads for steady-state and transient-state conditions and for selected sensitivity tests of the model

[M.S.E., mean square error, in feet squared; NA, not applicable; Kz, vertical hydraulic conductivity; Ss, specific storage; --, valnes not computed]

Model layers with values changed for sensitivity test	Model value changed	Change[1] ratio	Layer number/hydrogeologic unit									
			Alluvial aquifer		Dawson aquifer		Denver aquifer		Arapahoe aquifer		Laramie-Fox Hills aquifer	
			Mean[2] error (foot)	M.S.E.[2] (foot squared)	Mean[2] error (foot)	M.S.E.[2] (foot squared)	Mean[2] error (foot)	M.S.E.[2] (foot squared)	Mean[2] error (foot)	M.S.E.[2] (foot squared)	Mean[2] error (foot)	M.S.E.[2] (foot squared)
Steady-state conditions												
NA	NA	1.0	–0.26	13.99	–0.01	14.08	2.49	67.97	–18.83	2118.	–189.	71,008.
Sensitivity tests for steady-state conditions												
1	Kz	0.1	–0.61	1.32	–0.56	1.11	-0.44	.52	–0.25	.12	.01	.00
2	Kz	10	.00	.00	–0.02	.01	.09	.04	.16	.06	.04	.00
2	Kz	100	.00	.00	–0.02	.01	.09	.04	.17	.07	.04	.00
4	Kz	10	–0.12	.78	–0.40	3.52	–0.87	4.3	.34	.41	.19	.04
4	Kz	100	.07	.78	–0.23	3.60	–0.82	4.54	.41	.47	.15	.03
6	Kz	10	–0.08	.15	–0.08	.13	–0.11	.12	.77	7.77	.08	.01
6	Kz	100	–0.11	.27	–0.11	.22	–0.14	.20	.95	11.22	–4.88	44.35
8	Kz	10	.03	.00	.01	.00	.01	.00	–0.02	.00	–4.00	29.74
8	Kz	100	.03	.01	.02	.00	.01	.00	–0.02	.00	–4.96	44.70
2–8	Kz	.1	–0.68	.73	1.26	41.49	2.22	52.57	–6.19	198.	8.83	204.40
2–8	Kz	10	–0.16	.90	–0.47	3.84	–0.87	4.66	1.29	11.27	–4.05	29.93
2–8	Kz	100	–0.26	.89	–0.48	1.56	–1.05	6.17	1.49	15.24	–5.01	45.20
Transient-state conditions												
NA	NA	1.0	.46	5.03	--	--	--	--	--	--	--	--
Sensitivity tests for transient-state conditions												
2, 4, 6, 8	Ss	.67	–0.80	106.37	--	--	--	--	--	--	--	--
2, 4, 6, 8	Ss	1.33	–0.80	106.12	--	--	--	--	--	--	--	--
3, 5, 7	Ss	.1	–0.81	106.39	--	--	--	--	--	--	--	--
3, 5, 7	Ss	10	–0.77	105.68	--	--	--	--	--	--	--	--
3, 5, 7	Ss	100	–0.75	105.45	--	--	--	--	--	--	--	--

[1]The change ratio is the ratio of the value used in the sensitivity test to that used in the model.

[2]Mean error and mean square error (M.S.E.) of simulated heads for the model of steady-state conditions were computed, assuming that the starting heads for layers 1, 2, 4, 6, and 8 represented the true steady-state heads. Mean error and mean square error of simulated heads from sensitivity tests of the model of steady-state conditions were computed, assuming that the simulated heads for steady-state conditions represented the true steady-state head distribution. Mean error and mean square error for simulated heads in layer 1 at the end of stress period 4 (spring 1964 transient-state conditions) were computed, assuming that the 1964 water-table surface (McGovern and Jenkins, 1966) represented the true heads in the alluvial aquifer.

During a simulation, a cell may be simulated as going dry (become unsaturated); flow to or from adjacent cells is not computed and, in effect, the dry cell acts as an internal flow boundary. When a cell went dry during a simulation, the version of MODFLOW used in this study converted the cell to a no-flow cell and treated the cell as a boundary and did not allow the cell to become saturated during the remainder of the simulation. Consequently, simulated water levels in adjacent cells and some budget terms may be affected. In reality, if part of an aquifer is temporarily desaturated, water will continue to flow toward the desaturated area from upgradient saturated areas and through the overlying unsaturated zone from infiltration of precipitation and surface water, thus, the desaturated area eventually may become saturated. This limitation of the model may result in anomalous values in simulated water levels, particularly in cells adjacent to cells which go dry.

During the baseline projection, eight cells in the alluvial aquifer (layer 1) went dry, resulting in increasing simulated water levels in the alluvial aquifer upgradient from the dry cell and decreasing simulated water levels downgradient from the dry cell. The simulated ground-water budgets for the alluvial aquifer (layer 1) for the projection scenarios are shown in figure 35. Simulated change (depletion) of storage in the alluvial aquifer for the baseline projection decreased from about 5.5 ft^3/s to about 1.5 ft^3/s during the 50-yr simulation period (fig. 35A). The simulated cumulative change of storage in the alluvial aquifer was about 81,000 acre-ft, which is equivalent to an average water-level decline of about 8.6 ft over the entire alluvial aquifer. Projected water-level decline in the alluvial aquifer at the end of the baseline projection is shown in figure 36A. Net simulated vertical flow from the bedrock into the alluvial aquifer averaged about 0.47 ft^3/s. Cumulative vertical flow from the bedrock into the alluvial aquifer was about 17,000 acre-ft. Estimated withdrawal rates from the alluvial aquifer decreased during the 50-yr projection from about 11.6 ft^3/s in stress period 1 to about 8.5 ft^3/s in stress periods 6 and 7, as a result of cells in the alluvial aquifer in which wells were simulated, as going dry during the projection period.

During the 1-percent-depletion projection, 78 cells in the alluvial aquifer (layer 1) went dry, which represents the equivalent of dewatering a 19.5-mi^2 area of the alluvial aquifer. Consequently, simulated water levels in adjacent cells are affected and some budget terms were affected. Evapotranspiration from the alluvial aquifer was simulated to increase after 31 yrs of decreasing (fig. 35B). In reality, the rate of evapotranspiration would probably continue to decrease because water levels would not increase upgradient of the dry cells as simulated. The simulated ground-water budget for the 1-percent-depletion projection, shown in figure 35B, indicates an average simulated change in depletion of storage in the alluvial aquifer (layer 1) of 5.5 ft^3/s for the 50-yr projection period. Cumulative change of storage was about 200,000 acre-ft, which is equivalent to an average water-level decline of about 21 ft over the entire aquifer. Projected water-level decline in the area at the end of the 1-percent-depletion projection is shown in figure 36B. The simulated cumulative change of storage in the alluvial aquifer for the 1-percent-depletion projection was about 109,000 acre-ft more than the amount of depletion of the alluvial aquifer projected for baseline conditions. Net simulated vertical flow out of the alluvial aquifer (layer 1) averaged about 5.8 ft^3/s. Cumulative net vertical flow out of the alluvial aquifer was about 209,000 acre-ft, which is a net change of about 226,000 acre-ft from baseline conditions. The model results indicate that net flow is from the bedrock to the alluvial aquifer under baseline conditions, but that the flow direction would reverse under the 1-percent-depletion projection conditions. This reversal was simulated to occur during the first year in which withdrawal from the bedrock equaled 1 percent of the available storage in El Paso County.

During the 0.33-percent-depletion projection, 47 cells in the alluvial aquifer (layer 1) went dry, which represents desaturation of a 11.75-mi^2 area of the alluvial aquifer. The effect of a cell simulated as dry for the 0.33-percent-depletion projection is similar to that described for the 1-percent-depletion projection, except the effect on the budget is not as pronouced (fig. 35C). The simulated ground-water budget for the 0.33-percent-depletion projection, shown in figure 35C, had an average simulated change of storage in the alluvial aquifer (layer 1) of about 4.0 ft^3/s. Cumulative change in storage simulated for the 0.33-percent-depletion projection was about 145,000 acre-ft, which is about 64,000 acre-ft more than the amount of depletion projected for baseline conditions. Net simulated vertical flow out of layer 1 averaged about 3.5 ft^3/s. Cumulative vertical flow out of the alluvial aquifer was projected at about 126,000 acre-ft, which is a net change of 143,000 acre-ft from baseline conditions. Under the 0.33-percent-depletion scenario, flow directions also are reversed within the first year in which withdrawals from the bedrock equal 0.33 percent of the available bedrock storage in El Paso County. Water-level declines in the alluvial aquifer at the end of the 50-yr projections are shown in figure 36.

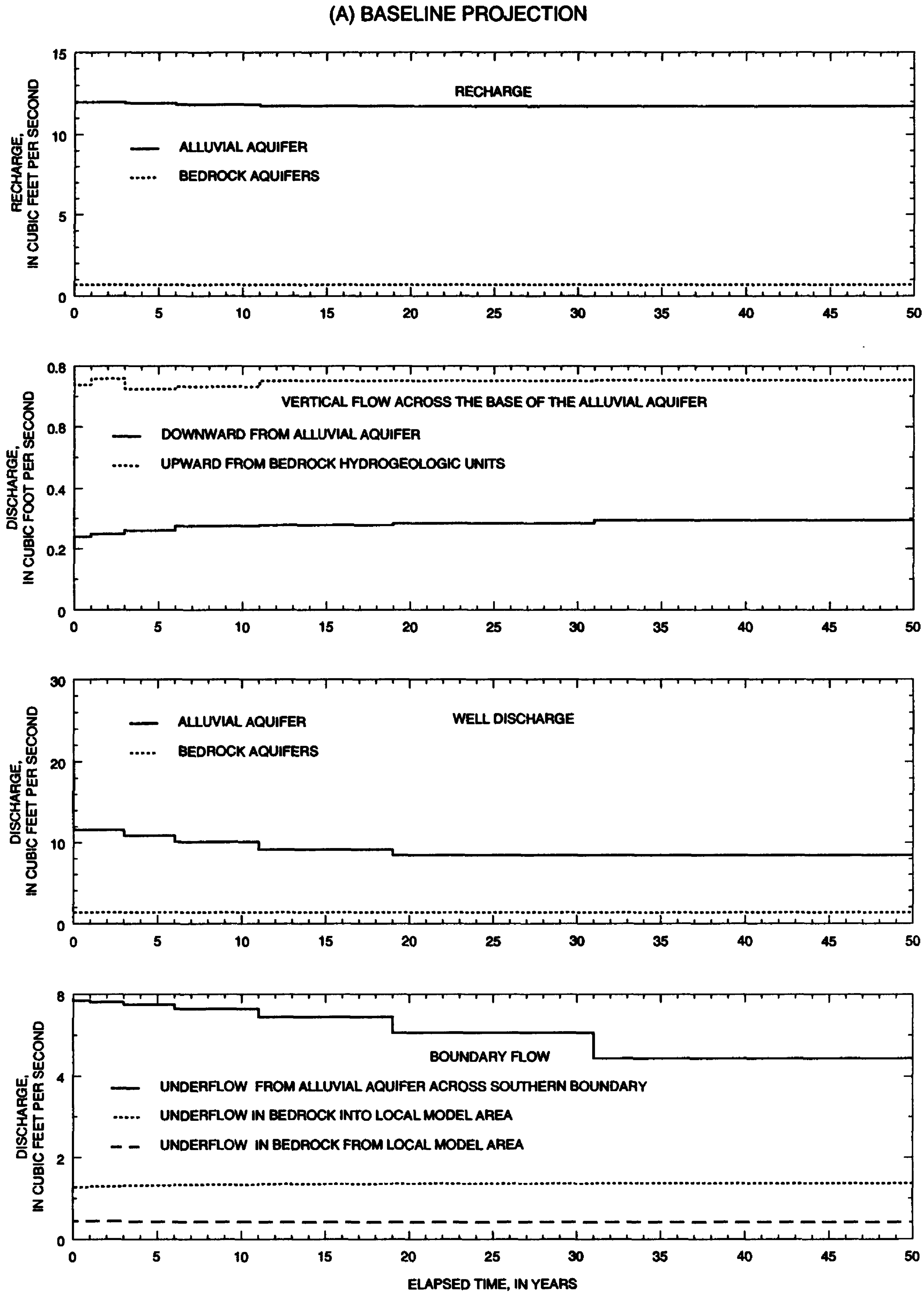


Figure 35. Simulated ground-water budgets for the alluvial aquifer and underlying bedrock in the upper Black Squirrel Creek Basin for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection.

(A) BASELINE PROJECTION

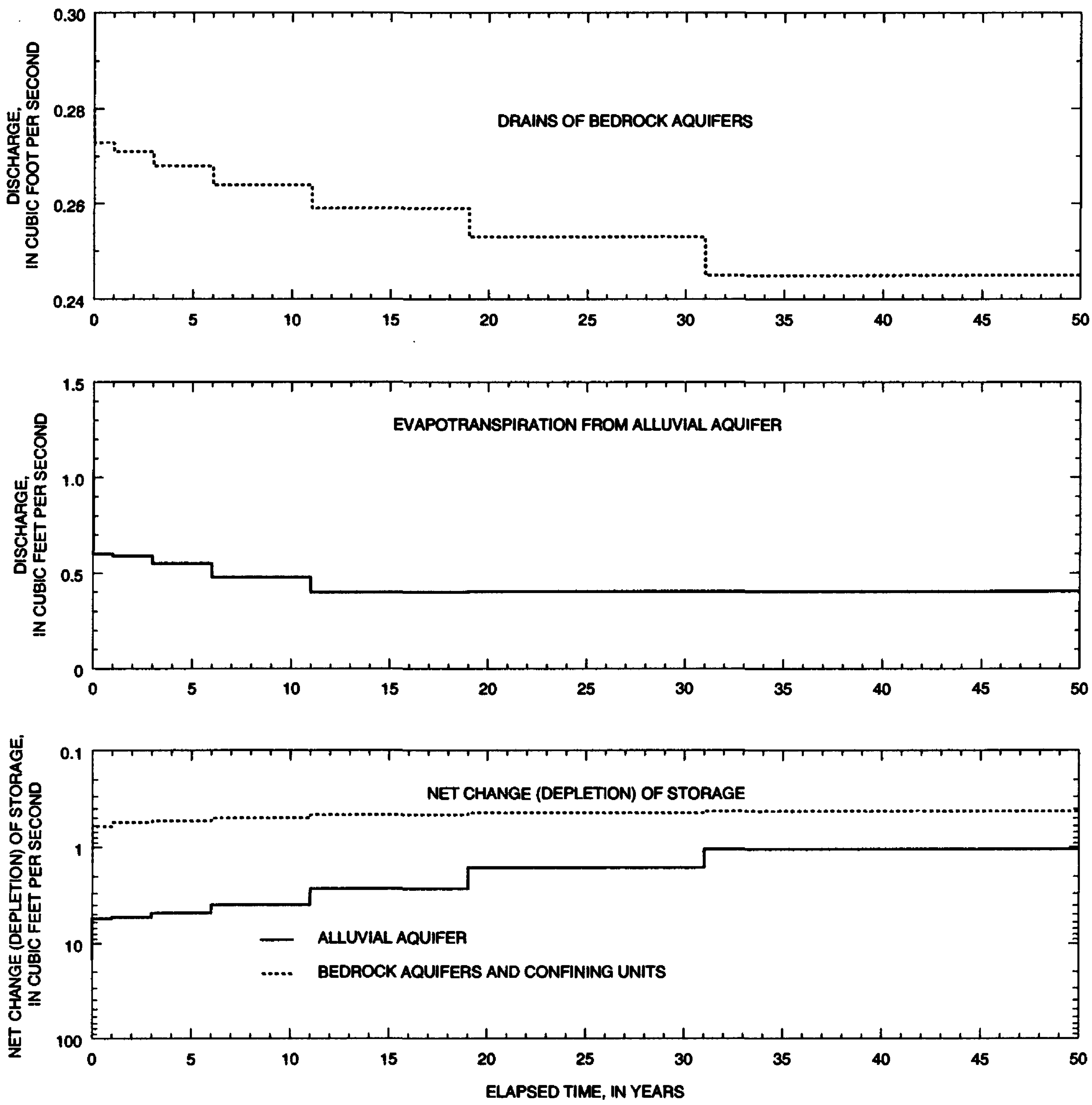


Figure 35. Simulated ground-water budgets for the alluvial aquifer and underlying bedrock in the upper Black Squirrel Creek Basin for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.

(B) 1-PERCENT-DEPLETION PROJECTION

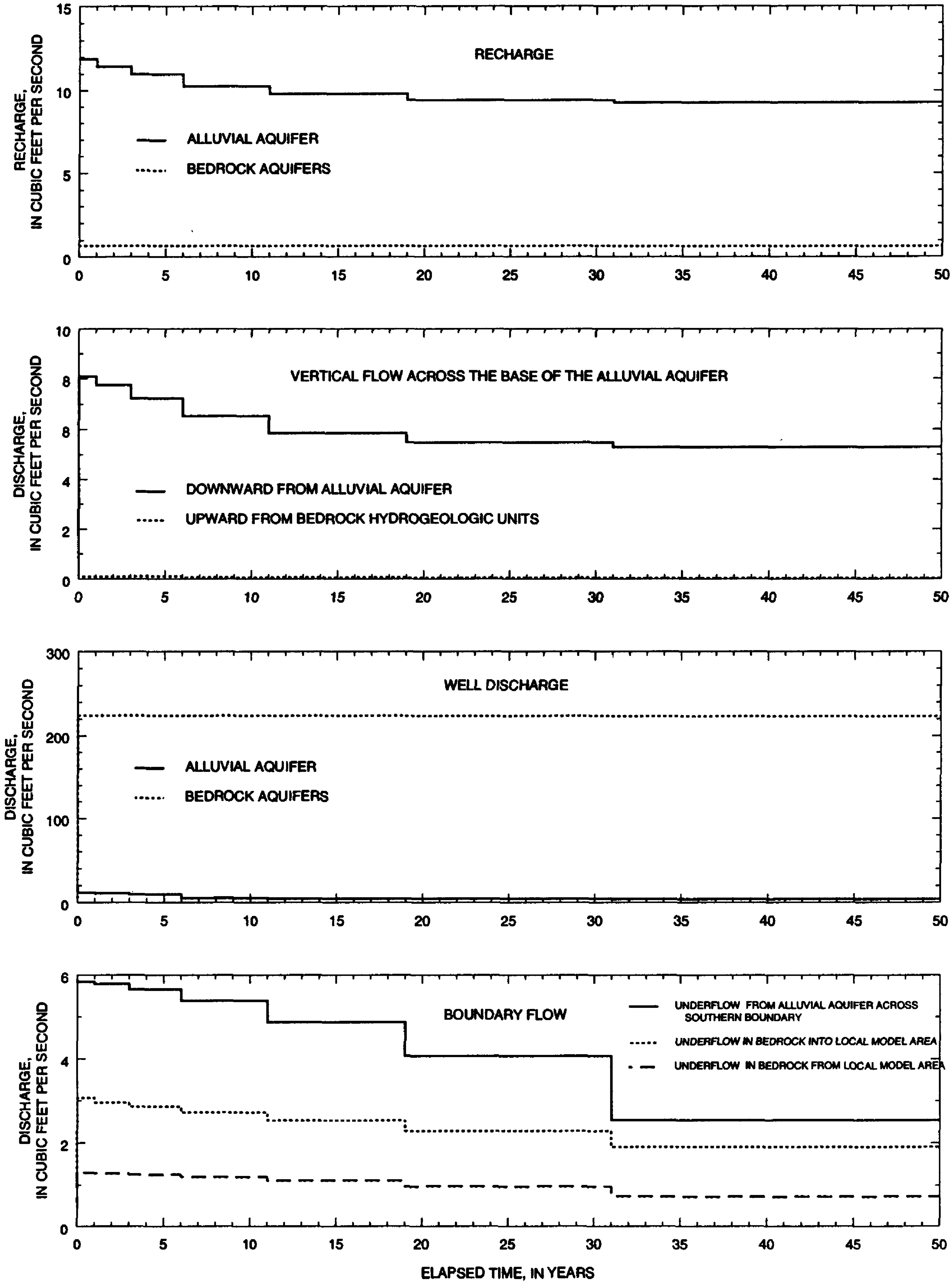


Figure 35. Simulated ground-water budgets for the alluvial aquifer and underlying bedrock in the upper Black Squirrel Creek Basin for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.

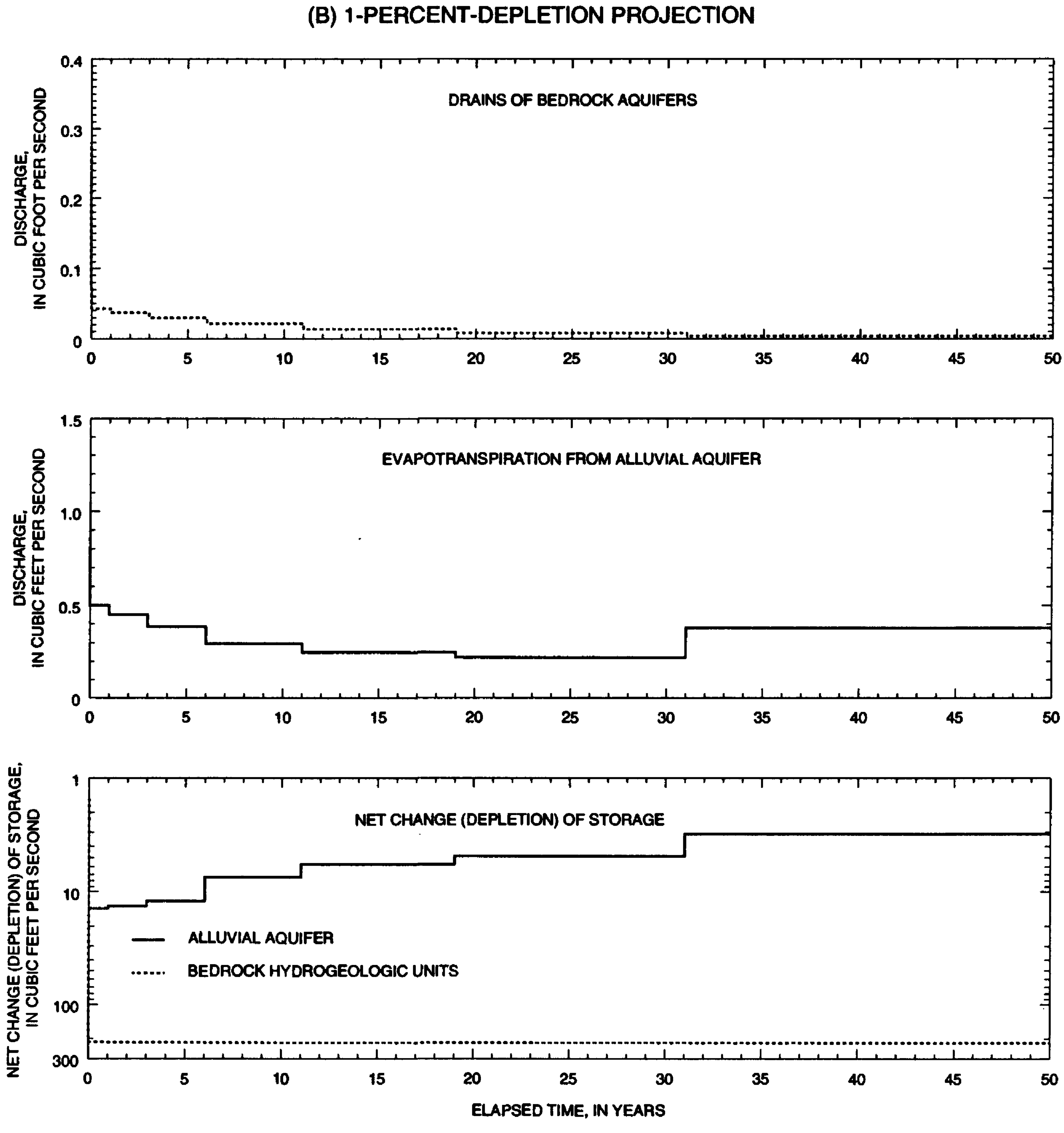


Figure 35. Simulated ground-water budgets for the alluvial aquifer and underlying bedrock in the upper Black Squirrel Creek Basin for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.

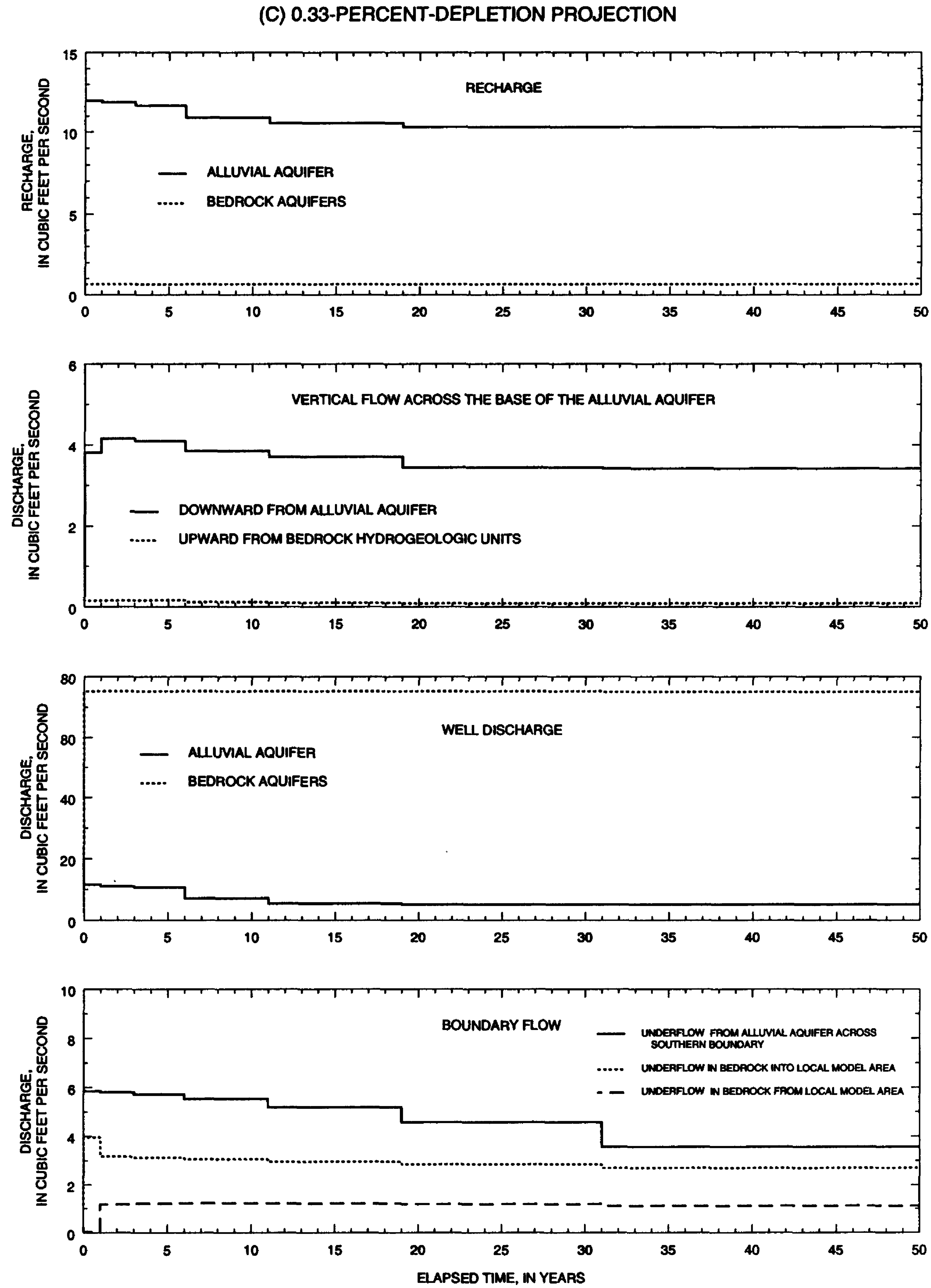


Figure 35. Simulated ground-water budgets for the alluvial aquifer and underlying bedrock in the upper Black Squirrel Creek Basin for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.

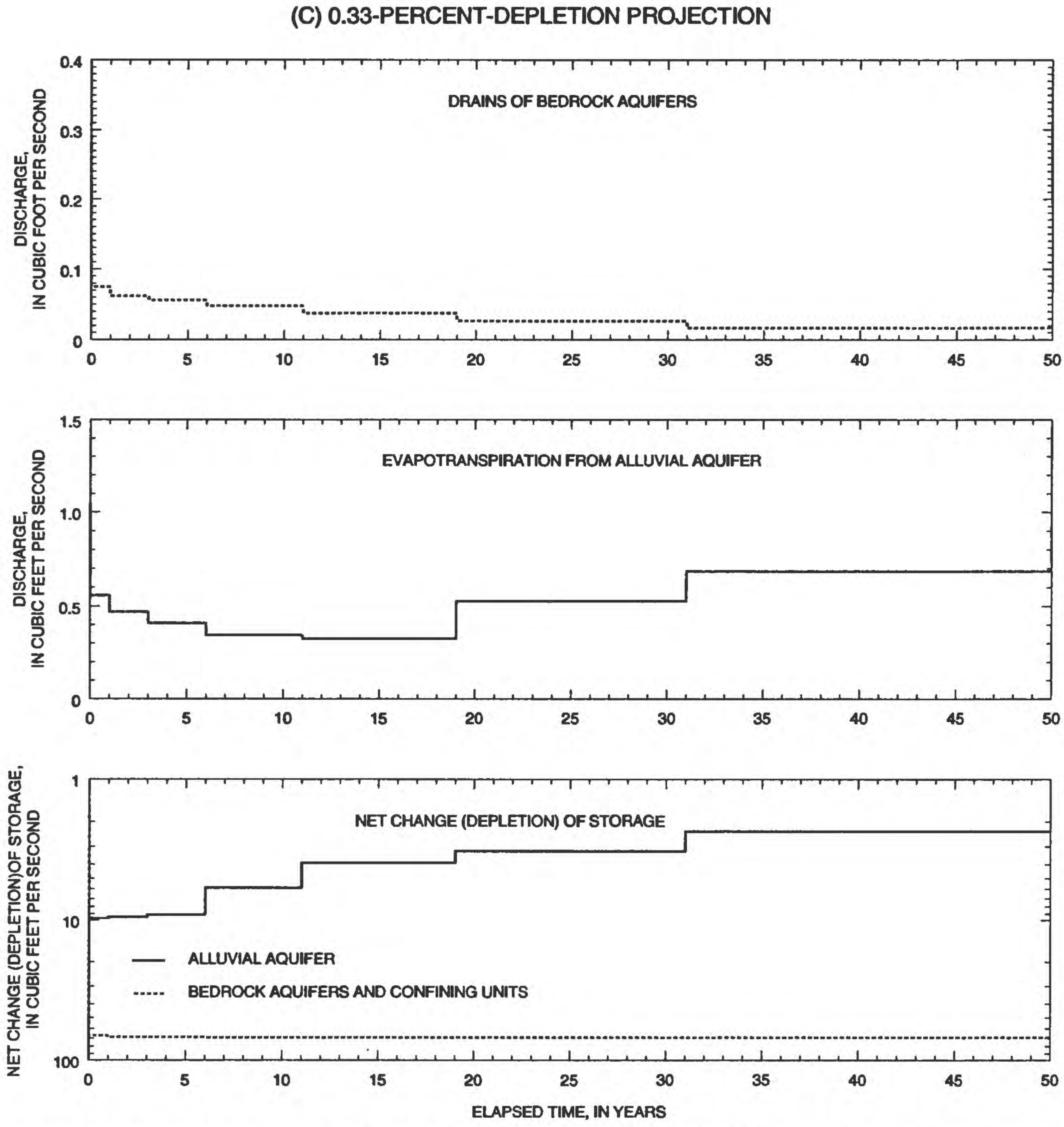


Figure 35. Simulated ground-water budgets for the alluvial aquifer and underlying bedrock in the upper Black Squirrel Creek Basin for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.

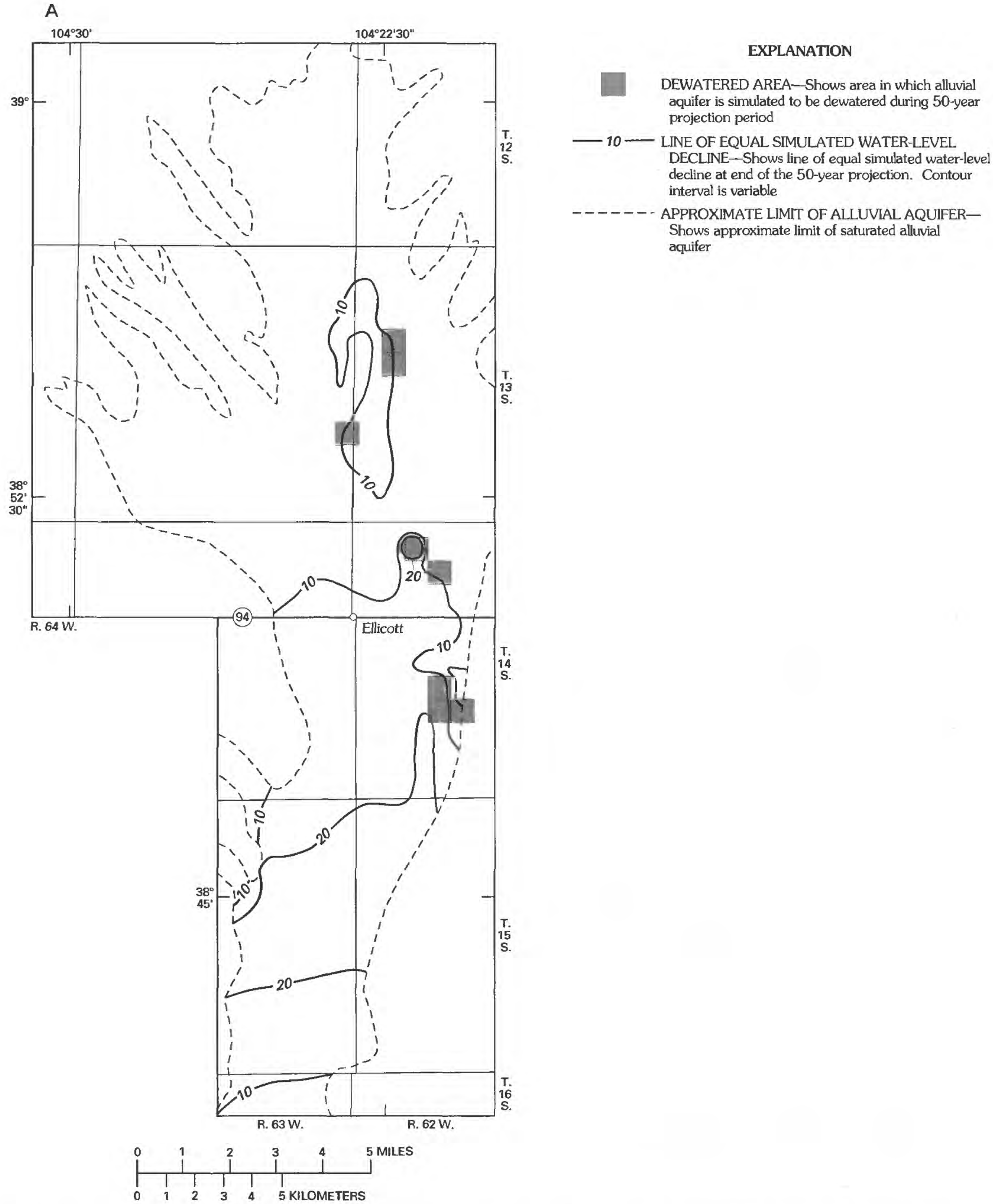


Figure 36. Simulated water-level declines in the alluvial aquifer after 50 years of withdrawal for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection.

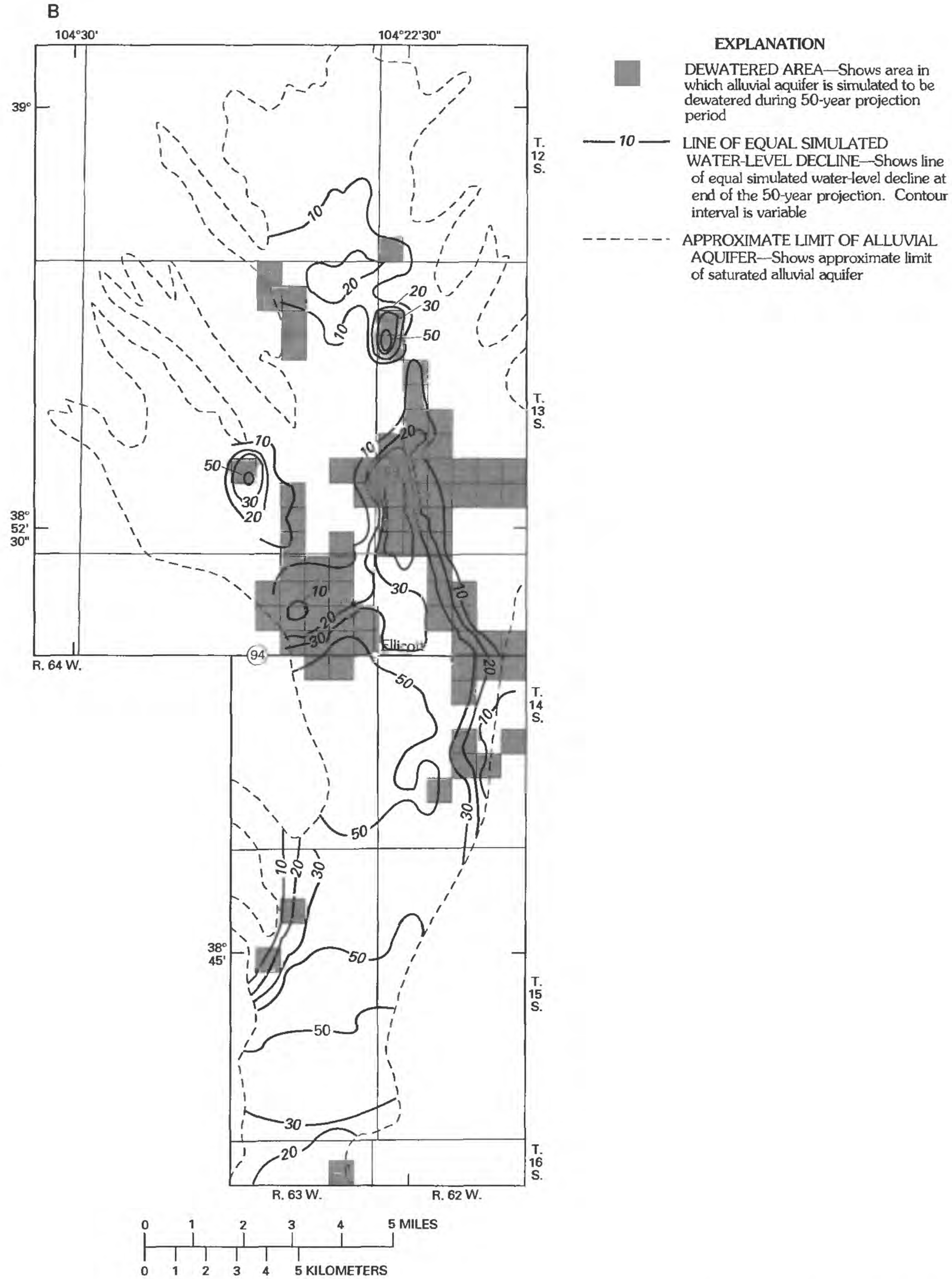
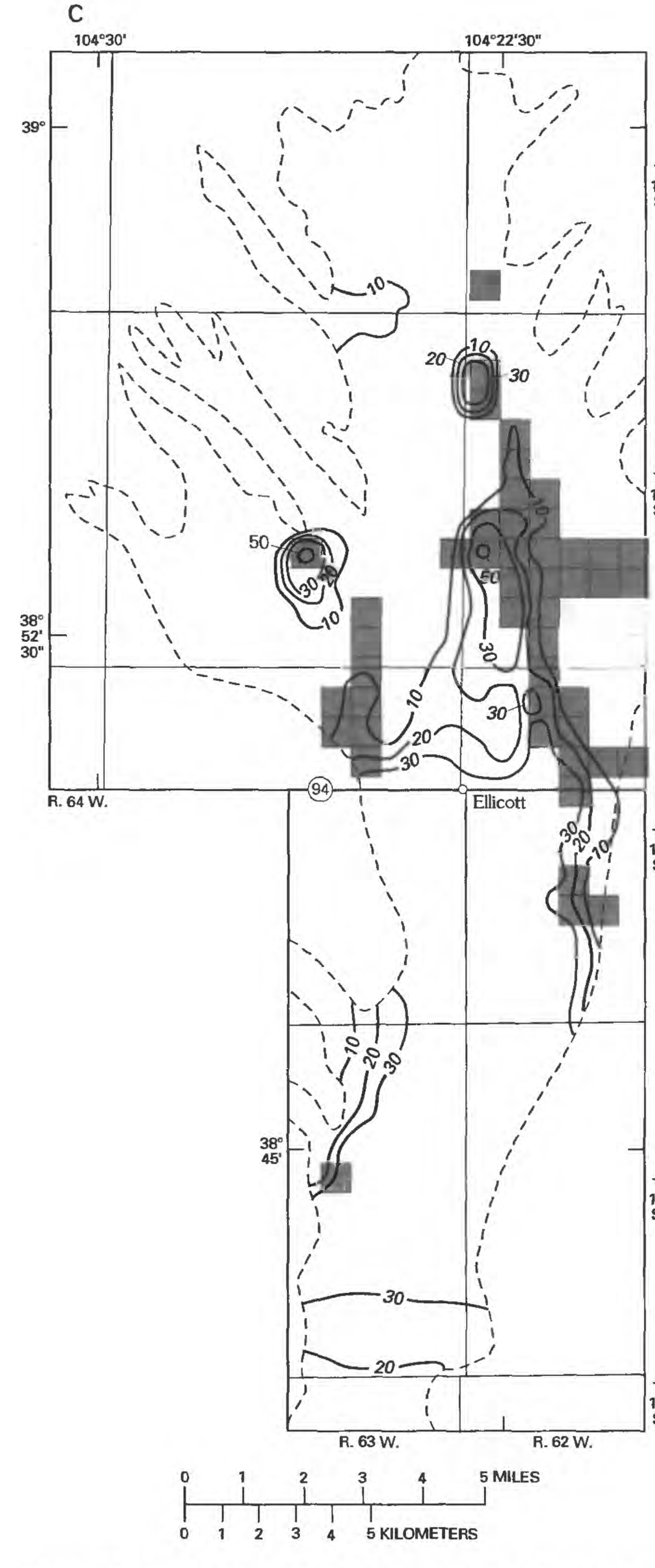


Figure 36. Simulated water-level declines in the alluvial aquifer after 50 years of withdrawal for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.



EXPLANATION

DEWATERED AREA—Shows area in which alluvial aquifer is simulated to be dewatered during 50-year projection period

—20— LINE OF EQUAL SIMULATED WATER-LEVEL DECLINE—Shows line of equal simulated water-level decline at end of the 50-year projection. Contour interval is variable

- - - - - - APPROXIMATE LIMIT OF ALLUVIAL AQUIFER—Shows approximate limit of saturated alluvial aquifer

Figure 36. Simulated water-level declines in the alluvial aquifer after 50 years of withdrawal for (A) the baseline projection, (B) the 1-percent-depletion projection, and (C) the 0.33-percent-depletion projection--Continued.

The water-level decline that resulted from continuation of October 1989 to September 1990 estimated rates of withdrawal from the alluvial aquifer caused desaturation of a smaller area (fig. 36A) than either the 1-percent-depletion projection (fig. 36B) or the 0.33-percent-depletion projection (fig 36C). Actual water-level declines that might occur in the future will differ from projected declines if withdrawal rates are substantially different from those simulated. Increases in withdrawals from the bedrock aquifers in El Paso County were simulated to: (1) Capture flow that currently discharges from the bedrock aquifers to springs and streams in upland areas and to the alluvial aquifer, (2) induce flow downward from the alluvial aquifer, and (3) accelerate the rate of water-level decline in the alluvial aquifer.

SUMMARY AND CONCLUSIONS

Water-level data, aquifer-test results, water-quality data, and model results indicate that the bedrock aquifers in the study area locally are hydraulically connected with the alluvial aquifer. During 1987–90, water-level hydrographs for some bedrock wells indicated that, locally, cyclic pumping from the alluvial aquifer caused temporary reversals in the direction of flow between the alluvial and bedrock aquifers. In general, the water-level data collected during this study indicate that the alluvial aquifer is recharged by flow from the Denver and Arapahoe aquifers, but water-level and water-quality data indicate that, locally, the Laramie-Fox Hills aquifer is recharged by flow from the alluvial aquifer.

Use of subsurface-temperature profiles to estimate specific discharge through leaky confining units was not found to be appropriate in the study area because: (1) Cyclic and seasonal pumping from the alluvial and the Laramie-Fox Hills aquifers cause large cyclic or seasonal fluctuations in water levels and; therefore, flow is not steady; (2) confining units generally are heterogeneous; and (3) horizontal flow in sandstones within the confining units affects the curvature of the temperature profiles.

The vertical hydraulic diffusivity of a 15-ft-thick interval of the lower Denver confining unit near well SC01306207BCB4 was determined by means of an aquifer test, based on the ratio method of Neuman and Witherspoon (1972), to be 1,000 ft^2/d. Specific storage of the leaky confining unit was an estimated 8×10^{-7} per ft, and estimated vertical hydraulic conductivity was 8×10^{-4} ft/d.

Similarities in ratios of major ions in water samples from the Laramie-Fox Hills and alluvial aquifers in the southern part of the study area locally indicate flow of water from the alluvial aquifer to the Laramie-Fox Hills aquifer. The relatively large concentration of dissolved nitrite plus nitrate, as nitrogen, in the Laramie-Fox Hills aquifer at SC01506313BBB3 also indicates flow from a near-surface source of nitrogen to the Laramie-Fox Hills aquifer.

The three-dimensional numerical model of flow for steady-state conditions indicated that the flow from the bedrock aquifers contributed about 7 percent of the total recharge to the alluvial aquifer. Net vertical flow into the alluvial aquifer from the bedrock was simulated to decrease by 0.32 ft^3/s, from a predevelopment (pre-1949) rate of 0.82 ft^3/s to 0.50 ft^3/s by April–September 1990.

The potential effects of future withdrawals from the bedrock aquifers were simulated for three hypothetical scenarios: (1) A continuation of October 1989–September 1990 estimates withdrawal rates from the alluvial and bedrock aquifers, (2) increased withdrawals from the bedrock aquifers equivalent to 1 percent per year of available storage, and (3) increased withdrawals from the bedrock aquifers equivalent to 0.33 percent per year of available storage. These projection scenarios, which simulated the withdrawals for a 50-yr period, indicated that, locally, the alluvial aquifer will be desaturated if future withdrawals from the alluvial and bedrock aquifers approximately equal the estimated October 1989–September 1990 rates. Furthermore, the projection simulations indicated that depletion of storage in the alluvial aquifer will accelerate if withdrawals from the bedrock aquifers in El Paso County increase substantially. At the end of the 50-yr projection period, withdrawals at the 1-percent-per-year depletion rate and at the 0.33-percent-per-year depletion rate from the bedrock aquifers in El Paso County were projected to cause cumulative depletion of storage in the alluvial aquifer of about 109,000 and 64,000 acre-ft more than occurred in the baseline projection.

At current rates of withdrawal, water levels in the alluvial aquifer in the upper Black Squirrel Creek Basin will continue to decline. If at some time in the future the water resources of the bedrock aquifers are developed and used, the increase in withdrawals from the bedrock aquifers could reverse the direction of flow between the alluvial and bedrock aquifers and accelerate depletion of storage in the alluvial aquifer. Because the numerical model of flow in the alluvial/bedrock aquifer system is based on limited field data in the study area, the vertical hydraulic properties of the bedrock aquifers and confining units are poorly defined, and future withdrawals from the alluvial and bedrock aquifers are unknown, future water-level changes in the alluvial and bedrock aquifers in the upper Black Squirrel Creek Basin may differ substantially from values predicted by the model.

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SYSTEM OF NUMBERING WELLS

The well locations in this report are given numbers based on the Bureau of Land Management system of land subdivision and show the location of the well by quadrant, township, range, section, and position within the section (fig. 37). The first letter "S" preceding the location number indicates that the well or spring is located in the area governed by the Sixth Principal Meridian. The second letter indicates the quadrant in which the well or spring is located. Four quadrants are formed by the intersection of the baseline and the principal meridian—"A" indicates the northeast quadrant, "B" the northwest, "C" the southwest, and "D" the southeast.

The first three digits of the number indicate the township; the next three digits, the range; and the last two digits, the section in which the well or spring is located. The letters following the section number locate the well or spring within the section. The first letter denotes the quarter section; the second, the quarter-quarter section; the third, the quarter-quarter-quarter section; and the fourth, the quarter-quarter-quarter-quarter section. The letters are assigned within the section in a counterclockwise direction, beginning with "A" in the northeast section and are assigned within each quarter-quarter section, in the same manner. Where two or more locations are within the smallest subdivision, consecutive numbers beginning with "1" are added in the order in which the data from the wells or springs were collected.

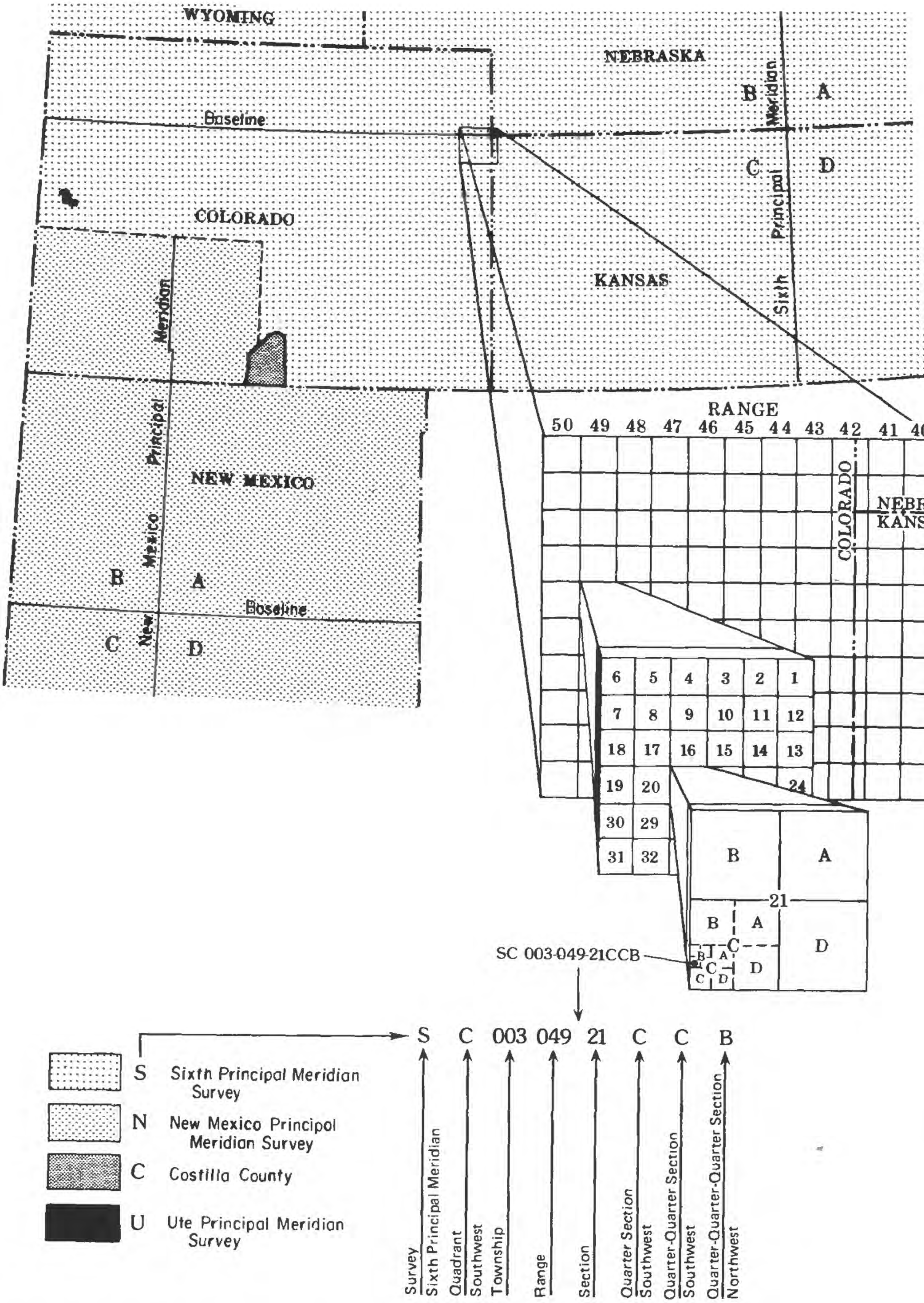

Figure 37. System of numbering wells.

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