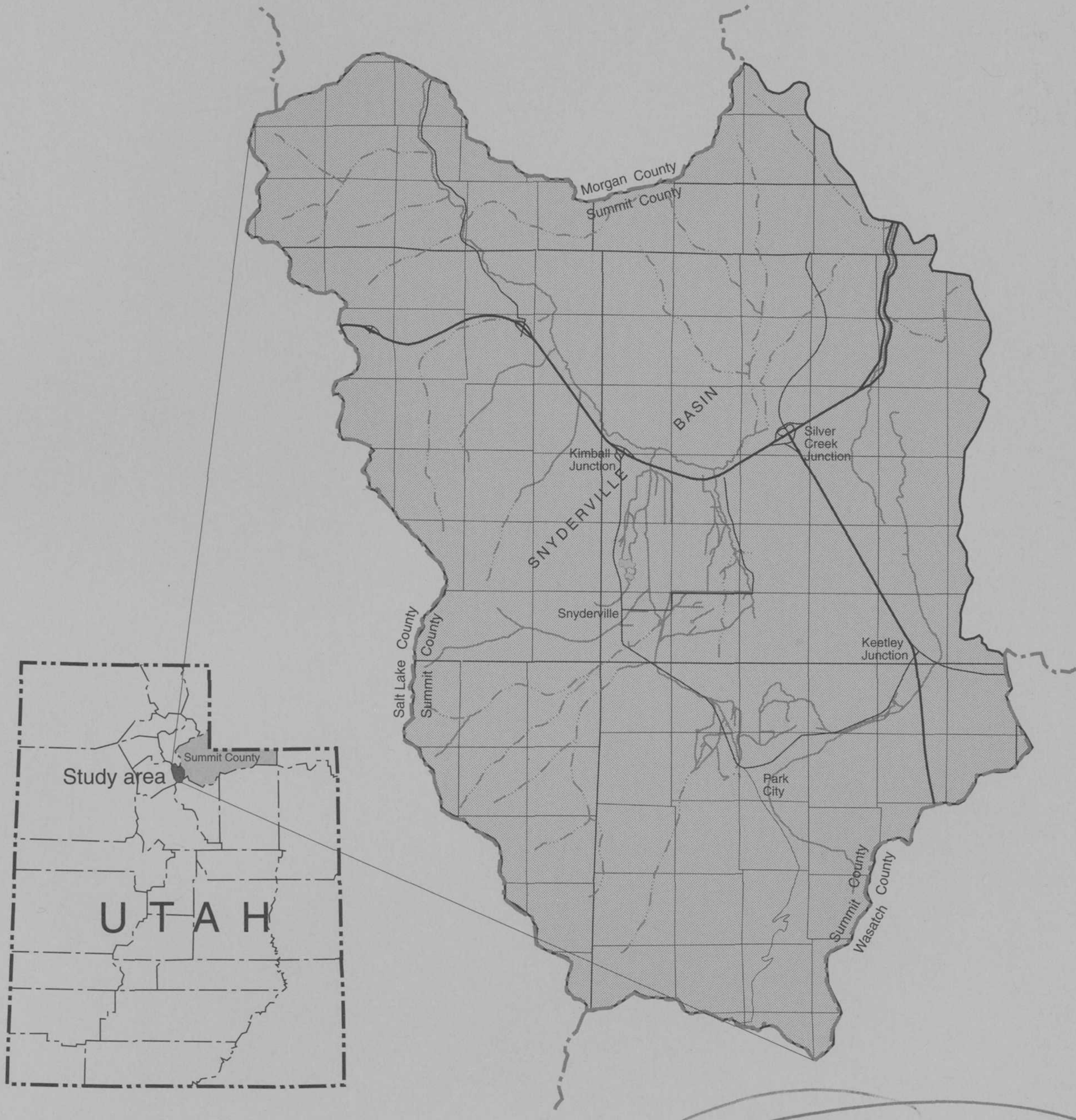
Hydrology and snowmelt simulation of Snyderville Basin, Park City, and adjacent areas, Summit County, Utah

Recd 6/30/98

Prepared by the U.S. Geological Survey



Technical Publication No. 115
State of Utah
DEPARTMENT OF NATURAL RESOURCES
1998

This report was prepared as a part of the Statewide cooperative water-resource investigation program administered jointly by the Utah Department of Natural Resources, Division of Water Rights and the U.S. Geological Survey. The program is conducted to meet the water administration and water-resource data needs of the State as well as the water information needs of many units of government and the general public.

Ted Stewart
Executive Director
Department of Natural Resources

Robert L. Morgan
State Engineer
Division of Water Rights

STATE OF UTAH
NATURAL RESOURCES
Division of Water Rights

Copies available at
Utah Department of Natural Resources
Bookstore
1594 West North Temple
Salt Lake City, Utah 84114

STATE OF UTAH
DEPARTMENT OF NATURAL RESOURCES

Technical Publication No. 115

HYDROLOGY AND SNOWMELT SIMULATION OF SNYDERVILLE BASIN, PARK CITY, AND ADJACENT AREAS, SUMMIT COUNTY, UTAH

By Lynette E. Brooks, James L. Mason, and David D. Susong

Prepared by the
United States Geological Survey
in cooperation with the
Utah Department of Natural Resources,
Division of Water Rights; Park City; Summit County;
and the Weber Basin Water Conservancy District
1998

The use of trade, product, industry, or firm names is for descriptive purposes only
and does not imply endorsement by the U.S. Government.

CONTENTS

PLATES

FIGURES

FIGURES—Continued

TABLES

CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS

Multiply	By	To obtain
acre	4,047	square meter
acre-foot (acre-ft)	1,233	cubic meter
	0.001233	cubic hectometer
acre-foot per year (acre-ft/yr)	0.00003907	cubic meter per second
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
foot per second (ft/s)	0.3048	meter per second
foot per month (ft/mo)	0.3048	meter per month
foot per year (ft/yr)	0.3048	meter per year
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	25.4	millimeter
mile (mi)	1.609	kilometer
mile per hour per foot ((mi/hr)/ft)	1.4665	meter per second per meter
pound per square foot (lb/ft^2)	4.8824	kilogram per square meter
pound per cubic foot (lb/ft^3)	16.0185	kilogram per cubic meter
square mile (mi^2)	2.590	square kilometer

Water temperature is reported in degrees Celsius (°C), which can be converted to degrees Fahrenheit (°F) by the following equation:

$$°F = 1.8(°C) + 32.$$

Snow temperature is reported in degrees Fahrenheit (°F), which can be converted to degrees Celsius (°C) by the following equation:

$$°C = 0.56(°F-32).$$

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

Chemical concentration is reported only in metric units. Chemical concentration is reported in milligrams per liter (mg/L) or micrograms per liter (µg/L). Milligrams per liter is a unit expressing the solute mass per unit volume (liter) of water. One thousand micrograms per liter is equivalent to 1 milligram per liter. For concentrations less than 7,000 milligrams per liter, the numerical value is about the same as for concentrations in parts per million. Specific conductance is reported in microsiemens per centimeter (µS/cm) at 25 degrees Celsius. Stable isotope concentration is reported as permil, which is equivalent to parts per thousand. Tritium concentration in water is reported as tritium units (TU). The ratio of 1 atom of tritium to 10^{18} atoms of hydrogen is equal to 1 TU or 3.2 picocuries per liter (pCi/L).

Vapor pressure is reported only in metric units as pascals (Pa). One pascal is 1 kilogram per meter per second squared.

HYDROLOGY AND SNOWMELT SIMULATION OF SNYDERVILLE BASIN, PARK CITY, AND ADJACENT AREAS, SUMMIT COUNTY, UTAH

By Lynette E. Brooks, James L. Mason, and David D. Susong
U.S. Geological Survey

ABSTRACT

Increasing residential and commercial development is placing increased demands on the ground- and surface-water resources of Snyderville Basin, Park City, and adjacent areas in the southwestern corner of Summit County, Utah. Data collected during 1993-95 were used to assess the quantity and quality of the water resources in the study area.

Ground water within the study area is present in consolidated rocks and unconsolidated valley fill. The complex geology makes it difficult to determine the degree of hydraulic connection between different blocks of consolidated rocks. Increased ground-water withdrawal during 1983-95 generally has not affected ground-water levels. Ground-water withdrawal in some areas, however, caused seasonal fluctuations and a decline in ground-water levels from 1994 to 1995, despite greater-than-normal recharge in the spring of 1995.

Ground water generally has a dissolved-solids concentration that ranges from 200 to 600 mg/L. Higher sulfate concentrations in water from wells and springs near Park City and in McLeod Creek and East Canyon Creek than in other parts of the study area are the result of mixing with water that discharges from the Spiro Tunnel. The presence of chloride in water from wells and springs near Park City and in streams and wells near Interstate Highway 80 is probably caused by the dissolution of applied road salt. Chlorofluorocarbon analyses indicate that even though water levels rise within a few weeks of snowmelt, the water took 15 to 40 years to move from areas of recharge to areas of discharge.

Water budgets for the entire study area and for six subbasins were developed to better understand the hydrologic system. Ground-water recharge from precipitation made up about 80 percent of the ground-water recharge in the study area. Ground-water discharge to streams made up about 40 percent of the surface water in the study area and ground-water discharge to springs and mine tunnels made up about 25 percent. Increasing use of ground water has the potential to decrease discharge to streams and affect both the amount and quality of surface water in the study area. A comparison of the 1995 to 1994 water budgets emphasizes that the hydrologic system in the study area is very dependent upon the amount of annual precipitation. Although precipitation on the study area was much greater in 1995 than in 1994, most of the additional water resulted in additional streamflow and spring discharge that flows out of the study area. Ground-water levels and ground-water discharge are dependent upon annual precipitation and can vary substantially from year to year.

Snowmelt runoff was simulated to assist in estimating ground-water recharge to consolidated rock and unconsolidated valley fill. A topographically distributed snowmelt model controlled by independent inputs of net radiation, meteorological parameters, and snowcover properties was used to calculate the energy and mass balance of the snowcover.

INTRODUCTION

The study area is in the southwestern corner of Summit County and includes all of the East Canyon Creek drainage within the county and the Silver Creek drainage from its headwaters to Tollgate Canyon, as shown in figure 1. This area includes the valley generally south of and straddling Interstate Highway 80 through which East Canyon Creek flows (Snyderville Basin), the area around Park City, and the area from south of Keetley Junction to north of Silver Creek Junction.

Population in this area has significantly increased from 1980 through the present (1998), and much of this

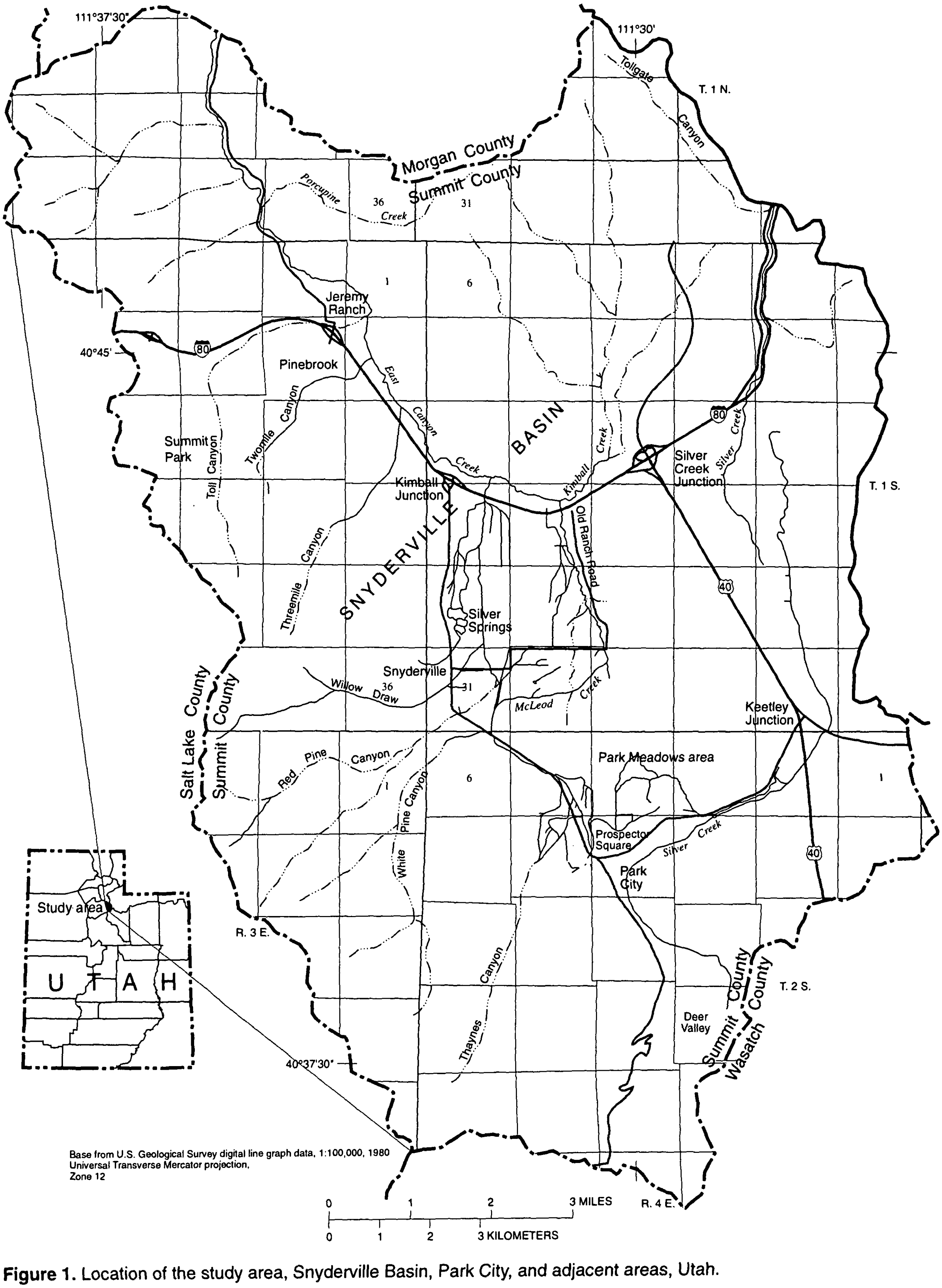


Figure 1. Location of the study area, Snyderville Basin, Park City, and adjacent areas, Utah.

increase has occurred after 1987. Industrial and commercial development in the area has increased, and ski areas are adding to their snow-making operations. Several venues for the 2002 Winter Olympics have been constructed in the study area. In 1990, the population of Summit County was 15,500 (Bureau of Economic and Business Research, University of Utah, written commun., 1994), an increase of 52 percent since 1980, and an increase of 173 percent since 1960. The approximate population of the study area in 1994 was 10,000. Retail trades and services are the businesses that employ the largest number of people in the study area.

This growth has placed increased demands on the ground- and surface-water resources in the area. One of the major constraints on development of additional residential areas and commercial activities is water supply. Surface water in the study area is considered to be fully appropriated. Because surface and ground water are interrelated, ground water also is considered to be fully appropriated. One method to obtain water rights for new development is to lease surface water stored in reservoirs, develop an equivalent amount of ground water, and release the surface water from the reservoirs to satisfy downstream users. There are concerns about how increased withdrawal of ground water might affect water levels in existing wells and springs, surface-water flows, and water quality within the study area and about how this development might affect surface-water flows and water quality downstream. Water-resource planners and agencies with water-management responsibilities need information and methods to manage existing water resources and to predict the effects of ground-water development. The U.S. Geological Survey, in cooperation with the Utah Department of Natural Resources, Division of Water Rights; Park City; Summit County; and the Weber Basin Water Conservancy District, completed a study of the water resources of the area from July 1993 to September 1997.

The quality of the water in the Park City area is good, but could degrade because of developing industry or if services such as centralized sewage treatment are not able to keep up with residential growth. Some of the water from mine tunnels contains concentrations of iron, manganese, sulfate, zinc, and dissolved solids that, before treatment or mixing, exceed State drinking-water standards. In September 1993, two public water suppliers indicated that sulfate concentrations are increasing in water being withdrawn from wells. The reason certain dissolved constituents are selectively increasing is not known, but the degradation of the ground water could be related to changes in the hydrologic system as a result of past development.

The objectives of the study were to define the geometry and character of the main aquifers, define how the hydrologic system works and how the different components interact, assess the existing quality of the water and the potential for degrading the quality, and provide data, analyses, and tools by which the effects of future development of water on the hydrologic system can be estimated. Specific objectives of the study included:

1. Define and describe the lateral and vertical extent of the principal aquifers in the area.
2. Describe the hydrologic system including the hydrologic properties of the aquifer; surface-water discharge, variability, and use; and ground-water recharge, direction of movement, discharge, storage, and use.
3. Improve available water budgets of the area.
4. Define and describe the interaction of ground water and surface water.
5. Describe the chemical quality of the surface water and the ground water, and identify potential sources of degradation.
6. Describe the hydrologic and hydrochemical effects that could be expected to occur as development creates a growing need for more water; for example, the effects on streamflow, water levels in wells, discharge of springs, quality of water, capture of natural discharge, and the quality and quantity of mine discharge.

The study was divided into two phases. The first phase was a 2-year data-collection period that was concurrent with an assessment of the geologic framework completed by the Utah Geological Survey. The second phase of the study was the synthesis and interpretation of the hydrologic and geologic data.

In the first phase, hydrologic data were collected during 1993-95 to provide a better understanding of the hydrologic system. Data-collection activities included the establishment of two surface-water gaging stations to complement four existing gaging stations; instantaneous measurement of streamflow for seepage analysis; measurement of ground-water discharge from springs; measurement of water levels in wells; and surface- and ground-water sampling for chemical analysis of major ions, isotopes, and chlorofluorocarbons (Downhour and Brooks, 1996). A snow data-collection site was operated during the winter months of 1993-95. Additional data were obtained from municipalities, water companies, the cooperative water-use program, and other climatic and snow-survey sites. These data included additional water-level measurements in wells, discharge from wells and springs, and miscellaneous streamflow measurements. The location of selected hydrologic-data sites is shown on plate 1. The numbering system for hydrologic-data sites in Utah is shown in figure 2.

In the second phase, hydrologic, chemical, and geologic data were used to determine water budgets for the study area and for six subbasins that were delineated on the basis of topography. Computer simulations of precipitation and snowmelt were done to determine the amount of water available for recharge and runoff in each subbasin. Where feasible, ground-water levels and geochemical modeling techniques were used to delineate probable direction of ground-water flow.

Purpose and Scope

This report describes the hydrologic system and documents the quantity and quality of water resources in Snyderville Basin, Park City, and adjacent areas. The report is based on the most recent interpretation of hydrologic data and geology and will add to the understanding of the hydrologic system and assist planners in assessing the effects of increased development on surface-water flows, ground-water levels, spring discharge, and the quality of the area's water resources. This report also indicates where additional ground-water monitoring would help to determine the extent of these effects. The results of this study provide a basis for comparison from which possible future changes to the hydrologic system can be identified. Information summarized in this report includes climatic data; surface-water flow; water levels in wells; discharge from springs, wells, and mine tunnels; and water-quality data.

Acknowledgments

Special acknowledgment is extended to the residents of Snyderville Basin, Park City, and adjacent areas, and to the officials of water companies for the information and access to sites they provided. Rich Hilbert of the Park City Water Department provided valuable data regarding flows from the Spiro Tunnel, Dority Springs, and other springs and streams in Park City. John Bollwinkel of Community Water Company provided his time and snowmobile to help measure water levels and collect samples for water-quality analysis. Cooperation from the Utah Department of Environmental Quality, Division of Drinking Water, is appreciated.

Special appreciation is given to Spring Creek Angus Ranch, Kimball Junction Properties, SCSC Incorporated, Spring Creek Associates, and Timberline Special Service District for their support to the Utah Department of Natural Resources, Division of Water Rights.

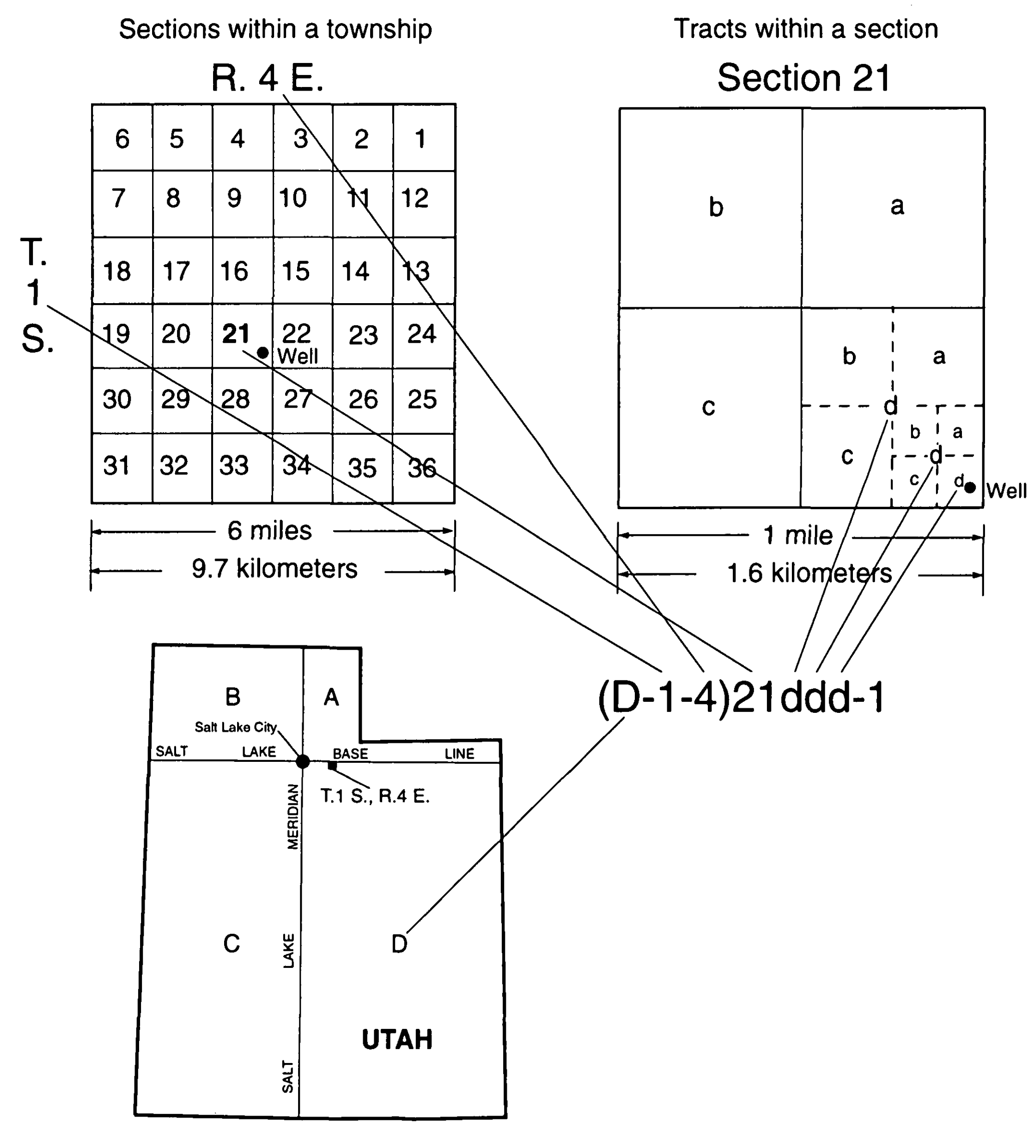
Previous Investigations

The first hydrologic investigation of the area was completed by Baker (1970) as part of a reconnaissance that assessed the water resources of the Heber, Kamas, and Park City areas. Thompson (1983) completed a study of the quality of surface water in the Weber River basin. This study included sampling of surface water for chemical analysis from Silver Creek and East Canyon Creek within the present study area. Holmes and others (1986) completed a detailed water-resources assessment of the Park City area, which included a part of the Provo River drainage adjacent to the study area to the south. Primary emphasis of Holmes and others (1986) was an analysis of the ground-water budget in the unconsolidated valley fill and consolidated rock. Mason (1989) completed a site-specific hydrologic and chemical assessment of the Prospector Square area in Park City. Solute and isotopic chemistry were used by Mayo and others (1992) to identify flow paths in the Wasatch Range, including the Park City area, and their relation to acid mine drainage. Additional site-specific information of the area is described in numerous well-head-protection studies submitted to the Utah Department of Health, Division of Environmental Quality.

Physiography

The study area lies within the Middle Rocky Mountain physiographic province (Fenneman, 1931). Altitude ranges from about 6,100 ft on the northern border where East Canyon Creek exits the study area to about 10,100 ft at the topographic divide on the southeastern boundary of the study area. The study area contains a low topographic divide that trends in a northerly direction from Park City to the intersection of Interstate Highway 80 and U.S. Highway 40. Surface water west of the divide is part of the East Canyon Creek drainage

The system of numbering wells and springs in Utah is based on the cadastral land-survey system of the U.S. Government. The number, in addition to designating the well or spring, describes its position in the land net. The land-survey system divides the State into four quadrants separated by the Salt Lake Base Line and the Salt Lake Meridian. These quadrants are designated by the uppercase letters A, B, C, and D, indicating the northeast, northwest, southwest, and southeast quadrants, respectively. Numbers designating the township and range, in that order, follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section and is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section—generally 10 acres for a regular section[1]. The lowercase letters a, b, c, and d indicate, respectively, the northeast, northwest, southwest, and southeast quarters of each subdivision. The number after the letters is the serial number of the well or spring within the 10-acre tract. When the serial number is not preceded by a letter, the number designates a well. When the serial number is preceded by an "S," the number designates a spring. A number having all three quarter designations but no serial number indicates a miscellaneous data site other than a well or spring, such as a location for a surface-water measurement site or tunnel portal. Thus, (D-1-4)21ddd-1 designates the first well constructed or visited in the southeast 1/4 of the southeast 1/4 of the southeast 1/4 of section 21, T. 1 S., R. 4 E.



[1]Although the basic land unit, the section, is theoretically 1 square mile, many sections are irregular in size and shape. Such sections are subdivided into 10-acre tracts, generally beginning at the southeast corner, and the surplus or shortage is taken up in the tracts along the north and west sides of the section.

Figure 2. Numbering system used for hydrologic-data sites in Utah.

and surface water east of the divide is part of the Silver Creek drainage.

Climate

Normal annual precipitation (1961-90) in the study area varies from about 19 in. at lower altitudes to 44 in. at higher altitudes (Utah Climate Center, 1996). About 65 percent of lower-altitude precipitation and 75 percent of higher-altitude precipitation occurs during the winter months (October-April). The altitude of the eastern part of the study area is about 2,500 ft higher than that of Salt Lake City, Utah, 25 mi to the northwest, but normal annual precipitation is only 3 in. greater than that in Salt Lake City. These data indicate that most of the study area is in a rain shadow of the mountains along the western edge of the study area.

Although no long-term weather station is located within the study area, the precipitation recorded at the station located at Silver Lake in Brighton, Utah, about 5 mi southwest of Park City, is representative of higher-altitude precipitation within the study area. Climatic data have been collected at this site since 1931, and 1961-90 normal annual and monthly precipitation has been calculated. Precipitation has been measured since July 1987 at the Thaynes Canyon snow-survey (SNOTEL) site in the study area (pl. 1) by the Natural Resources Conservation Service (formerly Soil Conservation Service). Monthly precipitation at Thaynes Canyon was compared to corresponding monthly precipitation at Silver Lake. Monthly precipitation at both sites has a high degree of correlation, with a correlation coefficient of 0.95. On the basis of this linear regression, normal monthly precipitation was calculated for the Thaynes Canyon site. Monthly precipitation and the departure from the calculated normal monthly precipitation for the Thaynes Canyon site are shown in figure 3.

All hydrologic estimates and water budgets presented in this report are calculated on the basis of a water year[1]. Precipitation for the 1994 water year was 32.9 in., 6.7 in. (17 percent) less than the calculated

[1] A water year is the 12-month period beginning October 1 and ending September 30 in the following year. The water year is designated by the calendar year in which it ends.

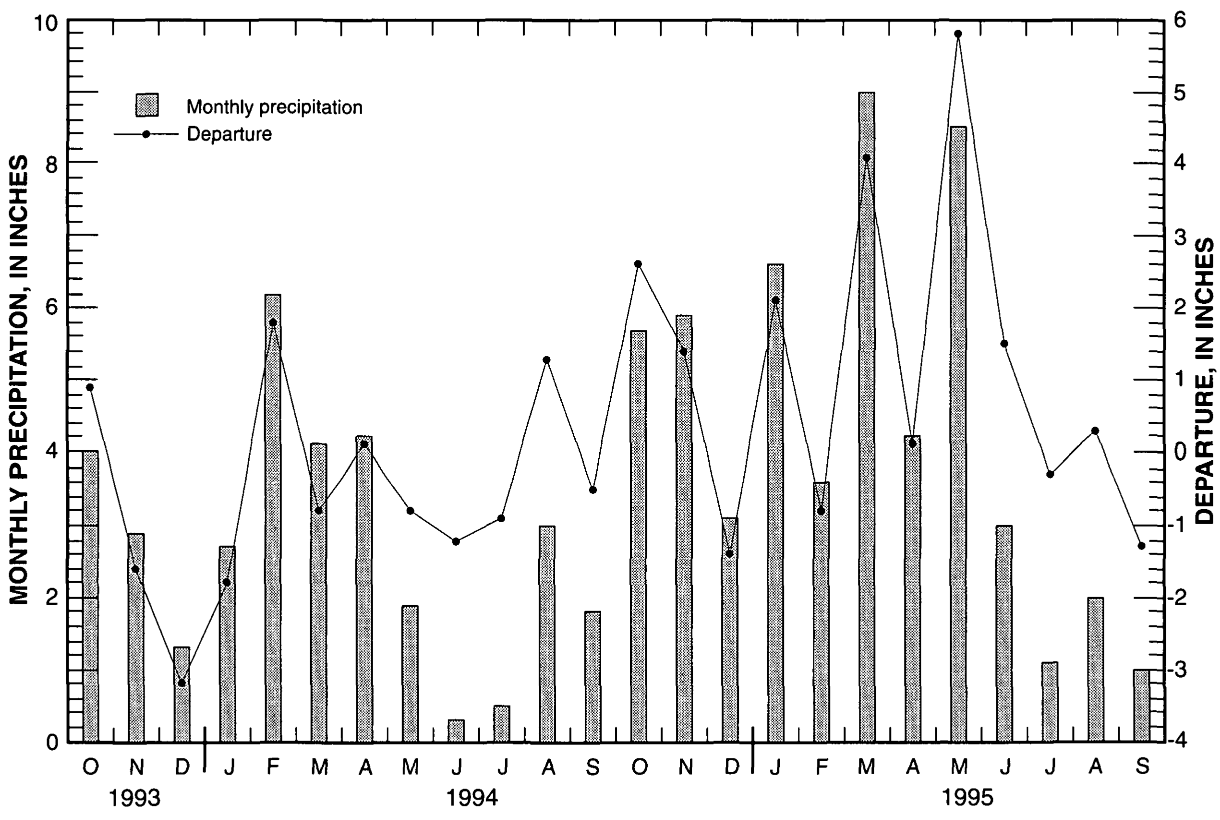


Figure 3. Monthly precipitation and departure from calculated normal monthly precipitation at Thaynes Canyon near Park City, Utah.

normal precipitation of 39.6 in. at the Thaynes Canyon site. Precipitation for the 1995 water year was 53.7 in., 14.1 in. (36 percent) more than the calculated normal precipitation. Precipitation during the 1995 water year was 20.8 in. more than during the 1994 water year. These comparisons indicate that the amount of water available to the study area is closer to normal for the 1994 water year than for the 1995 water year.

Compared with other areas of Utah, temperatures in the study area are mild during the summer months and cold during the winter months. Typically, summer maximum temperatures are below 90 °F and winter minimum temperatures are below 0 °F.

Land Use and Vegetation

The study area includes about 65,000 acres, most of which is undeveloped land with natural vegetation. Agricultural land is irrigated pasture, nonirrigated and wet pasture, and irrigated alfalfa. Ninety-eight percent of the irrigated land is irrigated with surface water, and some of the land is irrigated only when ephemeral and intermittent streams are flowing. Much of the pasture area is naturally irrigated by ground water. Land use, vegetation type, acres of each type, estimated water use of each type, and references for water use are listed in table 1. Areas of residential and commercial development, irrigated crops, nonirrigated pasture, and riparian vegetation were determined from a digital land use map (Utah Department of Natural Resources, Division of Water Resources, 1992). Areas of natural vegetation were determined from a digital Gap Analysis map (Utah Cooperative Fish and Wildlife Research Unit, 1995).

Water Use

Water demand in the study area is increasing as residential development continues, and as a result, hydrologic flow regimes in the study area might change. Water use for public supply was 1,300 acre-ft in 1980 (Utah Department of Natural Resources, Division of Water Rights, 1982, table 2) and 4,100 acre-ft in 1990 (Utah Department of Natural Resources, Division of Water Rights, 1993, p. 16). The amount of water used from wells, mine tunnels, springs, and surface water for public supply and domestic use in 1994 and 1995 is listed in the following table. In 1995, about 600 acre-ft of municipal water was consumed by evapotranspiration from lawns and gardens and about 600 acre-ft of municipal water recharged the ground-water system through irrigation of lawns and gardens (see "Methods" section later in this report). Areas such as Park City and Silver Springs use a higher percentage of water for lawn and garden irrigation than do areas such as Summit Park and Pinebrook, which have more natural vegetation.

Source of water	Water use (acre-feet) 1994 water year	1995 water year
Wells	2,800	2,400
Mine tunnels	2,400	2,300
Springs	1,400	1,700
Surface water	100	100
Total	**6,700**	**6,500**

Possible future increased treatment and use of mine-tunnel water and use of spring water for public supply would reduce the amount of streamflow exiting the study area. About 50 percent of the water used for public supply presently re-enters the streams through waste-water treatment plants. Only about 10 percent of the water is consumed (see "Water-budget analysis" section of this report). In areas where a higher percentage of water is being used to irrigate lawns and gardens, the percentage re-entering streams is less because water is lost to evapotranspiration. Natural streamflow also might be reduced by increased ground-water withdrawals for public supply. Increased ground-water withdrawals might cause seasonal and local water-level declines, which could decrease ground-water discharge to streams and springs. Because much of the increased ground-water withdrawal has occurred since 1990, the long-term effects on water levels and surface water are not yet known.

As undeveloped or agricultural land becomes urban and residential land, ground water and surface water will be affected. In areas of high ground-water levels, such as near Park Meadows and in Snyderville Basin, ground-water discharge to streams may increase as a result of urban or residential development. In these areas, ground-water discharge to crops and riparian areas likely would decrease as plants capable of using ground water are replaced by lawns or streets, land-surface altitudes are possibly raised by fill material, and drains or sewer lines that may carry ground water to streams are installed. In recharge areas such as White Pine Canyon, Red Pine Canyon, and Willow Draw, increased residential development could result in

Table 1. Area of land use or vegetation and estimated water use, Snyderville Basin, Park City, and adjacent areas, Utah

Land use or vegetation	Area (acres)	Estimated water use by vegetation (feet per year)	Reference for water use
Developed Land			
Residential	3,260	1.4	Utah State University, 1994, p. 293
Irrigated pasture and grass hay	1,620	1.6	Utah State University, 1994, p. 292
Nonirrigated pasture	1,400	1.6	Utah State University, 1994, p. 292
Wet pasture	840	1.6	Utah State University, 1994, p. 292
Golf courses	600	1.4	Utah State University, 1994, p. 293
Irrigated alfalfa	300	2.0	Utah State University, 1994, p. 292
Commercial	290		
Open water	80	2.7	Utah State University, 1994, p. 293
Total area of developed land	**8,390**		
Undeveloped Land			
Sagebrush and perennial grasses	21,670	All precipitation	Wight and others, 1986, table 2 Tomlinson, 1996b, p. 63
Aspen	15,980	1.7	Croft and Monninger, 1953, table 9 Brown and Thompson, 1965, table 3 American Society of Civil Engineers, 1989, p. 17
Gambel Oak	11,110	1.2	American Society of Civil Engineers, 1989, p. 19
Mountain shrub	2,240	0.8	Branson and others, 1970, figure 14
Pinyon-Juniper	2,210	1.7	American Society of Civil Engineers, 1989, p. 20
Dry meadow	1,930	1.4	Tomlinson, 1996a, table 5
Spruce-Fir	620	1.2	Brown and Thompson, 1965, table 3 Kaufmann, 1984, table 2
Riparian	520	2.4	Tomlinson, 1996a, table 5
Total area of undeveloped land	**56,280**		

decreased recharge. Natural seepage from streams could decrease if they are channelized or routed in culverts. Also, natural infiltration of precipitation through the land surface could decrease if much of the land surface becomes covered with impervious structures and surfaces such as buildings and roads.

Hydrogeology

The geology in the study area is very complex and is described in detail by Ashland and others (1996). Simplified hydrogeology of the area, including the principal water-bearing formations, primary structures such as folds and faults, and the outline of principal water-bearing unconsolidated valley fill, is shown on plate 2.

During the Sevier Orogeny, 66 to 100 million years ago, the consolidated rocks in the area underwent thrust and associated reverse and normal faulting (Mayo and others, 1992, p. 244). As a result of the structural deformation, most of the consolidated rocks, including all of the principal water-bearing formations, are extensively folded and fractured. The fractures provide paths for water to recharge consolidated rocks and be transmitted through these rocks to areas of discharge or withdrawal by wells.

Rocks in the study area range in age from Pennsylvanian to Holocene as shown on geologic maps by Bromfield and Crittenden (1971), Bryant (1990), and Crittenden and others (1966). In the northwestern, central, and southern parts of the study area, the principal consolidated rocks are sandstone, limestone, shale, and quartzite, which range in age from Pennsylvanian to

Jurassic. In the northern part of the study area, sedimentary rocks of Jurassic to Cretaceous age, which vary from siltstone to conglomerate, are overlain by Tertiary-age sandstone and conglomerate to the east. In the northeastern and eastern parts of the study area, the principal outcrops are extrusive volcanic deposits of Tertiary age. Extrusive igneous rocks of Tertiary age in the eastern part of the study area have not been considered to be hydrologically important in the past. Because of increased demands for water as a result of expected increased residential growth, however, these rocks are starting to be explored for possible ground-water production. In the southeastern part of the area, igneous intrusions have created additional deformation and faulting of the sedimentary rocks. Unconsolidated valley fill of Quaternary age primarily occurs in the valleys of the study area, but extends to higher altitudes as a veneer in the bottom of tributary drainages.

Unconsolidated Valley Fill

Unconsolidated valley fill in the study area mostly consists of alluvium, undifferentiated glacial outwash and alluvium, and glacial till (Ashland and others, 1996, p. 21; and Bryant, 1990). Alluvium of Holocene age generally underlies the larger perennial streams (Bryant, 1990). This alluvium consists of poorly sorted gravel and cobbles intermixed with clay, silt, and sand and is generally less than 10 ft thick (Ashland and others, 1996, p. 19). Alluvium of Pleistocene age is present along the lower parts of Silver Creek and East Canyon Creek and in the area north of Silver Creek Junction (Bryant, 1990). This alluvium consists of poorly sorted gravel, sand, and silt with estimated maximum thicknesses of 80 ft along lower Silver Creek, 50 ft along lower East Canyon Creek, and 30 ft in the area north of Silver Creek Junction (Ashland and others, 1996, p. 19). Undifferentiated glacial outwash and alluvium is present in the subsurface throughout much of the low-lying part of the study area (Bryant, 1990) and is generally poorly sorted but can be moderately sorted where coarser material was deposited near the foot of glaciers and finer material was carried into the valley (Ashland and others, 1996, p. 19). Glacial till is present generally as a veneer of poorly sorted coarse material in a clay and silt matrix in the upper reaches of tributary drainages in the southern part of the study area. Where glacial till deposits form moraines, the thickness can be much greater (Bryant, 1990).

Unconsolidated valley fill that has sufficient thickness to yield water to wells and springs has been delineated by Ashland and others (1996, fig. 7). These deposits cover about 18 percent of the study area (pl. 2). The areas of thin unconsolidated valley fill not included are in the upper reaches of tributary drainages in the south, along the lower part of East Canyon Creek, and most of the area north of Silver Creek Junction.

The thickness of the unconsolidated valley fill was contoured by Ashland and others (1996, fig. 12) in the Park City and Park Meadows area, lower Silver Creek, and Snyderville Basin. Contours show that the thickness of unconsolidated valley fill in the Park Meadows area is less than 80 ft. Lithologic logs reported by Mason (1989, table 3), however, indicate a maximum thickness of more than 130 ft in this area. In the lower Silver Creek area, between Keetley Junction and Silver Creek Junction, the thickness of unconsolidated valley fill probably is more than 80 ft along the north trending axis of the valley. In Snyderville Basin, the thickness of the unconsolidated valley fill probably is more than 270 ft.

Consolidated Rocks

The ability of consolidated rocks to accept recharge, transmit water, and yield water to wells and springs varies with lithologic character and prominence of fractures. Four consolidated-rock formations within the study area readily accept seepage of precipitation from land surface at higher altitudes and transmit the water through connected fractures to points of discharge: the Twin Creek Limestone, Nugget Sandstone, Thaynes Formation, and Weber Quartzite. For this reason, only these four consolidated-rock formations will be discussed in detail. Although other consolidated-rock formations in the study area transmit and contain some water, they are mentioned only as to their stratigraphic or structural relation, and location in the study area.

The extrusive Keetley Volcanics of Tertiary age unconformably overlie older consolidated rocks in the eastern part of the study area. The Ankareh Formation of Triassic age conformably underlies the Nugget Sandstone and overlies the Thaynes Formation. The Ankareh Formation crops out in several areas in the Wasatch Range along the western part of the study area and southwest of Park City and in the low hills that are just northwest of the Park Meadows area. The Woodside Shale, also of Triassic age, conformably underlies the Thaynes Formation and overlies the Park City Formation. The shale crops out in the western part of the study area, west and southwest of Kimball Junction, southwest of Park City, and in the Park Meadows area. The Park City Formation of Permian age conformably underlies and crops out in areas adjacent to the Woodside Shale.

Ashland and others (1996, p. 46, p. 55) suggest that water-bearing consolidated-rock formations can be divided into discrete "ground-water compartments" on the basis of confining units in stratigraphically adjacent formations and by major faults and folds. Major faults could restrict ground-water flow in a direction perpendicular to fault planes because of the presence of clay gouge, but would enhance ground-water flow parallel to fault planes because of the presence of intensely fractured zones (Ashland and others, 1996, p. 55). Although short-term water-level fluctuations from pumping of ground water might support these concepts, the lack of long-term, spatially distributed data throughout the complex ground-water system in the study area prevents definitive statements regarding ground-water compartmentalization.

The extent and the degree of connectivity between fractures within a consolidated-rock formation, in part, determine its ability to transmit and yield water to wells. The void spaces in rocks, formed by packing of grains, solution cavities, or fractures, are where ground water resides. These spaces determine the property known as porosity. An explanation of porosity is reported by Domenico and Schwartz (1990, p. 24-26). Total porosity is the ratio of the volume of void space in a given rock to the total volume of rock mass. Effective porosity is the ratio of interconnected void space in a given rock to the total volume of rock mass. Primary porosity is the ratio of the volume of void space to the total volume of rock mass in the formation after deposition and lithification before any chemical or physical alteration. Primary porosity of consolidated rock generally is much smaller than porosity of unconsolidated deposits. In a sedimentary consolidated rock, primary porosity will be affected by grain shape and arrangement and by chemical and physical processes that have affected the rock since deposition. Secondary porosity is the void space created by fractures or openings resulting from chemical dissolution. If fractures and solution openings mostly are unconnected, then the resulting effective porosity is much smaller than total porosity. If fractures and solution openings mostly are connected, then the resulting effective porosity can be nearly as large as total porosity. Secondary porosity can be greater than primary porosity in rocks where primary porosity is characteristically small, such as in some igneous and metamorphic rocks. Secondary porosity from fractures and solution openings, if interconnected, can enhance the transmissive properties of consolidated rock.

A qualitative indication of transmissiveness in consolidated rock can be obtained by mapping and defining fracture characteristics at land surface where the rock crops out or in underground mine workings. This task was completed by personnel of the Utah Geological Survey and the fracture characteristics and trends are summarized by Ashland and others (1996, fig. H. 1, tables G. 1, G. 2, H. 1, and H. 2, pls. 12-15). Fracture characteristics that are indicative of high secondary porosity include aperture (width), persistence (length), planarity, roughness, and degree of mineral infilling. Fracture types within the study area include joints, faults, bedding fractures (parallel to bedding), and cleavage fractures (Ashland and others, 1996, p. 30).

The degree of fracture connectivity, which can be used as a measure of effective porosity in three-dimensional space, is difficult to determine from flat, two-dimensional outcrops. Ashland and others (1996, p. 43) examined fracture characteristics in the Thaynes Formation, Woodside Shale, Park City Formation, and Weber Quartzite exposed in the Spiro Tunnel (pl. 1). They report that fracture characteristics are similar to those at land surface.

In addition to fracture characteristics, size and location of consolidated-rock outcrops are important in determining the ability of the formation to yield water to wells and springs. A consolidated-rock formation that crops out or is close to land surface at higher altitudes where snowmelt is greater than at lower altitudes will have more potential for water availability than a formation that crops out only at lower altitudes. Also, larger outcrops provide more area for the infiltration of water than do smaller outcrops. Topography, vegetative cover, soil cover, and other factors also influence infiltration.

Twin Creek Limestone

The Twin Creek Limestone of Middle Jurassic age consists of seven members as defined by Imlay (1967). Lithologic character varies from a red to brownish, soft siltstone in the basal Gypsum Spring and intermediate Boundary Ridge Members to thin- to medium-bedded, light- to purplish-gray limestone in most of the other members and becomes a silty to sandy limestone in the upper Giraffe Creek Member (Imlay, 1967, table 1). The basal Gypsum Spring Member is gypsiferous in parts of the study area and could be a confining layer (Ashland and others, 1996, p. 10). The Twin Creek Limestone is about 2,600 ft thick in the western part of the study area and is estimated to be about 1,400 ft thick near the eastern boundary (Ashland and others, 1996, p. 10).

Joints in the Twin Creek Limestone tend to be moderately open to open (0.04 to 0.4 in.), but persistence is very low (less than 3.3 ft) except for fractures parallel to bedding. Faults tend to be open (0.1 to 1.2 in.) and persistence is low to moderate with many exceeding 16 ft. Clay gouge and breccia zones are associated with the faults (Ashland and others, 1996, tables 2 and 3). Even though fracture persistence ranges from very low to moderate, the degree of fracture connectivity can be high if bedding thicknesses are less than 3.3 ft. Bedding joints tend to be more persistent. Because the Twin Creek Limestone generally is steeply dipping, bedding joints are exposed at land surface in the study area and where water is likely to infiltrate into the formation. From these observations, the capacity for transmitting water through the Twin Creek Limestone in the study area probably is most related to secondary porosity.

The Twin Creek Limestone crops out in the mountains to the west, south, and east of Snyderville Basin south of Interstate Highway 80 (pl. 2). The Twin Creek Limestone is covered by shallow valley fill in the Kimball Junction area and in the southern end of Snyderville Basin. Outcrops are visible in the Summit Park area at the western boundary of the study area. This block of the Twin Creek Limestone is steeply dipping and in some areas is overturned as a result of the folding in the upper block associated with the Mount Raymond-Absaroka thrust fault (pl. 2). The limestone also is bisected by the Toll Canyon fault. This outcrop area extends into Lambs Canyon, just west of the study-area boundary.

The Twin Creek Limestone in the low mountains and under the unconsolidated valley fill just north of Kimball Junction also is in the folded, upper block associated with the Mount Raymond-Absaroka thrust fault. This block dips to the northeast and is bounded on the north by an unnamed fault and on the south by the Mount Raymond-Absaroka thrust fault where it butts against another block of Twin Creek Limestone (Ashland and others, 1996, pls. 4 and 5). North of Interstate Highway 80, the Twin Creek Limestone is overlain by Tertiary sedimentary deposits of unknown thickness.

South of Kimball Junction, where the Twin Creek Limestone crops out in the mountains and underlies the unconsolidated valley fill, it forms the limbs of the northeast plunging Willow Draw anticline (pl. 2). This block is bounded by the Nugget Sandstone toward the core of the anticline and thrust faults at the outer edge of the anticline. Where the fault trace is queried, as shown on plate 2, two possible interpretations exist. Each interpretation has important geologic and hydrologic ramifications. Crittenden and others (1966) show a sinuous trace that crosses the Dutch Draw syncline just west of White Pine Canyon. New structual relations visible in outcrops exposed as a result of recent development have been used by Ashland and others (1996, fig. 3) to indicate that the Mount Raymond-Absaroka thrust fault trends in a northeast direction and that the fault located between the Willow Draw anticline and the Dutch Draw syncline is a separate backthrust fault. If the first interpretation is correct, the Twin Creek Limestone exposed in the Willow Draw anticline would be connected beneath the Dutch Draw syncline to the Twin Creek Limestone under the unconsolidated valley fill in the southern part of Snyderville Basin and where exposed east of the unconsolidated valley fill south of Interstate Highway 80. If the newer interpretation is correct, then the Twin Creek Limestone beneath unconsolidated valley fill in the southern part of Snyderville Basin and that crops out just to the east is not connected to the Twin Creek Limestone in the Willow Draw anticline and may not be connected to the Twin Creek Limestone in the mountains southwest of Snyderville. The east block of Twin Creek Limestone, therefore, could be isolated from the blocks exposed at higher altitudes and would not have the potential to yield a large amount of ground water to wells.

Nugget Sandstone

The Nugget Sandstone consists of fairly uniform pale red to reddish orange, very fine- to medium-grained, cross-bedded sandstone (Bromfield, 1968, p. 19; Crittenden and others, 1966). Bromfield (1968, p. 19) and Bryant (1990) have reported the age of the Nugget Sandstone as uncertain, Triassic?, or Jurassic? Ashland and others (1996, p. 9) and Crittenden and others (1966) report the age of the Nugget Sandstone as Jurassic. The Nugget Sandstone varies in estimated thickness from about 800 ft near Park City to about 1,400 ft near the western boundary of the study area (Ashland and others, 1996, p. 10).

Joints in the Nugget Sandstone are moderately open to open with apertures typically less than 0.2 in. and persistence is low with localized joints of medium or higher persistence (Ashland and others, 1996, table 2). Faults are tight with low persistence (Ashland and others, 1996, table 3). Breccia zones near faults generally are 6.5 ft with sandy gouge. The Nugget Sandstone near major faults is reported to be intensely fractured. Throughout most of the Nugget Sandstone, the capacity to transmit water would be related to primary porosity. Near major fault zones, the capacity to transmit water may be more related to secondary porosity.

Along the northwestern study-area boundary (pl. 2), a small outcrop of Nugget Sandstone is exposed just north of the Toll Canyon fault (Ashland and others, 1996, pl. 6 and 7). This outcrop extends into Lambs Canyon, just west of the study-area boundary. The block of Nugget Sandstone exposed south of the Toll Canyon fault is juxtaposed at depth to the block north of the fault (Ashland and others, 1996, pl. 3).

An arc-shaped block of Nugget Sandstone is exposed or underlies shallow unconsolidated valley fill north of Interstate Highway 80 (Ashland and others, 1996, pl. 6 and 7). At the west end of this block, the Nugget Sandstone is terminated by the Toll Canyon fault. At the east end, it is terminated by the Mount Raymond-Absaroka thrust fault.

The Nugget Sandstone is exposed or underlies a veneer of unconsolidated deposits in the mountains to the west, south, and east of Snyderville Basin south of Interstate Highway 80. The large exposed surface area in the higher mountains to the west and south provides the potential for a substantial amount of water from snowmelt to infiltrate into the Nugget Sandstone. As with the Twin Creek Limestone in this area, the structural relation of the Nugget Sandstone below land surface depends on the accepted interpretation of the Mount Raymond-Absaroka thrust fault. The sinuous fault trace of Crittenden and others (1966) indicates that the Nugget Sandstone exposed in the center of the Willow Draw anticline is connected in the subsurface to the outcrop east of the unconsolidated valley fill. The isolated outcrop of Nugget Sandstone shown in the upper block of the thrust fault, just northwest of the Dutch Draw syncline, is not connected to the other outcrops. The alternative interpretation presented by Ashland and others (1996, fig. 3, pl. 3) indicates that the isolated block would be connected to the outcrop east of the unconsolidated valley fill. The block of Nugget Sandstone in the Willow Draw anticline would then dip steeply to the southeast.

Thaynes Formation

The Thaynes Formation of Triassic age is a sequence of dark-brown and gray limestone, limy sandstone and siltstone, greenish micaceous shales, and red shale and siltstone (Bromfield, 1968, p. 17). Boutwell (1912, p. 55) reports that the red shale divides the two carbonate sequences into approximately equal thicknesses. Thin beds of gypsum have been reported in the Thaynes Formation (Withington, 1964, p. 184). Gypsum layers are only a few feet thick where exposed at the surface, but beds as much as 10 ft thick are present in the subsurface. Bromfield (1968, p. 17) estimated the thickness to range from 1,100 to 1,300 ft. Ashland and others (1996, table 1) report an estimated thickness for the Thaynes Formation of 2,200 ft in the upper block of the Mount Raymond-Absaroka thrust and 1,150 ft in the lower block.

Ashland and others (1996, table 2) describe joints in the Thaynes Formation as being tight to moderately open, typically less than 0.2 in. Joint persistence varies but is more than in the Twin Creek Limestone, with bedding joint persistence medium or higher, more than 9.8 ft. Faults are open to wide (0.1 to 1.2 in.) and persistence is generally low, but one-third of the faults that were recorded had a persistence of medium or higher (more than 9.8 ft) (Ashland and others, 1996, table 3). Because of the secondary porosity that results from the fractures described by these characteristics, some wells completed in the Thaynes Formation yield more water than other wells in the study area.

The Thaynes Formation crops out in the mountains west and southwest of Kimball Junction and underlies shallow valley fill to the northeast (pl. 2). This block of Thaynes Formation extends to the southwest beyond the study-area boundary into Lambs Canyon, just west of the study area. Because of folding associated in the upper block of the Mount Raymond-Absaroka thrust fault, the formation dips steeply and in some areas it is overturned (Crittenden and others, 1966). Additional faults are present west of Kimball Junction.

The Thaynes Formation also crops out in the mountains west and south of Park City. The formation underlies shallow valley fill in the Park Meadows area and crops out in the low hills to the northeast. The Thaynes Formation in this area dips steeply to the northwest (Ashland and others, 1996, pl. 8).

Weber Quartzite

The Weber Quartzite of Pennsylvanian age consists primarily of medium- to thick-bedded pale gray, tan-weathering, fine-grained quartzite and sandstone (Bromfield, 1968, p. 16). The lower part of the formation is interbedded quartzite and sandstone, the middle part is massive quartzite, and the upper part is interbedded quartzite and limestone. Limestone makes up about 15 to 20 percent of the formation (Ashland and others, 1996, p. 9). The thickness of the Weber Quartzite in the study area is probably 1,300-1,500 ft (Bromfield, 1968, p. 16).

The Weber Quartzite is brittle and easily fractured. Joints in the Weber Quartzite are tight to moderately open and persistence is generally low except in localized areas where persistence is medium and higher, second only to the Nugget Sandstone (Ashland and others, 1996, table 2). Faults are tight to moderately open with persistence being very low to medium. Intensely fractured fault zones were reported by Ashland and others (1996, table 3). Because of fractures and fractured fault zones, the capacity for transmitting water through the Weber Quartzite is primarily related to secondary porosity.

Outcrops of the Weber Quartzite are limited to the southeastern part of the study area (pl. 2). This block of Weber Quartzite extends beyond the study-area boundary to the south and east. The Weber Quartzite in this area contains many high-angled faults. Much of the southern part of this block is located within the Park City mining district, and much of the information regarding structural relations is proprietary. The Weber Quartzite is cut by the Frog Valley thrust fault (Ashland and others, 1996, pl. 3). The upper block associated with the thrust fault dips to the northwest and the lower block dips to the east (Ashland and others, 1996, pl. 10).

SURFACE-WATER HYDROLOGY

Surface water in the study area originates in the Wasatch Range on the southern and western borders of the area and exits through canyons to the north. Some streamflow is diverted near the mouths of canyons and used for irrigation. Some streamflow infiltrates into the ground in the canyons or near the mouths of canyons and recharges consolidated rocks and unconsolidated valley fill. Additional streamflow is derived from ground-water discharge in the lower parts of the study area, especially near Snyderville, Kimball Junction, and Park Meadows. Many of the original stream channels have been altered during mining, commercial and residential development, or irrigation.

The two major streams in the study area are East Canyon Creek and Silver Creek. East Canyon Creek begins as McLeod Creek where it receives water from snowmelt and ground water from the Spiro Tunnel and springs near Thaynes Canyon. McLeod Creek gains additional water from small perennial and ephemeral streams and ground-water discharge as it flows through Snyderville Basin to Kimball Junction. East Canyon Creek, which begins where McLeod Creek joins Kimball Creek, exits through a canyon in the northwestern part of the study area. Silver Creek receives water from snowmelt and ground water from mine tunnels and springs in the Deer Valley area along with water diverted from the Spiro Tunnel. Silver Creek gains additional water in the Park Meadows area where ground water discharges into the Pace-Homer Ditch, which joins Silver Creek downstream from the Prospector Square area. Silver Creek exits through a canyon in the northeastern part of the study area.

Streamflow

Streamflow-gaging stations in the study area and annual streamflows are listed in table 2. Locations are shown on plate 1. The daily mean flow of six streams is shown in figure 4. Streamflow is seasonal, with 70 to 100 percent of the flow in streams occurring from March to July. Streamflow throughout the study area was much higher during the 1995 water year than during the 1994 water year (fig. 4). As explained in the "Climate" section of this report, precipitation was much greater in the 1995 water year than in the 1994 water year. Runoff from snowmelt produced two discernible peaks in late winter and spring of 1995, whereas generally a single period of increased streamflow occurred in 1994.

McLeod Creek originates near the mouth of Thaynes Canyon where water discharges from Sullivan Springs, (D-2-4)8cab-S1. Streamflow in Thaynes Canyon upstream from Sullivan Springs rarely occurs, even during the peak of snowmelt runoff. Flow from the Spiro Tunnel also contributes to the initial flow in McLeod Creek. The flow from the Spiro Tunnel that is not used for municipal supply or irrigation and is not diverted to the Silver Creek drainage flows into McLeod Creek. Flow from the Spiro Tunnel is measured continually by the Park City Municipal Corpora-

Table 2. Annual flow at streamflow-gaging stations used in water-budget analysis, Snyderville Basin and adjacent areas, Utah

[All flows reported in acre-feet per year]

Low annual flow: Occurred in 1992 for all streams with records.
1995 flow: Also the high annual flow for all streams with records.
March through July: Percentage of annual flow that occurs from March through July. Percentage of mean annual flow if known, otherwise percentage of 1995 flow.

Streamflow-gaging station	Period of record	Mean annual flow	Lowest annual flow	1995 flow	March through July (percent)
McLeod Creek near Park City, Utah	October 1990—present[1]	9,190	4,650	16,220	70
East Canyon Creek above Big Bear Hollow near Park City, Utah	November 1989—present[1]	21,280	8,180	40,690	75
Kimball Creek above East Canyon Creek near Park City, Utah	October 1989—present[1]	1,420	150	3,870	97
White Pine Canyon near Park City, Utah	May 1994—September 1995	—	—	3,780	86
Unnamed Creek (Spring Creek) near Park City, Utah	August 1994—September 1995	—	—	7,380	88
Silver Creek near Wanship, Utah	October 1941—September 1946 July 1982—September 1985 October 1989—September 1996	6,150	3,060	10,700	74
Red Pine Canyon				[2]1,500	100
Willow Draw				[2]950	78

[1] Streamflow-gaging station in operation at time of publication of this report (1998).

[2] Estimated from weir readings (John Bollwinkel, Community Water Company, written commun., 1996).

tion. The flow that is diverted to McLeod Creek and the Silver Creek drainage is measured at Parshall flumes only during the summer months (Rich Hilbert, Park City Water Department, written commun., 1996); therefore, annual flows to each stream are estimated.

Streamflow in White Pine Canyon flows into McLeod Creek about 0.7 mi southeast of Snyderville. A gaging station (U.S. Geological Survey streamflow-gaging station 404039111325700) was operated in White Pine Canyon during the data-collection period of this study. Streamflow was about 0.5 ft^3/s during winter months and peaked at about 57 ft^3/s in June 1995 (fig. 4).

A gaging station (U.S. Geological Survey streamflow-gaging station 10133600) on McLeod Creek is about 3.2 mi northwest of Park City downstream from where streamflow from White Pine Canyon enters. Minimum streamflow was about 3 ft^3/s and maximum streamflow was 117 ft^3/s for the 1995 water year (fig. 4). Minimum streamflow occurs during winter months and during late summer months after snowmelt runoff ceases and when much of the streamflow has been diverted for irrigation. Maximum streamflow occurs during spring and early summer months from snowmelt runoff and discharge from Sullivan Springs and the Spiro Tunnel. Upper McLeod Creek drains some area at a lower altitude that results in a smaller peak in March 1994 and 1995 prior to the much larger peak in May 1994 and June 1995 (fig. 4).

Water from McLeod Creek just north of the gaging station is diverted to the Old Ranch Road area. Some of this water is used for irrigation, whereas some water seeps into the unconsolidated valley fill or returns to McLeod Creek. Ground water discharges from the unconsolidated valley fill into small stream channels and into McLeod Creek in the area south of Interstate Highway 80. Near Silver Creek Junction, McLeod

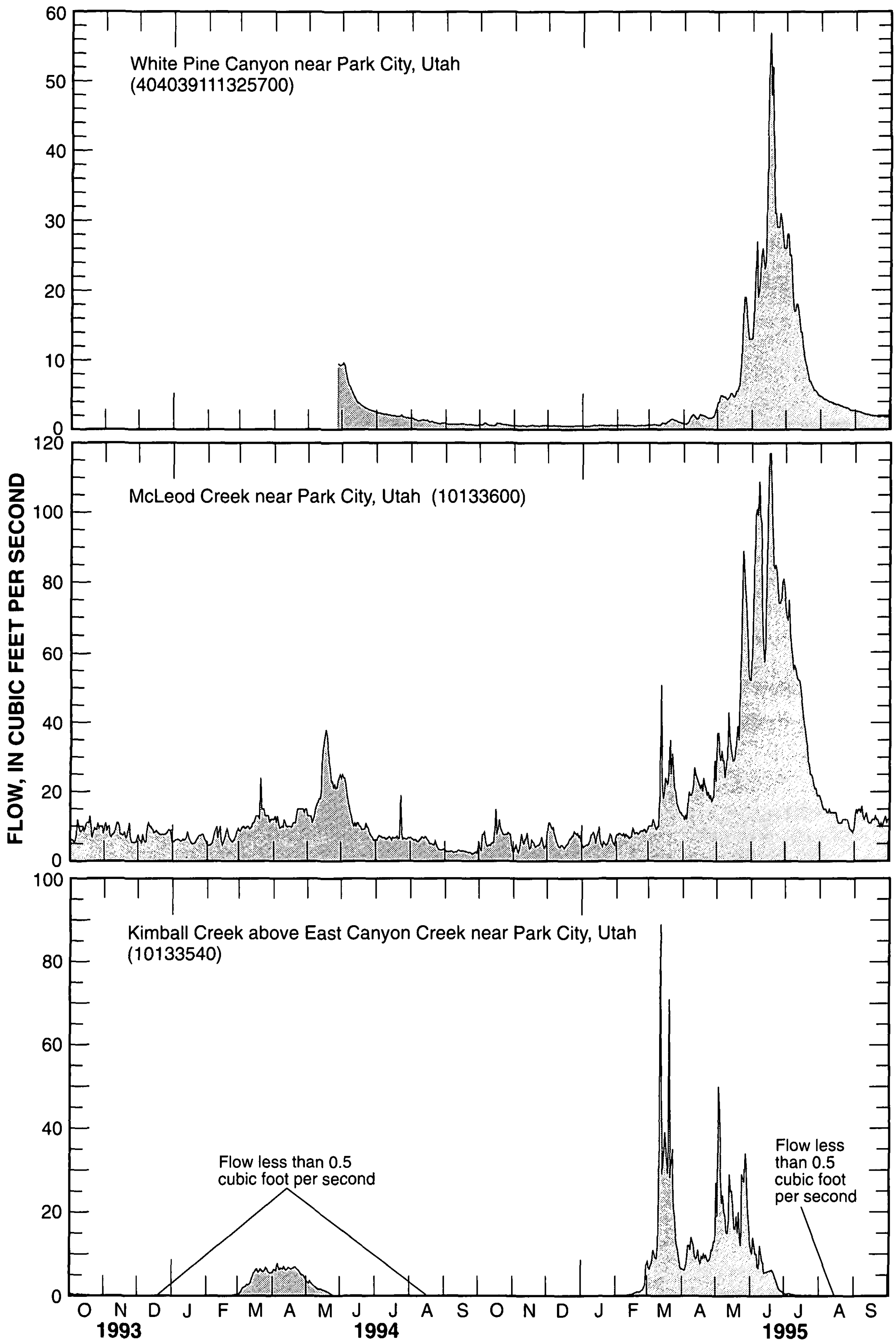


Figure 4. Daily mean flow of selected streams in Snyderville Basin and adjacent areas, Utah.

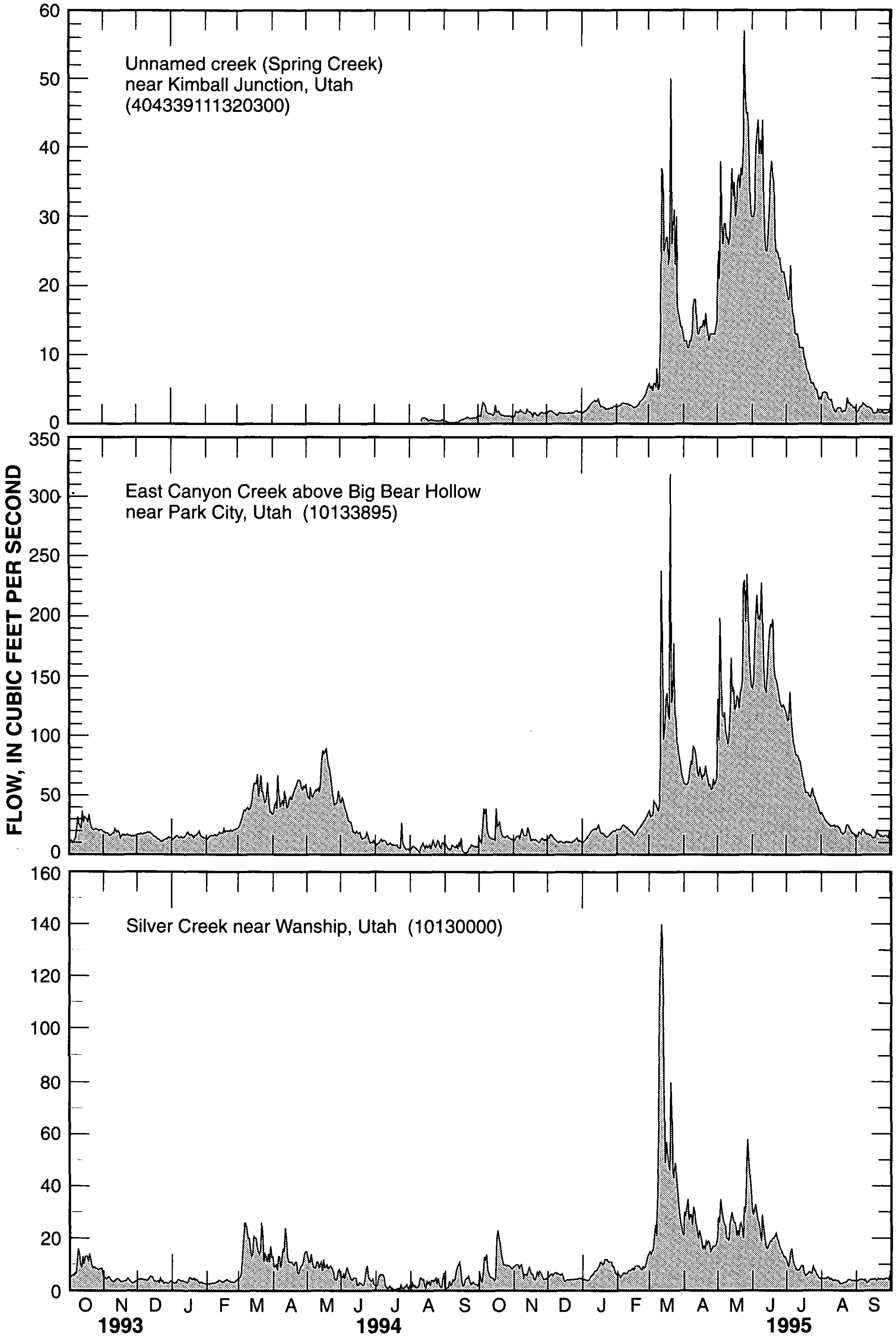


Figure 4. Daily mean flow of selected streams in Snyderville Basin and adjacent areas, Utah—Continued.

Creek merges with Kimball Creek to form East Canyon Creek.

The gage on Kimball Creek (U.S. Geological Survey streamflow-gaging station 10133540) is upstream from the confluence with McLeod Creek at a Utah Department of Transportation rest area 1.5 mi east of Kimball Junction. Streamflow in Kimball Creek for 1995 peaked in March at 89 ft^3/s during the runoff of low-altitude snowmelt. Minimum streamflow in Kimball Creek was less than 0.5 ft^3/s during most of the year (fig. 4). Runoff from snowmelt in the Kimball Creek drainage was much less in 1994 than 1995 because of less snowpack and drier conditions.

An unnamed creek (locally called Spring Creek) enters East Canyon Creek near Kimball Junction. During most of the year, water in the creek is derived from spring (D-1-3)36aad-S1 and from ground-water discharge between the spring and where the creek merges with East Canyon Creek. A gage (U.S. Geological Survey streamflow-gaging station 404339111320300) was operated near the mouth of this creek from August 1994 to September 1995. Streamflow in 1995 ranged from about 2 ft^3/s during low-flow conditions to a peak of 57 ft^3/s in May (fig. 4). Discharge measurements of spring (D-1-3)36aad-S1 provided by Silver Springs Water Company (David Polichette, Silver Springs Water Company, written commun., 1996) indicate that about one-half of the low streamflow and about one-quarter of the peak streamflow in the creek is derived from the spring. Similar to streamflow for the 1995 water year at other gaging stations in the study area, a distinct peak was present in March and a longer period of high streamflow was present through May, June, and into July. The peak in March is the result of runoff from low-altitude snowmelt. The prolonged second peak is the result of increased discharge from spring (D-1-3)36aad-S1 and ground-water discharge from the valley fill.

Red Pine Canyon and Willow Draw contribute no surface flow directly to McLeod Creek or the unnamed creek, but contribute recharge to consolidated rock in the upper reaches and unconsolidated valley fill in the lower reaches of both drainages. Streamflow in Red Pine Canyon and Willow Draw during high flows dissipates as it flows toward the residential area east of Snyderville. It is unknown whether coarse fill material or buried drains used in the construction of the residential area facilitates the seepage of water in these drainages into the valley fill. Much of this water probably resurfaces as ground-water discharge to the unnamed creek later in the year. Streamflow in Red Pine Canyon and Willow Draw was estimated from weir measurements (John Bollwinkel, Community Water Company, written commun., 1996). Some flow in Willow Draw is treated for municipal use by Community Water Company.

Several ungaged streams contribute flow to East Canyon Creek between Kimball Junction and the gaging station on East Canyon Creek above Big Bear Hollow, near Park City, Utah. These streams were only measured a few times during this study, but were measured during a previous study (Holmes and others, 1986, table 1).

Streamflow exiting the study area in East Canyon Creek is measured at a gage (U.S. Geological Survey streamflow-gaging station 10133895) upstream from Big Bear Hollow about 10 mi northwest of Park City, Utah. A gage was operated below Big Bear Hollow from 1982 to 1984 (U.S. Geological Survey streamflow-gaging station 10133900). Low streamflow in East Canyon Creek during 1995 was about 15 ft^3/s (fig. 4), of which about 10 percent was discharge from the wastewater-treatment plant as determined from data provided by Snyderville Basin Sewer Improvement District (Rex Osborne, Snyderville Basin Sewer Improvement District, written commun., 1996). Maximum streamflow for the 1995 water year was 319 ft^3/s in March during runoff of low-altitude snowmelt. Discharge from the wastewater-treatment plant also increased, but the wastewater percentage of total flow was much lower than it was during low streamflow conditions.

Ungaged streamflow entering East Canyon Creek from Big Bear Hollow downstream from the gaging station was estimated by developing a regression equation for Big Bear Hollow and East Canyon Creek for October 1982 through September 1984. This equation was used to estimate streamflow in Big Bear Hollow from October 1993 through September 1995. Streamflow was estimated to be 270 acre-ft in 1994 and 1,100 acre-ft in 1995.

Silver Creek originates at the base of the mountains in the southern part of the study area. Similar to streamflow in Thaynes Canyon, streamflow in Silver Creek through Park City is less than what is expected during late spring and early summer when snowmelt runoff should be at its peak. Although no gaging station was operated in the upper reaches of Silver Creek during this study, instantaneous measurements of streamflow were made (Downhour and Brooks, 1996, table 7).

No large streamflows were observed in upper Silver Creek during the runoff of snowmelt in the spring of 1995 as were measured in other drainages in the study area. This suggests that most of the runoff from snowmelt in the upper Silver Creek drainage seeps into the subsurface prior to reaching a stream channel. Upper Silver Creek flows through Park City and along the south edge of the Park Meadows residential area before entering a small canyon as the water flows toward Keetley Junction.

The Pace-Homer Ditch, located on the east side of the Park Meadows area, collects the discharge from Dority Springs ((D-2-4)4dca-S1), water diverted from the Spiro Tunnel, unused irrigation water from McLeod Creek, ground-water seepage to drains, and ground-water seepage directly to the ditch. Flume measurements of discharge from Dority Springs and flow in the Pace-Homer Ditch were provided by Park City Water Department (Rich Hilbert, Park City Water Department, written commun., 1996). The Pace-Homer Ditch exits the Park Meadows area to the east through a small canyon and parallels Silver Creek. Southwest of Keetley Junction, the Pace-Homer Ditch terminates where the water dissipates over the land surface or drains into Silver Creek.

North of Keetley Junction, most of the streamflow in Silver Creek is diverted for irrigation use in the area southeast of Silver Creek Junction. Streamflow in Silver Creek that exits the study area is measured at a gaging station (U.S. Geological Survey streamflow-gaging station 10130000) downstream from Silver Creek Junction. During most of the 1995 water year, streamflow at this gage ranged from about 3 to 10 ft^3/s (fig. 4). Much of this water is derived from ground-water seepage or discharge from the wastewater-treatment plant located near Silver Creek Junction. Discharge from the wastewater-treatment plant ranged from about 1 to more than 4 ft^3/s (Rex Osborne, Snyderville Basin Sewer Improvement District, written commun., 1996). Discharge from the wastewater-treatment plant is about 1 ft^3/s during normal-use periods and about 2 ft^3/s during high-use periods. The maximum flow in Silver Creek was 140 ft^3/s in March 1995 during runoff from low-altitude snowmelt. The peak discharge from the wastewater-treatment plant coincided with runoff from low-altitude snowmelt. This indicates that water from melting snow readily seeps into the subsurface and into wastewater transmission pipes.

The secondary peak in May 1995, which coincides with runoff from high-altitude snowmelt throughout the study area, is smaller than the peak in March 1995 (fig. 4). Much of the runoff from high-altitude snowmelt probably seeps into the subsurface prior to reaching a stream channel. Also, the high-altitude surface area is much smaller as compared to the low-altitude surface area in the Silver Creek drainage and thus contributes less to the runoff in this drainage. On the basis of estimated streamflow in Silver Creek exiting the Park Meadows area and in the Pace-Homer Ditch, about 50 percent of streamflow measured at the gage on Silver Creek during the peak runoff from high-altitude snowmelt in May 1995 was derived from the Pace-Homer Ditch. Only about 20 percent of the measured streamflow comes from upper Silver Creek and the remaining 30 percent comes from ground-water discharge to lower Silver Creek (see "Water-budget analysis" section later in this report).

Because data were insufficient to develop a regression equation, ungaged streamflow from Tollgate Canyon entering Silver Creek downstream from the gaging station was estimated on the basis of drainage area. The Tollgate Canyon drainage area is adjacent to the Kimball Creek drainage and has similar hydrologic and geologic characteristics. The drainage area for Tollgate Canyon is about 1,900 acres, about 24 percent of the Kimball Creek drainage area at the U.S. Geological Survey gaging station. Streamflow for Tollgate Canyon, therefore, was estimated to be 24 percent of the measured streamflow in Kimball Creek. Streamflow in Tollgate Canyon was estimated to be 180 acre-ft in 1994 and 900 acre-ft in 1995.

Surface-Water Quality

The quality of surface water depends on the source of the water. During snowmelt runoff, streamflow consists mostly of water from snow, which has a low dissolved-solids concentration, and the dissolved-solids concentration in a stream is reduced. During most of the year, streamflow primarily consists of base flow and discharge from springs and drain tunnels, which have a higher dissolved-solids concentration, and the dissolved-solids concentration in a stream is increased. This trend is indicated by data reported by Holmes and others (1986, table 13), Mason (1989, tables 6 and 8), and Downhour and Brooks (1996, table 6). The dilution of dissolved-solids concentration during high streamflow is evident in tributary streams where snowmelt runoff is largely relative to base flow, such as in White Pine Canyon. When streamflow was high in White Pine Canyon (U.S. Geological Survey streamflow-gaging station 404039111325700), more than 45 ft^3/s on June 14, 1995, the specific conductance was 165 μS/cm, almost one-half the specific conductance during low streamflow (Downhour and Brooks, 1996, table 6). Along some stream reaches, the predominant anion in the water, which is dependent upon the source of water, can change seasonally.

The specific conductance of surface water in the East Canyon Creek drainage ranged from 145 to 1,870 μS/cm during this study (Downhour and Brooks, 1996, tables 4, 6, and 7). Almost all values were less than 1,000 μS/cm. The highest value was measured in the creek emanating from Toll Canyon where it merges with East Canyon Creek.

Holmes and others (1986, p. 31) report that samples collected in East Canyon Creek drainage are generally of two water types or a mixture of both types. Water in the major tributaries generally is of a calcium bicarbonate type, and water in the main stem of East Canyon Creek is of a calcium sulfate, calcium sulfate bicarbonate, or calcium bicarbonate sulfate type. They attribute the predominance of sulfate in the East Canyon Creek drainage area to discharge from the Spiro Tunnel, which has a high sulfate concentration relative to most other sources in the study area. Data collected during this study indicate that this relation is still valid. An analysis of water from McLeod Creek near Interstate Highway 80 in March 1995 contained higher sodium and chloride concentrations (Downhour and Brooks, 1996, table 4) than previously reported. These higher concentrations are representative of the prevalent use of road salt during the winter months.

Water in the unnamed creek is derived primarily from spring (D-1-3)36aad-S1 and from ground-water discharge between the spring and where the creek merges with East Canyon Creek near Interstate Highway 80. Water that discharges from this spring is derived from the Twin Creek Limestone and is of a calcium bicarbonate type as determined from data reported by Holmes and others (1986, table 13). During high and low streamflow, water from the unnamed creek near Interstate Highway 80 similarly was of a calcium bicarbonate type. From an analysis of water collected in February 1995 (Downhour and Brooks, 1996, table 4), the water was a sodium calcium chloride type. High sodium and chloride concentrations in this sample (150 and 320 mg/L, respectively), relative to those reported by Holmes and others (1986, table 13), are a result of the use and storage of road salt that had dissolved and was transported by surface runoff. Much of the sodium and chloride probably seeps into the unconsolidated valley fill. Sodium and chloride might be prevalent in the unnamed creek near Interstate Highway 80 during low streamflow in the summer months, when most of the water is derived from ground-water discharge. Samples were not collected during low streamflow.

Specific conductance of surface water in the Silver Creek drainage ranged from about 400 to 1,730 μS/cm during this study, with almost all values more than 800 μS/cm (Downhour and Brooks, 1996, tables 4, 6, and 7). The generally higher specific-conductance values in the water from this drainage than of that in the East Canyon Creek drainage are the result of discharge from consolidated rocks that contain mineralized zones, or contact with mine tailings. The higher specific-conductance values generally are present upstream from where the Pace-Homer Ditch flows into Silver Creek.

Water in Silver Creek, upstream from where the Pace-Homer Ditch enters, generally is of a sodium calcium chloride to a sodium calcium chloride sulfate type. During the winter months, when the use of road salt is widespread, the water is a sodium calcium chloride type as represented by the sample collected in March 1995 (Downhour and Brooks, 1996, table 4). During the summer months, when the streamflow in Silver Creek is low, chloride is less prominent and sulfate becomes more prominent as represented by chemical analyses presented by Mason (1989, table 10).

Water in the Pace-Homer Ditch generally is a calcium sulfate bicarbonate or calcium magnesium sulfate bicarbonate type. Much of the calcium and sulfate are derived from water that is diverted to the Silver Creek drainage from the Spiro Tunnel to meet downstream water use. Much of the calcium, magnesium, and bicarbonate probably is derived from discharge from Dority Springs and upward leakage from consolidated rocks into the unconsolidated valley fill. Calcium and sulfate concentrations in water from the Park Meadows collection box and Pace-Homer Ditch increased noticeably from April 1988 to March 1995 (Downhour and Brooks, 1996, table 4). An increase in the amount of water diverted to the Silver Creek drainage from the Spiro Tunnel is the most probable source for the increase in calcium and sulfate.

Chemical analyses of water in Silver Creek at Keetley Junction and downstream from Silver Creek Junction (Holmes and others, 1986, table 14) indicate that the water is primarily a calcium sulfate bicarbonate type, similar to the water in the Pace-Homer Ditch. Most of the streamflow in lower Silver Creek is from the Pace-Homer Ditch and, hence, the similarity.

The only surface-water samples collected by the U.S. Geological Survey that have sulfate concentrations higher than 250 mg/L were collected from Silver Creek downstream from Park City during low flow, and from McLeod Creek near Park City where most of the flow is from the Spiro Tunnel. The only surface-water samples collected by the U.S. Geological Survey that have sulfate concentrations from 200 to 250 mg/L were collected from McLeod Creek near Park City, McLeod Creek at Interstate Highway 80, and Silver Creek near Keetley Junction.

GROUND-WATER HYDROLOGY

The ground-water system in Snyderville Basin, Park City, and adjacent areas is in consolidated rocks and unconsolidated valley fill. All public-supply wells in the study area are completed in consolidated rocks, mostly in the Twin Creek Limestone, the Nugget Sandstone, and the Thaynes Formation. A few wells are completed in the Ankareh Formation adjacent to the Nugget Sandstone in the Willow Draw and Pinebrook areas (Bryant, 1990, sh. 1). No wells are completed in the Weber Quartzite, but mine tunnels discharge water from the Weber Quartzite. The unconsolidated valley fill is less than 100 ft thick in most areas (as discussed in the "Hydrogeology" section of this report). Wells completed in unconsolidated valley fill typically produce sufficient water for domestic use for a single household but probably would not be sufficient for public supply because of the poorly sorted unconsolidated valley fill.

The consolidated rocks and unconsolidated valley fill form a heterogeneous, anisotropic, interconnected ground-water system. Ground-water withdrawals from consolidated rocks can affect water levels in the overlying unconsolidated valley fill. Also, residential development and other human activities can affect water quality in the unconsolidated valley fill and underlying consolidated rocks.

The complex geology (as discussed briefly in the "Hydrogeology" section of this report) and lack of spatially distributed water-level data make determining the connection between different consolidated rocks and between the consolidated rocks and unconsolidated valley fill difficult, but any water that is removed from one part of the system is no longer available to move to other parts of the system. Therefore, well withdrawals have an effect in the overall hydrologic system. In a homogeneous ground-water system, effects of ground-water withdrawals typically appear as declining water levels in nearby wells. The heterogeneity of a consolidated-rock ground-water system often results in inconsistent water-level declines. During testing and observation, water levels may be affected at only one of several observation wells, even though the observation wells may be the same distance from the pumped well. If that one well is not measured or no well is within the zone of water-level decline, no effects would be observed. In addition to causing water-level declines, ground-water withdrawals decrease ground-water discharge to streams and springs. These effects cannot be estimated by standard equations and models because of heterogeneity and uncertainty in fracture flow and may only be noticed by long-term monitoring of an extensive network of wells, springs, and streams.

Aquifer Characteristics

The rate at which water can move through unconsolidated valley fill or consolidated rock is proportional to the hydraulic conductivity of the fill or rock (Fetter, 1980, p. 473). Transmissivity is hydraulic conductivity multiplied by saturated aquifer thickness. For a well, saturated thickness is often assumed to be the perforated or open interval of the well. The amount of water a well can yield and the amount of ground water that can flow through a cross-sectional area are dependent upon transmissivity.

Hydraulic conductivity of unconsolidated valley fill is typically related to grain size and grain-size distribution. The hydraulic conductivity of a well-sorted gravel can be six orders of magnitude greater than the hydraulic conductivity of a fine, silty sand (Fetter, 1980, table 4.4). On the basis of specific capacities obtained from 13 drillers' logs and one aquifer test, estimates of the hydraulic-conductivity value for the unconsolidated valley fill in the study area range from 0.1 to 60 ft/d (Holmes and others, 1986, p. 19). Specific capacity is the rate of discharge of water from a well divided by the drawdown in the well for a specific time. On the basis of slug-test analysis of wells completed in fine sand, silt, and mixtures of sand, silt, and clay, estimates of the hydraulic-conductivity value of the unconsolidated valley fill in the Prospector Square area range from 1 to 14 ft/d (Mason, 1989, p. 22). No large production wells withdraw water from unconsolidated valley fill in the study area, and hydraulic conductivity of the unconsolidated valley fill was not determined at additional locations during this study.

With the possible exception of the Nugget Sandstone, consolidated rocks in the study area have little primary porosity (see "Hydrogeology" section of this report), and hydraulic conductivity is probably related to the number and size of fractures or solution openings. In the study area, the amount of withdrawal from a well depends on the well intersecting water-bearing fractures (Baker, 1970, p. 18). Holmes and others (1986, p. 29) report that vertical movement of water through the consolidated rocks probably is more prevalent than horizontal movement. Because many of the fractures and faults are nearly vertical, vertical hydraulic conductivity probably is larger than horizontal hydraulic conductivity. Also, in some places, gouge associated with faults may impede the horizontal movement of water. In areas of vertical or nearly vertical fractures, one well may intercept fractures and a nearby well may not. Baker (1970, p. 18) reports that the large discharge of water from the Weber Quartzite from mine tunnels near Park City should not be taken as an indication of the potential yield of wells. The tunnels drain many miles of mine workings that intersect fractures. Wells drilled in the Weber Quartzite may intersect only a few fractures.

Holmes and others (1986, table 6) estimated hydraulic properties of the consolidated rocks from aquifer tests of eight wells and one mine tunnel. The transmissivity values ranged from 3 to 7,400 ft^2/d. The highest transmissivity was in the Thaynes Formation and the lowest was in the extrusive igneous rocks. Because of the complex system of faults and fractures in the consolidated rocks, transmissivity might vary greatly both locally and throughout the study area. Results of aquifer tests, therefore, cannot be applied to entire formations and may not be representative even at nearby wells.

Several problems prevented the determination of additional transmissivity values for consolidated rocks during this study. One of the primary problems was that pumping activities at municipal wells could not be modified to assure proper aquifer-test data collection. Most of the problems with determining transmissivity of the consolidated rocks, however, relate to the fractured-rock geology. Aquifer-test analysis typically involves matching test data to theoretical curves. The most common curve to match is based on the Theis equation. The Theis equation assumes that (1) the aquifer is homogeneous and isotropic, (2) the water body has infinite areal extent, (3) the discharging well penetrates the entire thickness of the aquifer, and (4) the water removed from storage is discharged instantaneously with decline in head (Lohman, 1972, p. 15). The assumptions imply that water moves radially from all directions along horizontal flow paths and that the unit has a constant storage coefficient. Fractured rock is not homogeneous or isotropic, and most flow is along fractures oriented along specific directions. Drawdown during aquifer tests can exceed 300 ft, causing a steep vertical gradient around the well. Faults, stratigraphic contacts, or low-permeability zones cause the water body to not have infinite areal extent. Variations of the Theis assumptions have been developed to allow for vertical flow into the aquifer from higher or lower units and delayed yield from storage. Data available from wells in the study area do not match any of these type curves, and transmissivity could not be determined accurately. With sufficient flow-rate and water-level data in the discharging well and appropriately located observation wells, computer models could be constructed to analyze such aquifer tests. During well development, aquifer testing, and source-protection testing, water companies and consultants have estimated transmissivity, even though the Theis and other assumptions have not been met. Most of these data are only for the pumping well, and the well may have pumped for less than 24 hours. Information about tests done by water companies and consultants can be obtained from the water companies or the Utah Department of Environmental Quality, Division of Drinking Water.

The amount of water-level fluctuation for a given amount of recharge or discharge is inversely proportional to the specific yield of unconfined unconsolidated valley fill or consolidated rocks, or to the storage coefficient of confined consolidated rocks. Specific yield is the amount of water yielded from water-bearing material by gravity drainage, as occurs when the water level declines (Lohman, 1972, p. 6), and represents a dewatering of the pores. Storage coefficient is the volume of water a confined aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head (Lohman, 1972, p. 8). The storage coefficient does not represent dewatering, but rather the secondary effects of water expansion and aquifer compression caused by changes in water pressure.

Water in most of the unconsolidated valley fill in the study area is unconfined. Fetter (1980, table 4.2) reports the average specific yield of unconsolidated deposits ranges from 0.02 for clay to 0.27 for coarse sand. On the basis of descriptions of materials reported in drillers' logs, Holmes and others (1986, p. 21) determined the average specific yield of the unconsolidated valley fill in the study area to be about 0.15.

Water in the consolidated rocks in much of the study area is unconfined but becomes confined at lower altitudes where the consolidated rock is overlain by unconsolidated valley fill or a less permeable zone of consolidated rocks. Flowing wells completed in the Nugget Sandstone in Snyderville Basin indicate confined conditions. Holmes and others (1986, p. 30) report that the specific yield or storage coefficient of the consolidated rocks could not be determined from aquifer tests but report that storage-coefficient values determined by others range from 0.0004 to 0.013 (Holmes and others, 1986, table 6). The larger values probably represent specific yield of unconfined consolidated rock.

Specific yield and storage coefficient could not be determined during this study because of the problems discussed earlier in this section. The lack of suitable observation wells was another complication. Because transmissivity and specific yield are variable and difficult to determine in consolidated rocks, it is difficult to estimate the effect of additional ground-water withdrawals from specific areas. Some indication of the effect of ground-water withdrawals on water levels and spring discharge can be obtained during well-yield tests. Future tests would provide more information.

An aquifer test in February 1996 used well (D-1-3)12cca-1, completed in the Thaynes Formation, as the pumped well. The pumping rate ranged from 600 to 1,200 gal/min noncontinuously for about 48 hours. The water level in well (D-1-3)11ddb-1, also completed in the Thaynes Formation about 1,500 ft away from the pumped well, was affected within 25 minutes. Water levels in five other observation wells completed in the Thaynes Formation 2,000 ft to 3,600 ft away from the pumped well were not affected. The other observation wells were (D-1-3)11dbc-1, (D-1-3)11dbd-1, (D-1-3)11ddb-1, (D-1-3)12cca-1, and (D-1-3)13abb-2. These wells may have been unaffected because of the short pumping time, preferred flow direction in fractures, or because faults in the area act as hydrologic boundaries. Because the water-level data match no theoretical curve, aquifer characteristics could not be determined. The rapid reaction in well (D-1-3)11ddb-1, however, indicates that the storage coefficient is small.

An aquifer test near Kimball Junction in 1985 used well (D-1-4)19bdb-1, completed in the Twin Creek Limestone, as the pumped well. The pumping rate was about 850 gal/min for 24 hours. The water level in well (D-1-4)19bbc-2, completed in the Twin Creek Limestone 1,200 ft from the pumped well, was affected within 5 minutes. The rapid water-level decline in this observation well indicates a small storage coefficient, but because the water-level data match no theoretical curve, the values for aquifer properties could not be determined. The water level in well (D-1-4)19aba-1, completed in the Twin Creek Limestone and unconsolidated valley fill 2,400 ft from the pumped well, was not affected. This might indicate that the pumped well does not affect water levels 2,400 ft away during a pumping period of 24 hours, or that the Mount Raymond-Absaroka thrust fault (pl. 2) acts as a hydrologic boundary between the two wells. Water levels in three wells completed in the Nugget Sandstone 1,900 to 3,000 ft from the pumped well were not affected. This indicates that pumping for 24 hours from the Twin Creek Limestone does not affect water levels in the Nugget Sandstone at the distance of the observation wells. Horizontal water movement between the formations would be required to affect the wells completed in the Nugget Sandstone. Vertical movement is probably more prevalent locally than horizontal movement (Holmes and others, 1986, p. 29). If the Nugget Sandstone is present below the Twin Creek Limestone at the pumping site, water levels there may have declined. A spring discharging about 100 gal/min from unconsolidated valley fill and about 30 ft from the pumped well ceased flowing within 3 minutes from the start of the test. The cessation of flow indicates that withdrawals from the Twin Creek Limestone affect water levels in overlying valley fill. Well (D-1-4)19bbc-2 was used as a production well in 1995 and affected water levels in well (D-1-4)19bdb-1 and in well (D-1-4)19bca-2, completed in the Twin Creek Limestone and unconsolidated valley fill at a distance of about 700 ft. Well (D-1-4)19bca-2 was not monitored during the 1985 aquifer test. Water-level fluctuations in this area are shown in figure 5. Water-level measurements in well (D-1-4)19aba-1, about 3,000 ft from well (D-1-4)19bbc-2, do not indicate a water-level decline from May 1994 to May 1995.

An aquifer test in the Park Meadows area in 1988 used well (D-2-4)8aaa-1, completed in the Thaynes

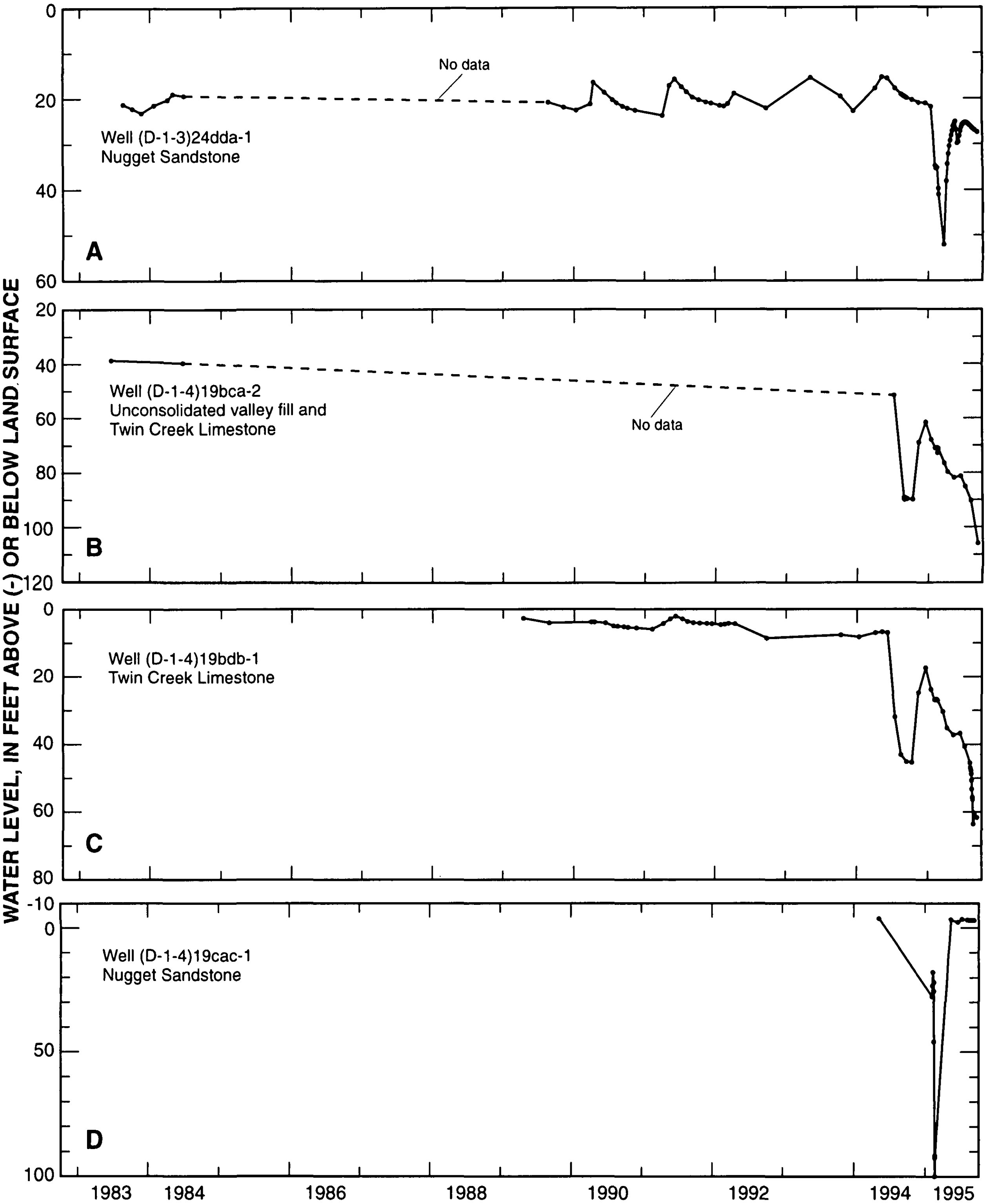


Figure 5. Water-level fluctuations in selected wells near Kimball Junction, Utah, 1983 to 1995.

Formation, as the pumped well (Mason, 1989, p. 25). The test determined that pumping this well decreased water levels in nearby wells completed in the unconsolidated valley fill just above the Thaynes Formation, eliminated discharge from spring (D-2-4)4dca-S1, Dority Springs, and may have decreased discharge from the unconsolidated valley fill and the Thaynes Formation into the Pace-Homer Ditch. The test also determined that pumping of this well did not affect water-levels in the Woodside Shale or water levels in unconsolidated valley fill overlying the Woodside Shale. The Woodside Shale is south of the Thaynes Formation in the area (Bryant, 1990, sh. 1).

An aquifer test in 1989 used well (D-2-4)4dda-1, completed in the Thaynes Formation, as the pumped well. Water levels in well (D-2-4)3dba-2, reportedly completed in the Thaynes Formation, and two other wells completed in unconsolidated valley fill just above the Thaynes Formation were affected (James M. Montgomery Engineers, 1990). Water levels in well (D-2-4)8aaa-1 were not affected and discharge from spring (D-2-4)4dca-S1 was not affected.

Water-Level Fluctuations

Water-level fluctuations in the study area are caused by fluctuations in ground-water recharge and ground-water withdrawals. Water-level fluctuations from 1983 to 1995 are shown in figures 5 and 6 and water-level fluctuations from 1993 to 1995 are shown in figure 7. Annual water-level fluctuations (A, B, D, F, I, and J in fig. 6) are caused by annual variation in ground-water recharge and are directly related to annual variation in precipitation. The relative amount of annual variation in water level is indicative of the storage coefficient of the formation in which the well is completed. Water-level fluctuations in wells completed in unconsolidated valley fill and Nugget Sandstone generally are smaller than water-level fluctuations in wells completed in consolidated rocks such as the Twin Creek Limestone, the Thaynes Formation, and shales and volcanic rocks in the study area. Water levels in unconsolidated valley fill also may fluctuate less because they are controlled by interaction with surface water.

In general, water levels in the study area have not changed significantly from 1983 to 1995. A statistical analysis using the Wilcoxon Signed-Rank test (Ott, 1993, p. 297) indicated no significant difference between the highest water level measured in 19 wells during 1983 and 1984 (Holmes and others, 1986, table 5) and the highest water level measured in the same wells during 1994 and 1995 (Downhour and Brooks, 1996, table 3). In some areas, however, high levels in 1995 were lower than high levels in 1983 and 1984 (B, E, and F in fig. 6). Continued water-level monitoring would help determine how much of this change is caused by decreased precipitation and how much is caused by increased ground-water withdrawals in these areas. The water-level increase in well (D-1-4)20dab-2 (C in fig. 6) may be a response to decreased nearby pumping. In 1993, a production well near this well was replaced by a well farther away.

Increased ground-water withdrawal from 1983 to 1995 generally has not affected ground-water levels in the study area, probably because ground-water withdrawal is a minor part of the ground-water budget (see "Water-budget analysis" section of this report). Seasonally and in some areas, however, ground-water withdrawal affects ground-water levels. In the Kimball Junction area (fig. 5), increased withdrawal for testing and production in 1994 and 1995 caused seasonal fluctuations and also caused peak water levels in 1995 to be lower than peak water levels in 1994, despite much-greater recharge in 1995.

Water levels in consolidated rocks generally increase from March through May or June, decrease throughout the summer and fall, and remain low during the winter (figs. 6 and 7). The water-level rise from March through May is caused by recharge from snowmelt and rainfall. The increased pressure head causes increased ground-water discharge to streams and springs. When recharge stops, the discharge continues until water levels are lowered.

Rapid snowmelt and fractures that allow the water to reach the ground-water system probably cause the rapid water-level changes observed in some wells (B in fig. 6, and B and F in fig. 7). Water levels in well (D-2-4)8dbd-3 (H in fig. 7) represent consolidated-rock discharge to unconsolidated valley fill. The small water-level increase in February and March 1995 was probably caused by low-altitude recharge to the valley fill. The larger water-level increase in May probably is caused by increasing upward movement from the Thaynes Formation caused by high-altitude recharge to the Thaynes Formation.

Water levels in a few wells have low levels in mid-summer and increase during the fall (C and E in fig. 6). The low levels are probably caused by increased ground-water withdrawals for lawn and garden irriga-

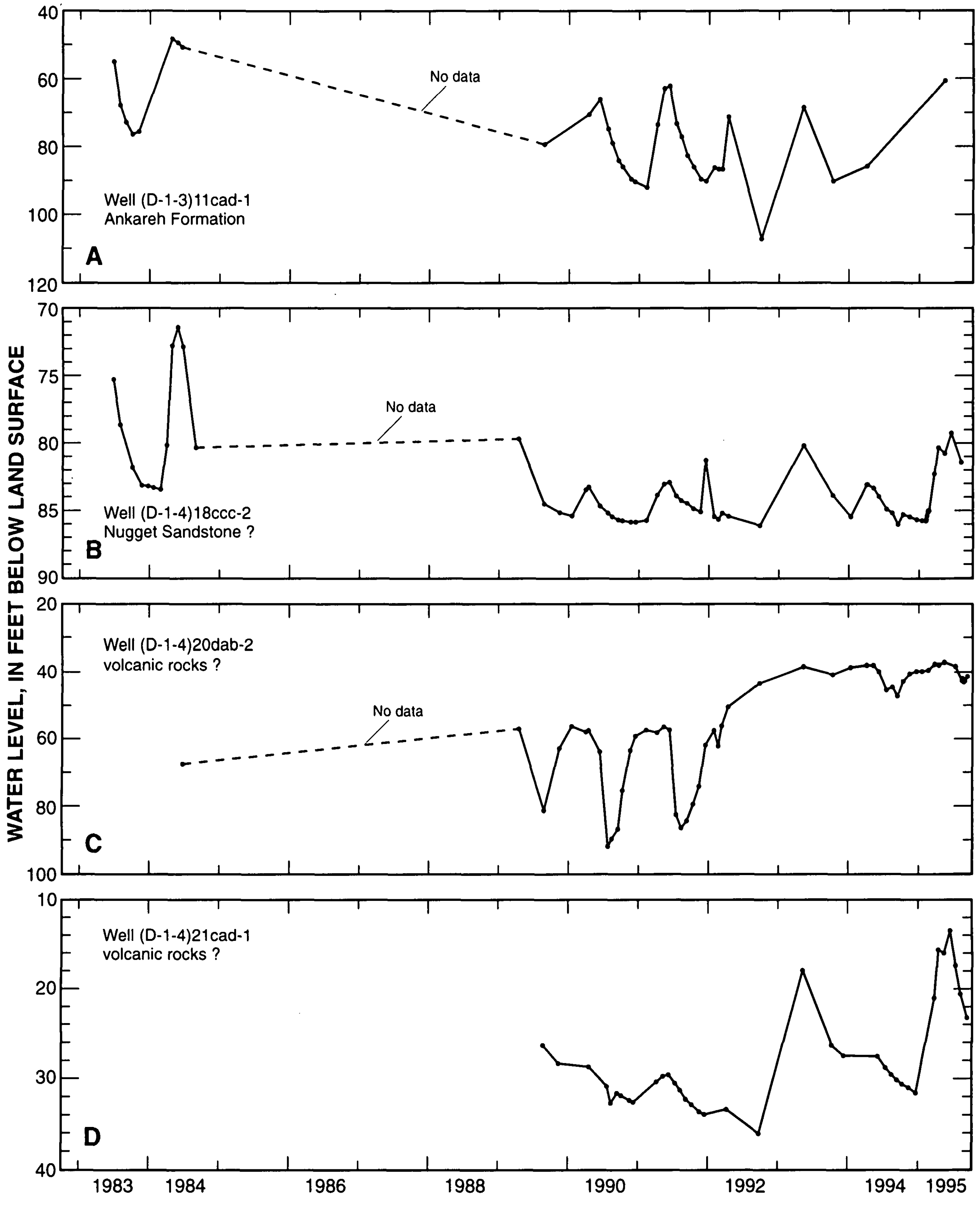


Figure 6. Water-level fluctuations in selected wells in Snyderville Basin, Park City, and adjacent areas, Utah, 1983 to 1995.

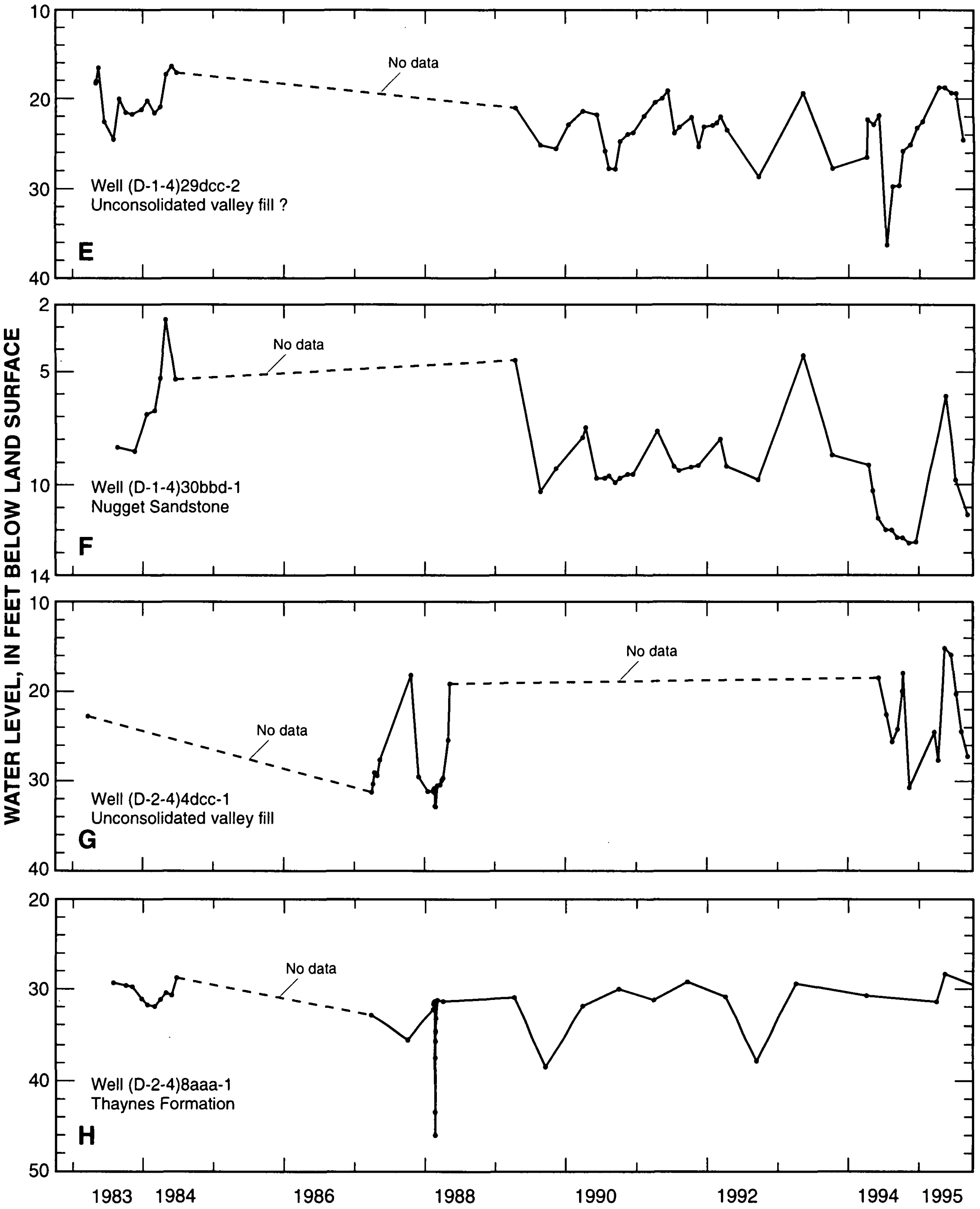


Figure 6. Water-level fluctuations in selected wells in Snyderville Basin, Park City, and adjacent areas, Utah, 1983 to 1995—Continued.

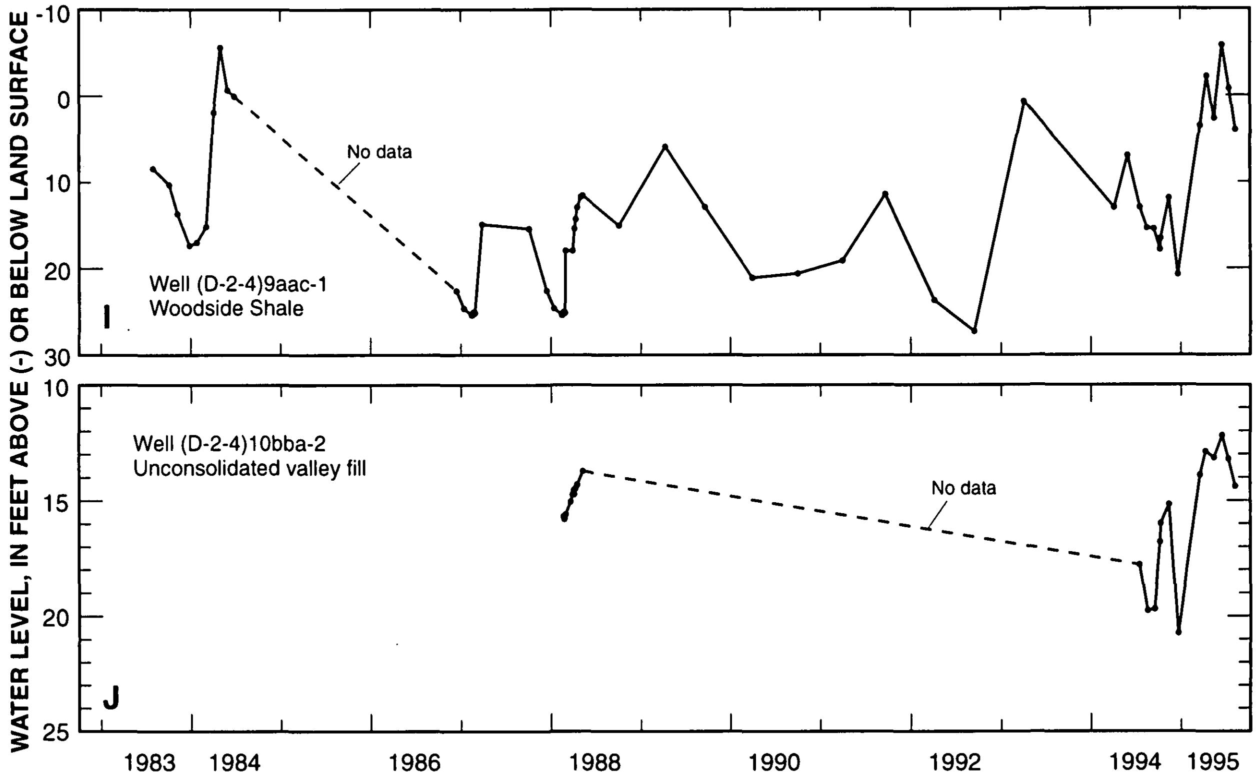


Figure 6. Water-level fluctuations in selected wells in Snyderville Basin, Park City, and adjacent areas, Utah, 1983 to 1995—Continued.

tion. The water-level rises during the fall are probably recovery from pumping and do not represent recharge to the ground-water system. Water levels in unconsolidated valley fill may also increase in the fall because of decreased discharge by evapotranspiration of crops and riparian areas and continued upward movement from consolidated rocks.

Recharge

Recharge to the ground-water system occurs through infiltration of snowmelt and rainfall, ground-water inflow from south of the study area through consolidated rock, and infiltration of streamflow, unconsumed irrigation water, and septic-tank effluent. Most of the recharge is derived from snowmelt at high altitude in the western and southwestern parts of the study area. Recharge to consolidated rock occurs in the spring after the soil veneer has thawed and become saturated, thus allowing snowmelt and stream runoff to infiltrate through the soil to the consolidated rock (Holmes and others, 1986, p. 21). This section of the report discusses sources and processes of recharge. Methods of data collection and analysis are discussed in the "Water-budget analysis" section of this report. Monthly recharge to and discharge from the ground-water system are shown in figure 8. Recharge from ground-water inflow is not shown because the monthly distribution is not known and recharge from septic tanks is not shown because it is only about 6 acre-ft per month and would be distributed evenly throughout the year.

Snowmelt

Recharge from infiltration of snowmelt is the largest source of recharge in the study area and occurs mainly from April through June (fig. 8). The amount of snowmelt available for recharge to the ground-water system is controlled mainly by four factors: the amount of water in the snowpack, the amount of water that is sublimated directly from the snowpack, the amount of water that runs off the soil surface to streams, and the amount of water that is needed to replenish soil moisture and be used by plants. The amount of water remaining after sublimation, runoff, the replenishment of soil moisture, and use by plants becomes ground-water recharge.

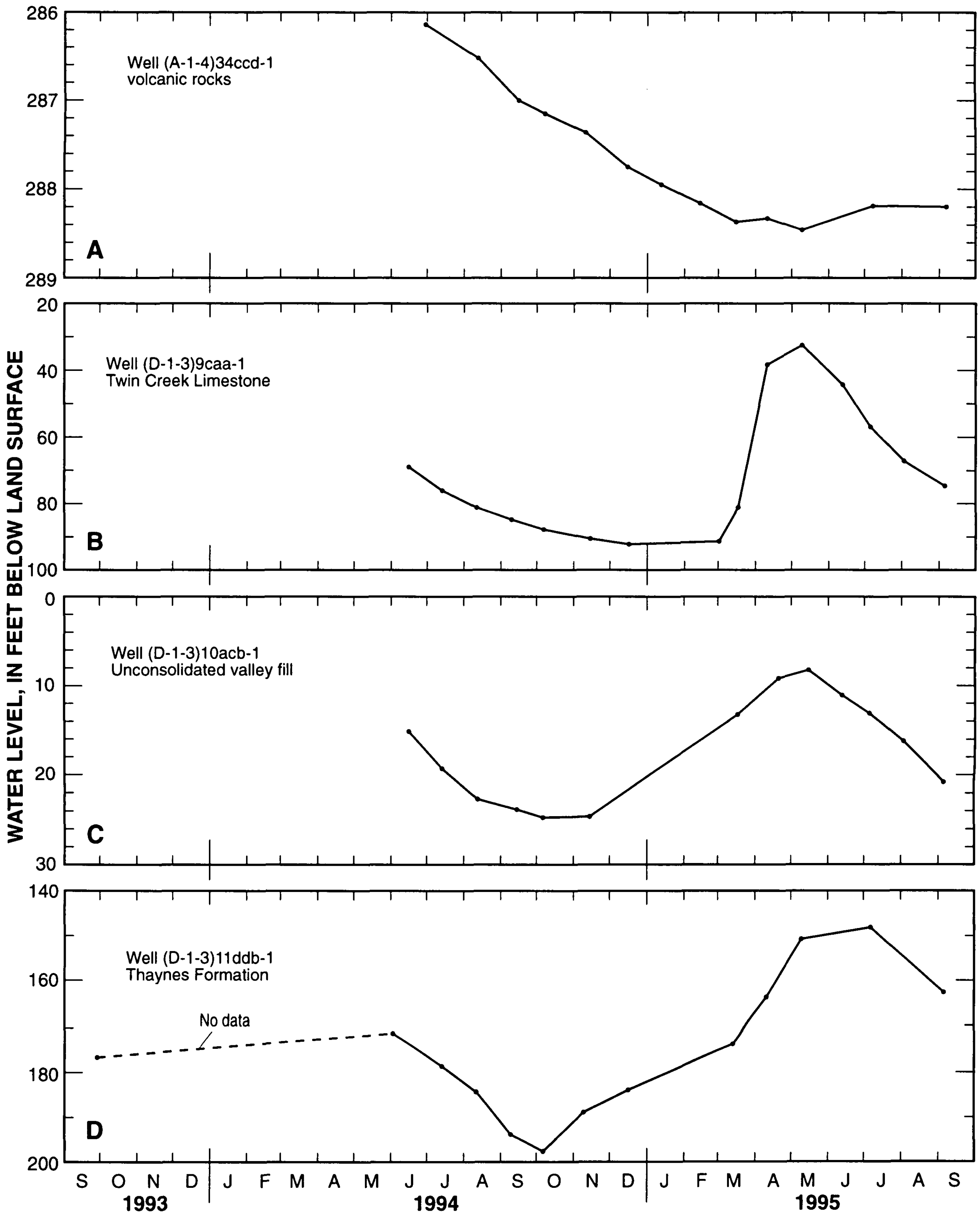


Figure 7. Water-level fluctuations in selected wells in Snyderville Basin, Park City, and adjacent areas, Utah, 1993 to 1995.

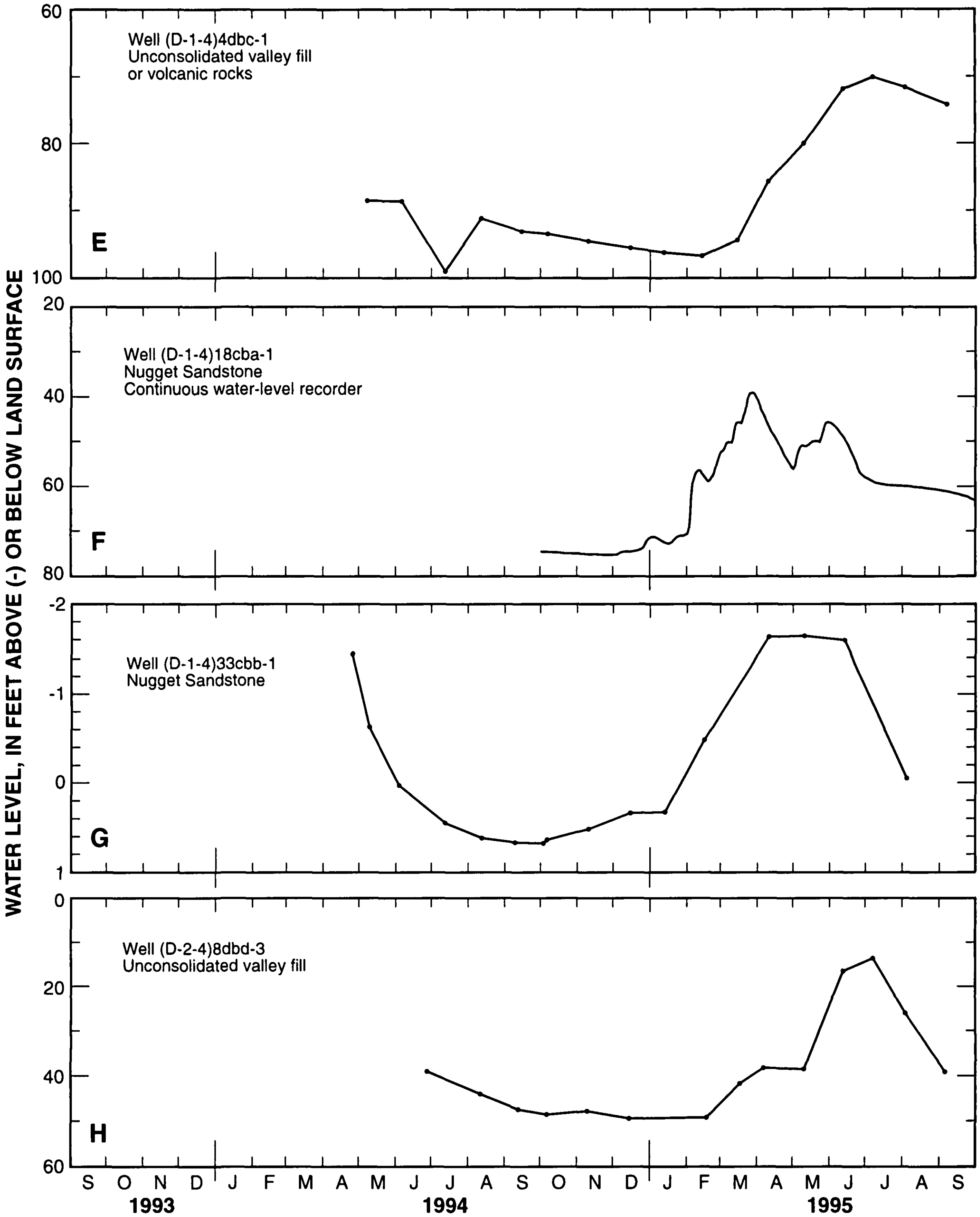


Figure 7. Water-level fluctuations in selected wells in Snyderville Basin, Park City, and adjacent areas, Utah, 1993 to 1995—Continued.

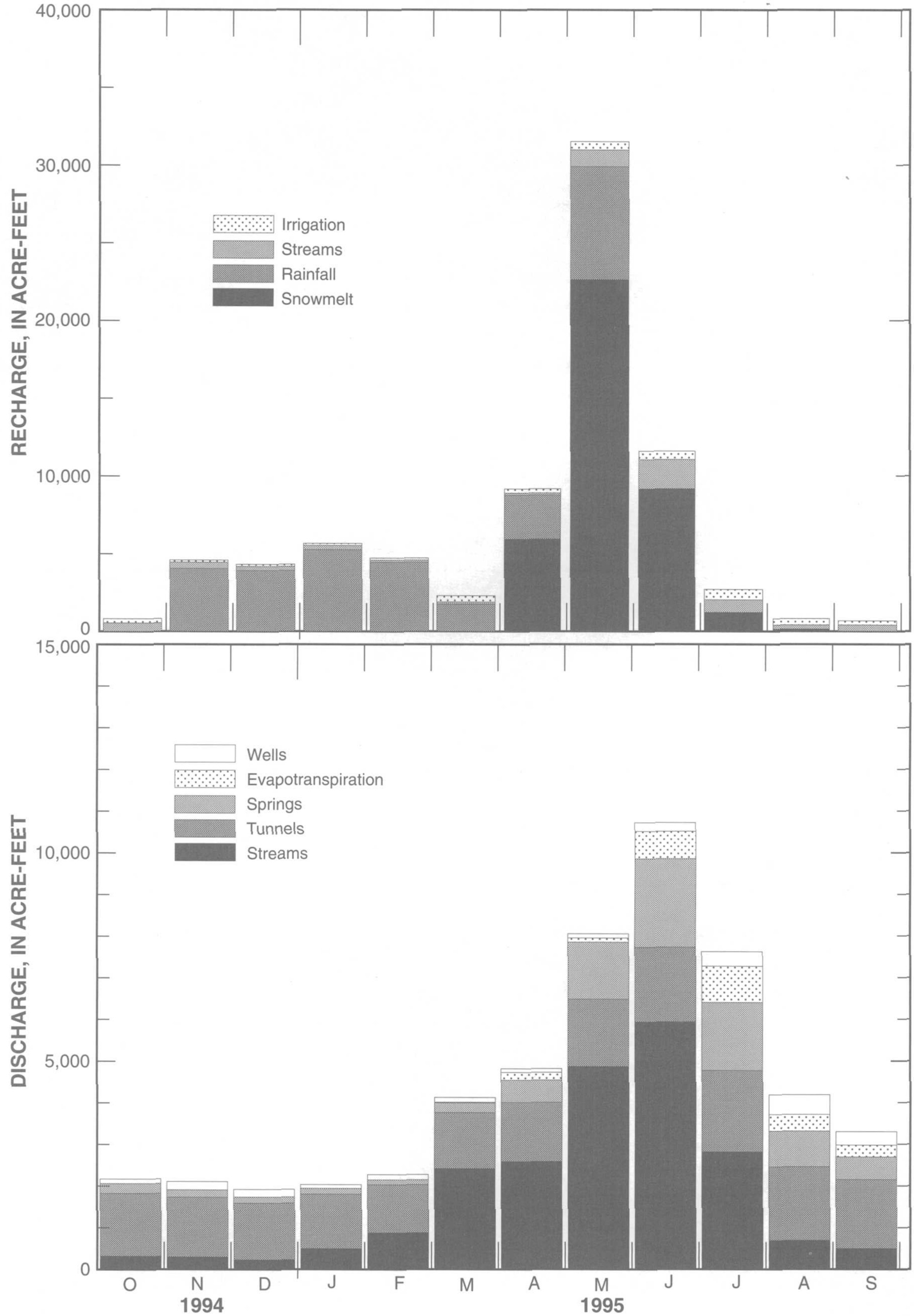


Figure 8. Recharge to and discharge from the ground-water system in Snyderville Basin, Park City, and adjacent areas, Utah, 1995.

Rainfall

Recharge from infiltration of rainfall is the second largest source of recharge in the study area and occurs mainly from November through May (fig. 8). Summer rainfall mostly is consumed by natural vegetation and crops or evaporates. Less rainfall is consumed by vegetation in the fall, but the rainfall mostly replenishes soil moisture. At low altitude, winter rainfall and snowmelt during winter thaws contribute ground-water recharge. At low altitudes, spring rainfall (March through June) can recharge the unconsolidated valley fill but mostly is used by vegetation. In areas where the soil is saturated, the water runs off to streams. At higher altitudes, spring rainfall adds to the water content of the snowpack or contributes directly to ground-water recharge as the water seeps through the snowpack. In areas of residential or commercial development, paving causes surface runoff to be greater and recharge from snowmelt and rainfall less than that in undeveloped areas.

Ground-Water Inflow

Ground-water inflow occurs across the southern and southwestern boundaries of the study area through consolidated rocks, mine tunnels, or fractures that intersect mine tunnels. Consolidated-rock formations that crop out southwest of the study area dip toward the east and also crop out in the Park City area (Holmes and others, 1986, p. 23). Some of the ground water south of the study area may flow in the direction of dip and enter the Park City area. Forster and Smith (1988, fig. 1) show hypothetical flow lines that indicate that topographic divides in mountainous regions are not necessarily ground-water divides.

Infiltration of Streamflow

Infiltration of streamflow generally contributes a small proportion of recharge to the ground-water system (fig. 8). This recharge occurs mostly in the western and southwestern parts of the study area and primarily to unconsolidated valley fill. Seepage to consolidated rock occurs in Red Pine Canyon and possibly White Pine Canyon, Willow Draw, and Toll Canyon. If streams in these canyons are channelized or enclosed in pipe as part of residential or commercial development, ground-water recharge from streams may decrease.

Ground-water recharge from streams fluctuates seasonally and annually with precipitation and ground-water levels. In areas where the ground-water level is always less than the streambed altitude, such as Red Pine Canyon and parts of White Pine Canyon, Willow Draw, and Toll Canyon, the stream will lose water to the subsurface as it flows across permeable deposits of fractured consolidated rock and unconsolidated valley fill. The amount of infiltration varies with the level in the stream. In areas where the ground-water level fluctuates above and below the altitude of the streambed, streamflow infiltrates to the ground-water system typically only in the late summer through winter months, when ground-water levels are lowest. This form of recharge generally occurs where streams overlie unconsolidated valley fill and primarily occurs near Park City and along the lower parts of Silver Creek and East Canyon Creek. If ground-water withdrawal from wells lowers ground-water levels near streams to below the altitude of the streams, streamflow infiltration to the ground-water system will increase. The amount of recharge depends on the gradient from the streams to the ground water and on the permeability of the deposits underlying the streams.

Infiltration of Irrigation Water

Recharge from infiltration of irrigation water contributes a small proportion of recharge to the unconsolidated valley fill (fig. 8). Irrigation water recharges the ground-water system when the amount of water that reaches the root zone exceeds the consumptive use of the plants. Most of this recharge occurs along lower Silver Creek, where water from the creek is used to flood irrigate pasture. Irrigation along lower Silver Creek recharges the ground-water system throughout most of the year because water is allowed to flow across fields during all months. During winter months, some of this flow may directly re-enter Silver Creek because the frozen ground prevents infiltration.

Direction of Flow

Ground-water flow in the study area is of four types, similar to those described by Mayo and others (1992, p. 244) for the central Wasatch Range. The applicable types of ground-water flow include shallow and deep flow through fractures in limestone, sandstone, and quartzite; shallow and deep flow between consolidated-rock formations along faults; artificially induced ground-water flow toward mine tunnels; and shallow ground-water flow through unconsolidated valley fill.

Ground-water flow in consolidated rock generally is from recharge areas at higher altitudes in the mountains to discharge areas at lower altitudes in the valleys where water is withdrawn from wells, discharges from springs, or seeps upward into overlying unconsolidated valley fill (pl. 2). Ground water flows along paths of least resistance from a higher to a lower altitude or potential. Ground water, therefore, moves preferentially along fractures and joints rather than through tight interstices of the consolidated rocks. Orientation, size, and degree of connection of the fractures determines the ability of the rock to transmit water. If fracture orientation is not similar to the direction of the hydraulic gradient within a consolidated-rock formation, then the ground-water flow path is tortuous and depends on the degree of fracture connection. If fracture orientation is similar to the direction of the hydraulic gradient, then ground-water flow is less tortuous and residence times should be shorter. If a major fault transects a formation or is between formations, then ground-water flow can be restricted in a direction perpendicular and enhanced in a direction parallel to the fault (Ashland and others, 1996, p. 55).

The network of mine shafts and tunnels in the mining area south of Park City probably has changed the direction of ground-water flow in the Weber Quartzite and adjacent consolidated rocks. No water-level data from wells are available to delineate the present or prior-to-mining-development hydrologic condition; however, artificial hydraulic gradients probably have been imposed as consolidated rock has been dewatered through ground-water flow toward and discharge to these tunnels.

Ground-water flow through unconsolidated valley fill is generally from areas of recharge near consolidated rocks to areas of discharge near streams. Near a stream, ground water generally flows along a gradient similar to the stream gradient. Ground water preferentially moves through areas of well-sorted sands and gravels rather than through clays or poorly sorted deposits.

Ground-water flow directions were determined using the altitude of springs and water levels measured in 49 wells during October 3-7, 1994 (Downhour and Brooks, 1996, table 3). Because water levels can change rapidly in consolidated rock, water-level measurements only from a short time period must be used for comparative purposes. Water levels tend to be more stable during late fall and early winter months than during late winter through summer months when recharge and increased pumping occurs. The water levels measured in October 1994 represent point data at many locations and in several rock types. Because of the lack of spatially distributed wells, of which only some were suitable for measuring water levels, the general direction of flow could be determined only in the Thaynes Formation near Park City, the Nugget Sandstone near Snyderville, consolidated rocks near Pinebrook and Summit Park, and the unconsolidated valley fill in the vicinity of Snyderville and McLeod Creek.

Water levels in wells measured during October 1994 and the altitude of springs (D-2-4)8cab-S1, (D-2-4)8dab-S1, and (D-2-4)4dca-S1 indicate that ground-water flow in the Thaynes Formation near Park City generally is in a northeasterly direction from the higher-altitude recharge area to a lower-altitude discharge area in the Park Meadows area (pl. 2). In late spring and early summer, when water levels are high in the Thaynes Formation and the two municipal wells are not pumped, some water probably flows upward into the unconsolidated valley fill beneath McLeod Creek and the Park Meadows area. Water levels also indicate an upward gradient from the Woodside Shale to the overlying unconsolidated valley fill in the Prospector Square area. Withdrawal from municipal wells in the Park Meadows area lowered water levels, decreased discharge from spring (D-2-4)4dca-S1 (Dority Springs), and caused downward flow of water from the overlying unconsolidated valley fill into the Thaynes Formation (Mason, 1989, p. 25-33).

In general, water in the Nugget Sandstone near Willow Draw, which is in the lower block of the Mount Raymond-Absaroka thrust fault, flows east to the area around Silver Springs (pl. 2). Water levels in the Nugget Sandstone south and southeast of Snyderville indicate northeasterly flow. This would indicate flow from the White Pine Canyon area to the area around Snyderville. Water in the Nugget Sandstone probably flows through unconsolidated valley fill to discharge at springs and streams, or may remain in the consolidated rock and fractures and flow northeasterly.

Water in consolidated rocks near Summit Park and Pinebrook generally flows northeasterly (pl. 2). Folds, faults, and the lack of spatially distributed water-level data make the delineation of ground-water flow paths difficult. Many wells in the area are used regularly, and are therefore not suitable for water-level measurements. Water that infiltrates into the Thaynes Formation at higher altitudes probably is confined by the shales on the west and east as it flows toward a lower-altitude discharge area. The Toll Canyon fault might restrict the northeasterly flow in the Thaynes Formation but also might enhance mixing with water from other formations. Water in the Ankareh Formation northwest of and adjacent to the Thaynes Formation (Bryant, 1990, sh. 1) might be flowing northeasterly through near-vertical bedding planes. Water levels measured in May 1993 and April 1994 indicate that an upward gradient might exist in the Ankareh Formation in this area.

Water in the sandstones and conglomerates of Tertiary age in the northeastern part of the study area, north of Silver Creek Junction, probably flows south toward Kimball Creek or east toward Silver Creek. The lack of suitable monitoring wells for water-level data in this area prevents the delineation of distinct ground-water flow paths. Similarly, in the northwestern part of the study area, north of Interstate Highway 80, water-level data is available only in the Nugget Sandstone, where the hydraulic gradient follows the strike direction to the south toward East Canyon Creek. No wells are known to be completed in the north-dipping Cretaceous rocks in the northwestern part of the study area. Some water might follow the strike direction to the west toward East Canyon Creek as it exits the study area.

Water levels in October 1994 in the unconsolidated valley fill around Snyderville indicate northeasterly flow and flow away from Willow Creek toward McLeod Creek and Kimball Creek. This movement appears to be different than the potentiometric surface shown by Holmes and others (1986, fig. 6), which may be influenced by an incorrect land-surface altitude at one well. Correction of the altitude data used to construct the potentiometric contours shown by Holmes and others (1986, fig. 6) shows a potentiometric surface similar to that discussed above. Water-level contours for the Nugget Sandstone and the unconsolidated valley fill indicate an upward gradient from the Nugget Sandstone to the unconsolidated valley fill near Snyderville. A set of wells completed in the unconsolidated valley fill and in the Nugget Sandstone was not available to verify possible upward flow.

Ground water also flows upward from consolidated rock into the unconsolidated valley fill in other parts of the study area. This upward flow helps sustain streamflow and riparian areas. Seasonally, ground water may flow downward from unconsolidated valley fill to underlying consolidated rocks. Streamflow measurements indicate sections of gain or loss that may result from discharge from consolidated rock through unconsolidated valley fill to streams, or recharge from streamflow through unconsolidated valley fill to consolidated rock. These ground-water/stream interactions were determined during the water-budget analysis and are discussed in the appropriate subbasin section of the "Water-budget analysis" section of this report. Water-quality data indicate that water from the unconsolidated valley fill is flowing downward to the Twin Creek Limestone near Kimball Junction and from the unconsolidated valley fill downward to the Thaynes Formation near Park City as discussed in the "Ground-water quality" section of this report. Nested wells completed at various depths, in which to determine vertical hydraulic gradients between consolidated rocks, unconsolidated valley fill, and streams, do not exist.

Discharge

Discharge from the ground-water system occurs as seepage to streams, discharge to mine tunnels and springs, evapotranspiration, withdrawal from wells, and possible ground-water flow out of the study area. Most discharge occurs in the southern and western parts of the study area. Except for wells, discharge varies naturally with seasons (fig. 8). Generally, the highest rates of discharge occur during late spring and summer and the lowest rates in late winter, before snowmelt begins. The rapid increase in discharge that results from the recharge effects of snowmelt is indicative of a ground-water system with little storage. This increase in discharge is a pressure response to the infiltration of snowmelt into the ground-water system and is not direct discharge of newly melted snow.

Seepage to Streams

Ground-water seepage to streams is the largest component of ground-water discharge in the study area and occurs from unconsolidated valley fill when ground-water levels are at higher altitudes than stream levels. Seepage occurs as flow directly into stream channels from stream banks or vertically through the streambed and as diffuse ground-water discharge from small ungaged springs and riparian areas from which cumulative flow enters streams. Ground-water seepage to streams varies seasonally, with most of the discharge occurring in spring and early summer when ground-water levels are highest (fig. 8).

Seepage from unconsolidated valley fill into streams occurs primarily in the upper reaches of Silver

Creek through Park City and the Park Meadows area, the lower reaches of Silver Creek near Silver Creek Junction, Kimball Creek, most of McLeod Creek, the unnamed creek north of Silver Springs, and most of East Canyon Creek downstream from Kimball Junction.

Ground-water seepage to streams is a major component of surface-water outflow from the study area. If ground-water seepage to streams is reduced by lowering ground-water levels, surface outflow could be reduced unless replenished from other sources. An example is ground-water withdrawal for public supply. The withdrawals may lower ground-water levels and cause less ground-water seepage to streams, but much of the water withdrawn enters the surface-water or ground-water system after use. In the study area, about 50 percent of the water used for public supply re-enters streams as treated wastewater and about 10 percent is consumed by lawns, gardens, and domestic use. The remaining 40 percent recharges the ground-water system, enters streams after irrigation of lawns and gardens, or is unaccounted for as a result of measurement and estimate errors.

Mine Tunnels

Three mine tunnels discharge ground water in the study area. Baker (1970, p. 18) reports that most of the water in the mine workings around Park City appears in tunnels that penetrate the Weber Quartzite. The Ontario #2 Drain Tunnel discharges into the Provo River drainage southeast of the study area. Some of the discharge from the Spiro Tunnel is used for public supply and the remainder forms the beginning of McLeod Creek or is diverted into the Silver Creek drainage. All of the discharge from the Judge Tunnel is used for public supply. Mine-tunnel discharge varies seasonally and annually, but not to the same degree as seepage to streams or discharge from springs (fig. 8). This may be because the mine tunnels can fill with water, which is then released gradually through bulkheads and portals or is pumped to the Ontario #2 Drain Tunnel.

Springs

Five springs discharged more than 200 acre-ft each in 1995. The five springs are in the southern, southwestern, and western parts of the study area (pl. 1) and discharge water primarily from the Thaynes Formation and Twin Creek Limestone. The discharge of these springs is shown in figure 9. About 30 percent of this spring discharge is typically used for municipal supply. The remainder becomes streamflow. Several small springs discharge a negligible amount of ground water and are seasonal.

Discharge from the springs varies seasonally, much like the fluctuation in streamflow, with most discharge occurring in late spring and early summer. Discharge from springs also varies annually and is proportional to the variation in yearly snowpack. The discharge of a spring is dependent upon water levels in the consolidated rocks near the spring. The extreme seasonal and annual fluctuation in spring discharge, therefore, indicates extreme seasonal and annual water-level fluctuation. Such extreme water-level fluctuations probably result from a small storage capacity of the Thaynes Formation and the Twin Creek Limestone.

Evapotranspiration

Evapotranspiration is a small component of total ground-water discharge (fig. 8). Evapotranspiration by plants directly from ground water occurs in areas of natural riparian vegetation and pasture where ground-water levels are near land surface. This use of ground water may decrease as these areas become residential or urban developments or if ground-water withdrawals decrease water levels below the root zone of the plants.

Wells

Withdrawal from wells is a small component of total ground-water discharge (fig. 8). The amount of ground-water withdrawal and its ratio to total ground-water discharge increases in the late summer months. The increase in withdrawals is caused by the need for more water to irrigate lawns and gardens during the late summer, and also by decreased spring discharge and more reliance on well withdrawals for municipal supply.

Any water withdrawn by wells affects the ground-water system. All ground water moves from a place of recharge to a place of discharge; the average rate of discharge equals the average rate of recharge; and under natural conditions previous to withdrawal from wells, aquifers are in a state of approximate equilibrium (Theis, 1940, p. 277). Withdrawal from wells, therefore, must be balanced by an increase in recharge, a decrease in natural discharge, a loss of storage in the ground-water system, or a combination of these. Recharge from precipitation cannot generally be increased. Recharge from streams can be increased by

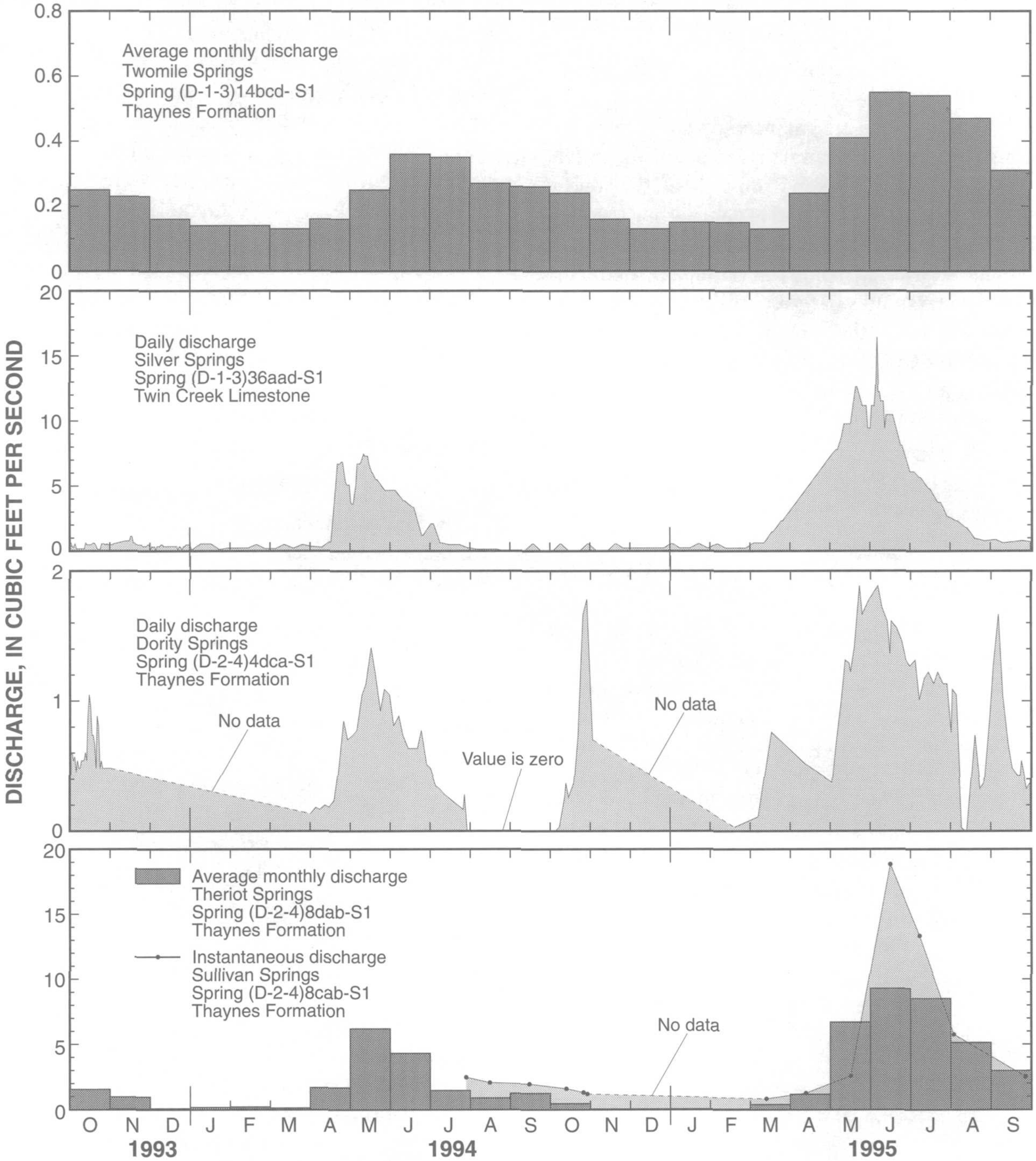


Figure 9. Discharge of selected springs in Snyderville Basin, Park City, and adjacent areas, Utah, 1993 to 1995.

streams to below the altitude of the streams. Natural discharge can be decreased by decreasing ground-water seepage to streams, ground-water discharge to springs, and ground-water use by evapotranspiration of crop and riparian areas by lowering water levels between the withdrawal area and the discharge area. Until water levels are lowered enough to increase recharge from streams or decrease natural discharge by the same amount of water withdrawn by wells, water withdrawn by wells will continue to be balanced by a loss of storage in the ground-water system and water levels will continue to decline. The amount and areal extent of water-level declines are dependent upon aquifer characteristics. Because of the heterogeneity, anisotropy, and fracture flow in the study area, aquifer characterisarea, aquifer characteristics probably vary at different locations within the same formation.

In most confined aquifers, little water is taken from storage, the cone of depression spreads rapidly, and most withdrawals are balanced by an increase in recharge or decrease in natural discharge. In unconfined aquifers, more water is taken from storage, the cone of depression spreads slowly, and ground-water levels may decline for years before reaching a new equilibrium. In the study area, many of the consolidated rocks probably have a small storage coefficient similar to that of confined aquifers, the cone of depression spreads rapidly in some directions, and withdrawals may be balanced by an increase in recharge or decrease in natural discharge.

Except in limited areas, the effect of large production wells on the ground-water system in the study area is not known. In some areas of increased withdrawals, no monitoring wells exist to measure the effects of ground-water withdrawals. Near Park City and Kimball Junction, withdrawal from wells has caused ground-water-level declines and may be reducing ground-water seepage to streams. Ground-water withdrawal from well (D-2-4)8aaa-1, completed in the Thaynes Formation in the Park Meadows area, eliminates discharge from spring (D-2-4)4dca-S1, also in the Thaynes Formation, and causes water-level declines in the overlying unconsolidated valley fill (Mason, 1989, p. 25-26). An aquifer test done by the U.S. Geological Survey near Kimball Junction in 1985 used well (D-1-4)19bdb-1, completed in the Twin Creek Limestone, as the pumped well. A spring about 30 ft from the pumped well discharging about 100 gal/min from unconsolidated valley fill ceased flowing within 3 minutes from the start of the test. The cessation of flow indicates that withdrawals from the Twin Creek Limestone affect water levels in overlying valley fill and reduce spring discharge. Other springs in the study area have not been carefully measured during withdrawals from nearby wells. Discharge from other springs would likely decrease if increased ground-water withdrawals cause a decline in the ground-water level in the area of the springs.

Ground-Water Outflow

Ground-water probably flows out of the study area in consolidated rocks and unconsolidated valley fill. Flow in consolidated rocks may enter the Wasatch Range mountain block and flow to other areas. Hypothetical flowpaths for this deep flow are shown by Forster and Smith (1988, fig. 1). Flow out of the study area also may occur to the north through the consolidated rocks north of Interstate Highway 80. The rock units dip to the north, and if the hydraulic gradient is from south to north, flow may occur in the direction of dip. Flow from the study area in unconsolidated valley fill is limited to the thin unconsolidated valley fill near the stream channels of East Canyon Creek and Silver Creek. If ground water is leaving the study area, these processes will continue regardless of annual variations in recharge that result from climate variations. In years of less-than-normal precipitation, ground water would still flow out of the study area through consolidated rocks and unconsolidated valley fill.

Ground-Water Quality

The chemical composition of ground water in the study area is influenced primarily by the lithology of the consolidated rocks through which the ground water flows. Dissolution and weathering of limestone and sandstone contribute calcium, magnesium, bicarbonate, and other constituents to the water. Dissolution of gypsum in shale or gypsiferous limestone contributes calcium and sulfate to the water. Typically, the smaller the intersices in the rock through which the water flows and the longer the water is in contact with the rock, the higher the concentration of dissolved solids in the water. In certain areas, the application and storage of road salt have influenced the chemical composition of water in the unconsolidated valley fill. Dissolution of road salt contributes sodium and chloride to much of the ground water in unconsolidated valley fill.

Water from wells generally has a dissolved-solids concentration that ranges from 200 to 600 mg/L (Downhour and Brooks, 1996, table 4). Water discharging from large springs in the Twin Creek Limestone and the Thaynes Formation generally has a dissolved-solids concentration of less than 300 mg/L. Lower dissolved-solids concentrations in water from springs than in water from wells may indicate that water moves through fractures or solution openings to most springs more rapidly and with less contact with consolidated rocks than the ground water withdrawn from wells, or that flow paths from recharge areas to springs are shorter than flow paths from recharge areas to wells, resulting in shorter contact times.

Water samples from the Spiro Tunnel bulkhead and portal have dissolved-solids concentrations that range from 540 to 760 mg/L and sulfate concentrations that range from 280 to 440 mg/L (Downhour and

Brooks, 1996, table 4). Water from the Spiro Tunnel bulkhead and portal is a calcium sulfate type. Water discharging from the Judge Tunnel has a dissolved-solids concentration of 220 mg/L and a sulfate concentration of 70 mg/L. Water from the Judge Tunnel is a calcium bicarbonate sulfate type. Holmes and others (1986, table 14) report that discharge water from the Ontario #2 Drain Tunnel had a dissolved-solids concentration of 630 mg/L and a sulfate concentration of 360 mg/L, which is considerably higher than that of water from the Judge Tunnel. Water from the Ontario #2 Drain Tunnel is a calcium sulfate type. The Judge Tunnel passes through the Weber Quartzite, whereas the Spiro Tunnel and Ontario #2 Drain Tunnel pass through the Weber Quartzite, the Park City Formation, the Woodside Shale, and the Thaynes Formation. Higher calcium and sulfate concentrations in water from the Spiro Tunnel and Ontario #2 Drain Tunnel probably are derived from the dissolution of gypsum in the additional formations through which it passes.

Water samples from wells, springs, and drains in the Park Meadows area of Park City generally have a higher dissolved-solids concentration than does ground water elsewhere in the study area. Higher sulfate and chloride concentrations cause most of the difference. Water from well (D-2-4)4dda-1 and well (D-2-4)8aaa-1, both completed in the Thaynes Formation, has a dissolved-solids concentration of about 600 mg/L and a sulfate concentration of 220 mg/L (Downhour and Brooks, 1996, table 4). Water from well (D-2-4)4dda-1 has a chloride concentration of 96 mg/L whereas water from well (D-2-4)8aaa-1 has a chloride concentration of 37 mg/L. Water from spring (D-2-4)4dca-S1, which discharges from the Thaynes Formation, has a dissolved-solids concentration of about 500 mg/L and a sulfate concentration of 210 mg/L. In both wells and the spring, the water is a calcium bicarbonate sulfate type. Water samples collected from drains installed in the unconsolidated valley fill in the Park Meadows area have dissolved-solids concentrations ranging from 680 to 830 mg/L, sulfate concentrations ranging from 220 to 330 mg/L, and a chloride concentration of 240 mg/L in one sample. Water from these drains is of a calcium sulfate or calcium sulfate chloride type. The prominence of chloride in this water is probably the result of dissolution and infiltration of road salt. The prominence of sulfate is probably the result of infiltration of water discharging from the Spiro Tunnel.

Mason (1989, p. 36) reports that the concentration of major ions varies areally and vertically within the unconsolidated valley fill near Prospector Square and that sulfate concentrations were higher than 250 mg/L in water collected from 10 wells and 2 drains near Prospector Square. The wells were all completed in unconsolidated valley fill at depths ranging from 16.5 to 95.5 ft. Mason (1989, table 10) also reports sulfate concentrations ranging from 200 to 250 mg/L in two wells completed in unconsolidated valley fill at depths of 13 and 44.5 ft near Prospector Square. Water from most of the monitoring wells and drains was a calcium sulfate type. Water from a few wells that may have been influenced by dissolved road salt from nearby snow storage was a sodium chloride type. The monitoring wells completed near the base of the unconsolidated valley fill generally yield water with low specific conductance values and pH values greater than 7.0 (Mason, 1989, p. 37). The low dissolved-solids concentrations in water derived from the base of the unconsolidated valley fill beneath the Prospector Square area might indicate that ground water in the shallow unconsolidated valley fill does not appear to have substantial downward movement even though the hydraulic gradient is downward.

Ground-water samples collected near Kimball Junction indicate that the dissolution of applied or stored road salt is affecting ground-water quality in this area. In addition to the application of road salt, the Utah State Department of Transportation and Summit County have both maintained stockpiles of road salt near Kimball Junction for many years. Low streamflow in an unnamed creek near the junction during late summer is derived primarily from unconsolidated valley fill. As discussed in the "Surface-water quality" section of this report, chloride is very prevalent in water from the unnamed creek. Water from well (D-1-4)19bab-1, completed in unconsolidated valley fill or Twin Creek Limestone, had a dissolved-solids concentration of 640 mg/L, a sodium concentration of 37 mg/L, and a chloride concentration of 250 mg/L in 1983 (Holmes and others, 1986, table 14). Water from well (D-1-4)18cda-1, completed in the Twin Creek Limestone, had a dissolved-solids concentration of 600 mg/L, a sodium concentration of 33 mg/L, and a chloride concentration of 110 mg/L in 1995 (Downhour and Brooks, 1996, table 4). The sodium to chloride mole ratio in water from wells (D-1-4)19bab-1 and (D-1-4)18cda-1 is 0.23 and 0.45, respectively. This indicates either another source of chloride or a sink for sodium. Since the high chloride concentrations occur near major roads and no other sources of chloride are known, it is assumed that sodium and chloride are derived from road salt. Ion exchange and cyclic wetting and drying may be removing sodium from the ground water (Drever, 1988, p. 240).

In comparison, water from areas not near road salt has smaller sodium and chloride concentrations. Water from well (D-1-3)16baa-1, completed in the Twin Creek Limestone, had a dissolved-solids concentration of 304 mg/L, a sodium concentration of 6.2 mg/L, and a chloride concentration of 8.7 mg/L in 1995. Water from spring (D-1-3)36aad-S1 (Silver Springs), discharging from the Twin Creek Limestone, had a dissolved-solids concentration of 210 mg/L, a sodium concentration of 5.4 mg/L, and a chloride concentration of 3.6 mg/L in 1995. The high chloride concentration in water from the wells near Kimball Junction indicates that downward movement has occurred from the unconsolidated valley fill into the underlying Twin Creek Limestone. With current and anticipated ground-water withdrawals causing water-level declines in the Twin Creek Limestone and Nugget Sandstone in this area, water with high chloride concentrations may continue to move from the unconsolidated valley fill to the Twin Creek Limestone or Nugget Sandstone.

High silica concentrations are indicative of water from the volcanic rocks. Silica concentration ranges from 45 to 60 mg/L in water from wells completed in the volcanic rocks, whereas the silica concentration is generally less than 20 mg/L in water from other wells and springs sampled in the study area (Downhour and Brooks, 1996, table 4).

Because of the complex geology and the lack of spatially distributed wells in the study area, ground-water flow paths cannot be determined solely from water-level data. In addition to standard water-quality analyses, other water-quality methods were used to help delineate probable ground-water flow paths and determine residence times in the ground-water system. These methods included the analysis of isotopes and chlorofluorocarbons, determination of normal and simple salts in conjunction with cluster analysis, and geochemical modeling to determine probable mixing.

Isotope Chemistry

Each element has a distinctive number of protons (atomic number) but can have a different number of neutrons in the nucleus, which will result in a slightly different mass. Atoms with the same atomic number but different mass are isotopes of an element. An isotope is stable if it does not undergo radioactive decay. Analysis of stable isotopes determines the ratio of a rare isotope to a common isotope and compares this ratio to a standard. Stable isotopes of sulfur ($^{34}S/^{32}S$), oxygen ($^{18}O/^{16}O$), and hydrogen ($^{2}H/^{1}H$) were analyzed during this study to help determine ground-water flow paths. Tritium (^{3}H), an unstable isotope of hydrogen, was analyzed to help determine ground-water age.

Isotopic composition is described by use of a delta value (δ), which is reported in parts per thousand or permil ($^{o}/_{oo}$) deviation from a reference standard. If the δ value is positive, then the heavy isotope is enriched relative to the standard. Conversely, if the δ value is negative, then the heavy isotope is depleted relative to the standard. Sulfur isotope values are reported relative to sulfur in troilite from the Canyon Diablo iron meteorite. Oxygen and hydrogen isotope values for water are reported relative to Vienna Standard Mean Ocean Water (VSMOW). Deviations from a standard are in the form:

$$\delta R = \left[\frac{Rsample}{Rstandard} - 1\right] \times 1,000 \qquad (1)$$

where

δR is the delta value in the water sample, in permil,

Rsample is the isotope ratio of an element in a water sample, and

Rstandard is the isotope ratio of a reference standard for the respective element.

Dissolved sulfate in water in the study area can be derived from the dissolution of gypsum or anhydrite or from oxidation of pyrite. Because of fractionation processes during the formation of pyrite, the sulfur in pyrite has a $\delta^{34}S$ value of about 0.0 permil (Faure, 1977, p. 412). Sulfur in evaporites, such as gypsum and anhydrite, formed during the Triassic and early Jurassic periods has $\delta^{34}S$ values ranging from 10 to 23 permil (Faure, 1977, fig. 21.2), but can vary depending on the amount of biological reduction prior to lithification. Precipitation in the study area has a $\delta^{34}S$ value of 6.5 permil as determined from the analysis of a composite snow sample collected in April 1995. Because snowmelt is the major component of ground-water recharge in the study area, recharge to the ground-water system probably has a $\delta^{34}S$ value similar to 6.5 permil. Seasonal and long-term variations in the $\delta^{34}S$ value of precipitation in the study area, however, are not known. Chemical interaction of the recharge water with minerals in consolidated rocks determines the $\delta^{34}S$ value of the sampled water. Dissolution of gypsum has contributed most of the sulfate if the sampled water has a $\delta^{34}S$ value greater than 6.5 permil. Oxidation of pyrite has contributed most of the sulfate if the sampled water has a $\delta^{34}S$ value less than 6.5 permil. Nielsen and Mayo (1989, p. 130) report that water from fault-controlled, large-discharge, carbonate ground-water systems in the central Wasatch Range has a mean $\delta^{34}S$ value of 7.1 permil and derives its isotopic composition from the dissolution of evaporite minerals.

Water from most mine tunnels in the central Wasatch Range has a mean $\delta^{34}S$ value of -0.5 permil and derives its isotopic composition from the oxidation of pyrite. Water collected from the Spiro Tunnel, however, has $\delta^{34}S$ values of 8.7 and 12.1 permil (Downhour and Brooks, 1996, table 4), which indicate that dissolution of gypsum is the major source of sulfate in this water. Water collected from the Judge Tunnel has a $\delta^{34}S$ value of 0.1 permil, which indicates that oxidation of pyrite is the major source of sulfate in this water. Water collected from drain (D-2-4)3ccd-2 has a $\delta^{34}S$ value of 4.0 permil. The low $\delta^{34}S$ value in this water probably can be attributed to pyrite in buried mine tailings. Water collected from well (D-1-3)16baa-1 has a $\delta^{34}S$ value of 3.3 permil. The cause of the low $\delta^{34}S$ value at well (D-1-3)16baa-1 could not be determined because this well is not along a flow path with known pyrite mineralization. All other samples collected from wells, springs, and surface water had $\delta^{34}S$ values ranging from 7.3 to 28.5 permil. This indicates that virtually all sulfate in water in the study area is derived from the dissolution of gypsum. For this reason, $\delta^{34}S$ could not be used to trace water flowing through mineralized zones in the southern part of the study area.

Because the temperature at which precipitation forms is the principal factor controlling the isotopic composition of most ground waters, isotope analyses of oxygen and hydrogen might provide insights into the recharge and geothermal history of ground-water systems (Nielsen and Mayo, 1989, p. 126). Isotope values for ^2H (deuterium) and ^{18}O are compared to a straight-line relation called the global meteoric water line. The equation for this line is

$$\delta^2 H = 8\delta^{18} O + d \qquad (2)$$

where

d is the excess ^2H parameter.

The mean value for the slope of 8 and the mean value of 10 for d were derived from about 400 water samples from rivers, lakes, and precipitation around the world (Craig, 1961). The slope of the line is constant at different locations, but the value of d can differ substantially with location. Isotope values for ^2H and ^{18}O that plot along a line with a slope other than 8 indicate that water has undergone geothermal or evaporation processes.

Values for δ^2H and $\delta^{18}O$ in water from the study area (Downhour and Brooks, 1996, table 4) plot close to the global meteoric water line (fig. 10), indicating that no significant processes have affected the water. The variation in the stable-isotope values probably reflects orographic and seasonal effects and cannot be related to geology or recharge areas. Mayo and others (1992, p. 246) report that orographic relations and microclimatic effects are responsible for the distribution of many of the observed isotopic values in the central Wasatch Range; isotopically depleted δ^2H (mean value of -134 permil) and $\delta^{18}O$ (mean value of -17.1 permil) values tend to occur on the western sides of the topographic divides and isotopically enriched δ^2H (mean value of -126 permil) and $\delta^{18}O$ (mean value of -16.0 permil) values tend to occur on the eastern sides of the topographic divides. Water from the Spiro Tunnel (Mayo and others, 1992, table 2, sample 14) had a δ^2H value of -139 permil and a $\delta^{18}O$ value of -17.7 permil, which indicate that the tunnel receives water from west-facing slopes. Isotopes collected from the Spiro Tunnel in February 1995 had δ^2H values of -131 permil and -132 permil and $\delta^{18}O$ values of -17.72 permil and -17.88 permil. The range of δ^2H for all samples collected within the study area during 1995 was from -115 to -134 permil. The range of $\delta^{18}O$ for all samples was from -15.17 to -18.00 permil. Isotope data were insufficient, however, to delineate recharge sources from western slopes.

Cooper and others (1991, p. 2171) report wide variation in $\delta^{18}O$ values between summer and winter precipitation, and even between the amount of precipitation from individual storms. They report a large difference in stream-water oxygen-isotope composition between the time of peak snowmelt and 1 month later. Water in the unsaturated soil was enriched in $\delta^{18}O$ relative to snowmelt, and rainfall was the most enriched. The use of oxygen-isotope analysis in this study area to determine recharge sources, flow paths, and residence times would require the analysis of composite snow samples and precipitation from major rain storms (such as thunderstorms during July 21-23, 1994, and rain during May 1995). In addition, surface-water samples need to be collected during peak flow and base flow, and ground-water samples need to be collected before and after snowmelt. Such intensive sampling was beyond the scope of this study.

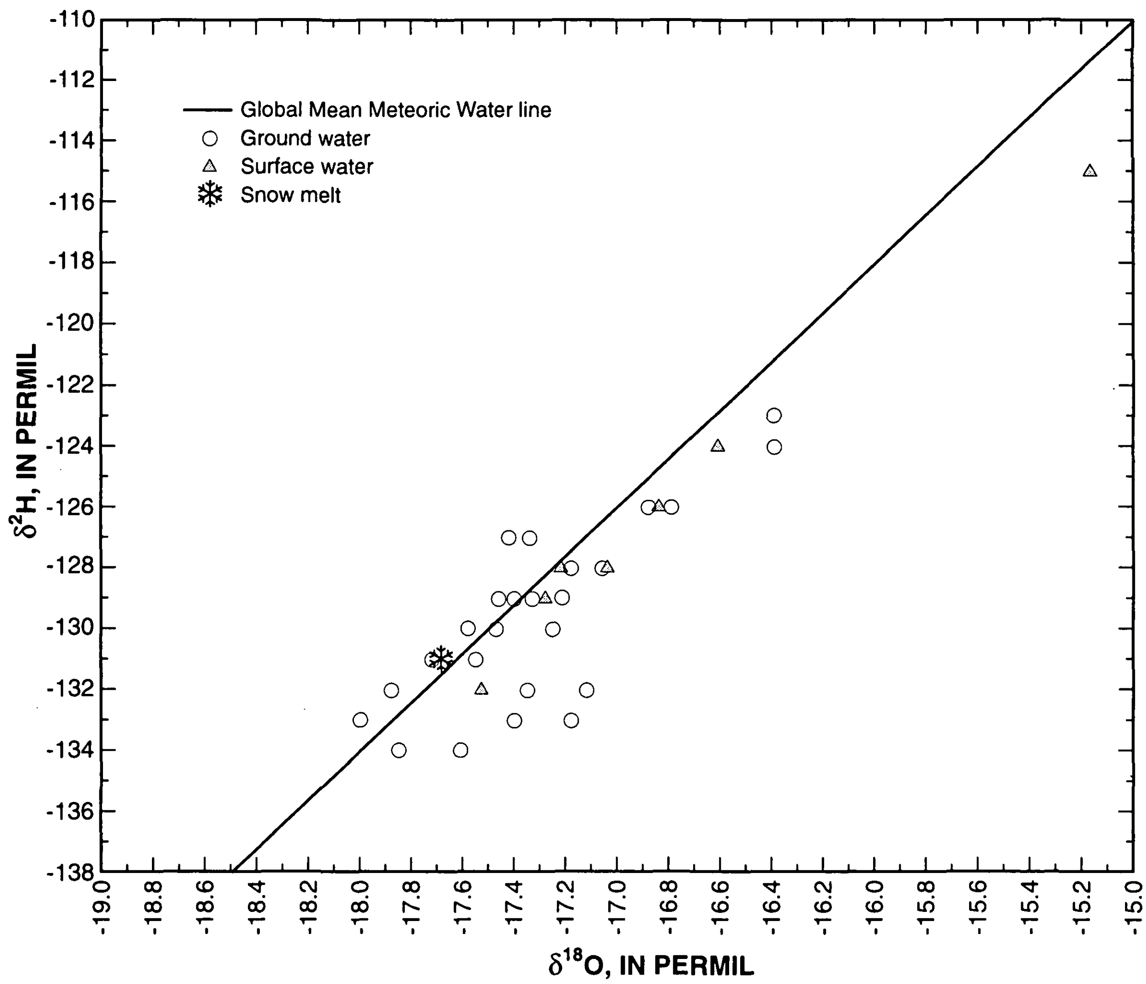


Figure 10. The relation between δ^2H and $\delta^{18}O$ values in water from Snyderville Basin, Park City, and adjacent areas, Utah, 1995.

Tritium concentration was determined in 12 samples collected from the study area (Downhour and Brooks, 1996, table 4). Tritium is a radioactive or unstable isotope of hydrogen that decays and has a half-life of about 12.3 years. Tritium occurs naturally in the atmosphere, but the largest source has been atmospheric nuclear testing from 1952 to 1969. The natural level for ^3H prior to atmospheric nuclear testing ranged from 2 to 8 TU. During large-scale atmospheric nuclear testing during 1962-63, ^3H levels were reportedly more than three orders of magnitude larger than natural concentrations (Plummer and others, 1993, p. 258). At the present time (1998), as a result of radioactive decay and the cessation of most atmospheric nuclear testing, atmospheric ^3H values are again approaching naturally occurring levels. Tritium concentrations in precipitation generally increase with increasing distance from the ocean and increasing latitude (Plummer and others, 1993, p. 258). Generally, the lowest ^3H values occur during the winter or early spring. Because of the lack of spatially distributed wells and uncertainties in determining ground-water flow paths within the study area, ^3H values in ground water were used only to estimate a relative time at which water entered the subsurface as recharge from precipitation.

Tritium was analyzed in water from 10 wells, the Spiro Tunnel, and a composite snow sample. The locations of samples collected for tritium and chlorofluorocarbon analyses are shown in figure 11. Tritium values range from less than detection limit to 58 pCi/L (Downhour and Brooks, 1996, table 4). This range is equivalent to 0 to 18 TU. These ^3H values indicate that water infiltrated into the ground-water system before and after atmospheric nuclear testing, which reached its peak during 1962-63. The ^3H value for the composite snow sample is 6.8 TU, which is representative of present recharge water.

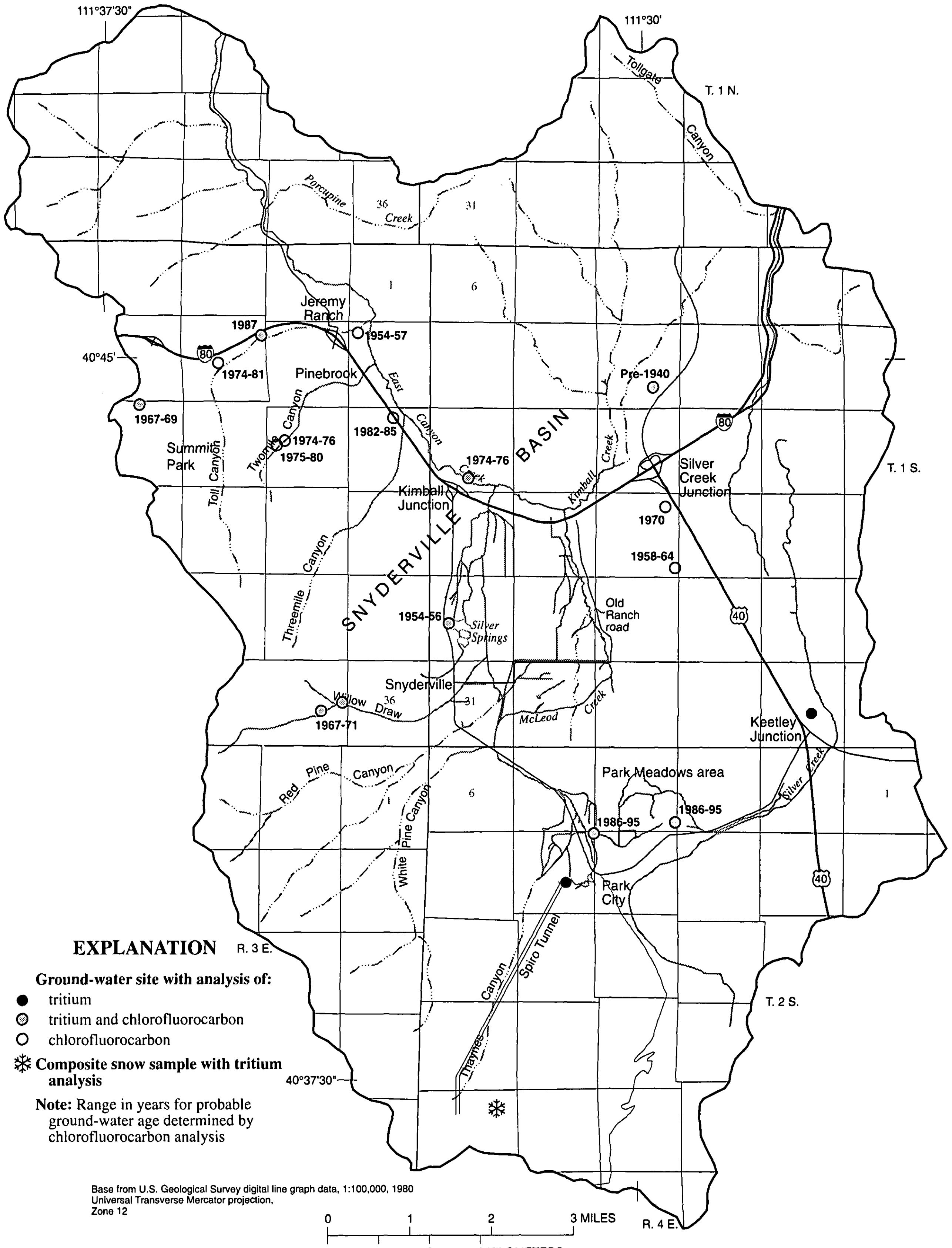


Figure 11. Age of ground water and location of selected ground-water sites with tritium and chlorofluorocarbon analyses, Snyderville Basin, Park City, and adjacent areas, Utah.

Water from wells (D-1-4)9dbd-1, (D-1-4)30cba-1, and (D-1-4)35dbb-1 had ^3H values of virtually 0 TU. These data indicate that water in these wells infiltrated into the subsurface prior to 1953, has not undergone mixing with younger water containing higher ^3H concentrations, and has a low flow velocity or long flow path. Well (D-1-4)9dbd-1 is located in the area north of Silver Creek Junction and is probably completed in the Tertiary conglomerate or volcanic rocks beneath shallow valley fill. Well (D-1-4)30cba-1 is located north of Snyderville and is completed in the Nugget Sandstone. Well (D-1-4)35dbb-1 is located just north of Keetley Junction and is probably completed in the volcanic rocks beneath unconsolidated valley fill. The low yield of this well indicates a relatively low permeability.

Water from well (D-1-3)35daa-1 has a ^3H value of 3.4 TU and water from well (D-1-3)35dba-1 has a ^3H value of 5.3 TU. Both wells are in the Willow Draw drainage, west of Snyderville. Well (D-1-3)35daa-1 is completed in the Nugget Sandstone and well (D-1-3)35dba-1 is completed in the Ankareh Formation. These ^3H values indicate mixing of water that infiltrated into the subsurface prior to large-scale atmospheric nuclear testing with relatively recent recharge water.

Tritium values in water from wells (D-1-3)10aad-2, (D-1-3)16baa-1, (D-1-4)18cda-1, (D-2-4)4dda-1, and (D-2-4)8aaa-1, and from the Spiro Tunnel range from 11.2 to 18.1 TU. These values indicate that recharge of these waters occurred either in the early 1970s or relatively recently, and that little or no mixing has occurred with water that was recharged prior to atmospheric nuclear testing.

Chlorofluorocarbons

Chlorofluorocarbons (CFCs) are stable volatile organic compounds that can be used for age-dating ground water that has infiltrated into the subsurface since the 1940s (Plummer and others, 1993, p. 268). All CFCs produced are eventually released to the atmosphere, where they are partitioned into water by gas-liquid exchange equilibria. The CFC concentration of ground water, therefore, is determined at the time when recharge water enters the saturated aquifer material and is dependent upon the atmospheric concentration. The concentration of CFCs in the atmosphere has been measured since the mid-1970s. Atmospheric concentration of CFCs has been estimated from 1940 to the mid-1970s by the use of production records (Plummer and others, 1993, p. 269). The concentration of CFCs in ground water can be used to determine a range of years for the date precipitation recharged the ground-water system. Three CFCs were analyzed during this study (Downhour and Brooks, 1996, table 4). These were trichlorofluoromethane (CCl_3F or F-11), dichlorodifluoromethane (CCl_2F_2 or F-12), and trichlorotrifluoromethane ($C_2Cl_3F_3$ or F-113).

The ground-water age determined for a certain concentration of CFCs depends on the recharge temperature, which is the temperature of the water at the base of the unsaturated zone. In deep unsaturated zones, which occur in the consolidated rocks throughout most of the study area, the recharge temperature probably corresponds to the mean annual temperature (Plummer and others, 1993, p. 271). The recharge temperature, however, can be less than the mean annual temperature in late winter and early spring when most recharge occurs. Age determinations can be incorrect if CFCs in the recharging water have been affected by CFCs in the local air, where concentrations higher than mean atmospheric concentrations can result from anthropogenic sources such as plastic containers, air conditioners, and aerosol cans, or if excess air is incorporated in the recharge water. The uncertainty in ground-water age increases with large perforated or open intervals in sampled wells because of increased mixing of water with possibly different ages. Also, more representative samples are obtained by using low pumping rates to minimize drawdown around the casing during sampling. These and other limitations of age dating using CFCs are discussed in Plummer and others (1993).

Samples were collected from 15 wells and 1 spring (fig. 11) (Downhour and Brooks, 1996, table 4). Recharge dates determined for the study area range from pre-1940 to 1995 (Eurybiades Busenberg, U.S. Geological Survey, written commun., 1995). A date of pre-1940 indicates that no CFCs were detected. Chlorofluorocarbon concentrations indicate that even though recharge raises ground-water levels throughout the study area within a few weeks of snowmelt, water typically takes 15 to 40 years to move through the ground-water system.

Samples were collected from wells (D-2-4)4dda-1 and (D-2-4)8aaa-1, both completed in the Thaynes Formation in the Park Meadows area. Values for CFCs indicate that recharge occurred during 1986-95, which is indicative of mixing from sources with different ages. The most recent age indicates that downward leakage occurs from unconsolidated valley fill to the underlying Thaynes Formation. This conclusion is supported by the presence of sodium and chloride in water from these wells, which probably is derived from the dissolution of road salt. Although chemical concentrations are higher in water in the overlying unconsolidated valley fill, they are diluted readily by mixing with water in the Thaynes Formation.

Samples were collected from two wells and one spring in the Thaynes Formation in the Pinebrook area. Values for CFCs in water from well (D-1-3)13abb-1, located near Interstate Highway 80, indicate that recharge occurred during 1982-85. Values for CFCs in water from well (D-1-3)14bcc-1 indicate that recharge occurred during 1974-76. Values for CFCs in water from spring (D-1-3)14bcd-S1, located close to well (D-1-3)14bcc-1, indicate that recharge occurred during 1975-80. Well (D-1-3)14bcc-1 and spring (D-1-3)14bcd-S1 are at higher altitudes than well (D-1-3)13abb-1 and therefore are considered to be upgradient of well (D-1-3)13abb-1 under normal ground-water flow paths from high-altitude recharge areas to lower-altitude discharge areas. The younger age for water from well (D-1-3)13abb-1 indicates probable mixing with younger water. This water can be derived from a closer source of recharge such as overlying unconsolidated valley fill or through faster flow paths such as fault zones and fractures from the distant source of recharge.

Samples were collected from three wells completed in the Twin Creek Limestone in the Summit Park area. Values for CFCs in water from well (D-1-3)10aad-2 indicate that recharge occurred during 1987. The CFC and other water-quality analyses indicate that water from this well might be withdrawn from the Twin Creek Limestone and unconsolidated valley fill. Values for CFCs in water from well (D-1-3)10caa-1 indicate that recharge occurred during 1974-81. Values for CFCs in water from well (D-1-3)16baa-1, located at a higher altitude than well (D-1-3)10caa-1, indicate that recharge occurred during 1967-69. Both wells are completed in the same block of Twin Creek Limestone, which has no major faults. Despite no obvious barriers to flow, water from the lower-altitude well in this block of Twin Creek Limestone has either a closer source of recharge or a faster flow path than water from a well at higher altitude that is presumably closer to the major recharge area.

Values for CFCs in water from well (D-1-3)12bbc-1, completed in the Nugget Sandstone near Jeremy Ranch, indicate that recharge occurred during 1954-57. The older date indicates that the water has taken a longer flow path, that mixing has not occurred with water from more recent sources, or that the Nugget Sandstone is not intensely fractured and has lower hydraulic-conductivity values and longer travel times than the nearby Twin Creek Limestone and Thaynes Formation.

Samples collected from three wells west of Snyderville show a conventional trend in ground-water age with the oldest water farthest from areas of recharge. Values for CFCs in water from well (D-1-3)35daa-1, completed in the Nugget Sandstone, and in water from well (D-1-3)35dba-1, completed in the Ankareh Formation, indicate that recharge occurred during 1967-71. The period for recharge assigned to water from both wells on the basis of values for CFCs contradicts the interpretation made from ^{3}H data. If the recharge water entered the subsurface during 1967-71 with little or no subsequent mixing of younger water, then ^{3}H values would have to be greater than those measured. Values for CFCs in water from well (D-1-4)30cba-1, completed in the Nugget Sandstone downgradient from these wells, indicate that recharge occurred during 1954-56. Fractures are not as prevalent in the Nugget Sandstone in this area as in the Summit Park and Pinebrook areas; therefore, most of the ground-water flow probably occurs in pore spaces.

Values for CFCs in water from well (D-1-4)21ddd-1, which is completed in the Twin Creek Limestone south of Silver Creek Junction, indicate that recharge occurred during 1958-64. The range in age indicates possible mixing from different sources. Water of this age is indicative of no local recharge and long flow times from the recharge area. This water is much older and has a higher dissolved-solids concentration than other water derived from the Twin Creek Limestone within the study area, which indicates longer ground-water travel times and probable longer flow paths from the area of recharge.

Values for CFCs in water from well (D-1-4)21aac-1, completed in the volcanic rocks near Silver Creek Junction, indicate that recharge occurred during 1970. The only sample with no detectable CFCs was collected from well (D-1-4)9dbd-1, completed in the volcanic rocks north of Silver Creek Junction. These data indicate that little local recharge and low hydraulic conductivity in this area result in long travel times from recharge areas.

Cluster Analysis

Hierarchal cluster analysis of simple salt assemblages for 33 samples collected in 1994 and 1995 identified four distinct hydrochemical facies. The location and facies of the samples are shown in figure 12. The computer program SNORM (Bodine and Jones, 1986) calculates normative salt and simple salt assemblages from the chemical composition of a natural water. A salt norm is an ideal equilibrium mineral assemblage that would crystallize if a sample of water is evaporated to dryness at 25 °C and 1 bar pressure and atmospheric pressure of CO_2. A simple salt assemblage is recast from the normative salt assemblage to a simplified representation that is composed of major solutes.

The normative salt assemblages in Facies 1 are generally composed of anhydrite, dolomite, calcite, halite, flauberite, and syngenite. This salt assemblage is characteristic of carbonic acid hydrolysis, and all samples in this facies were collected from wells, springs, mine tunnels, and surface water that are completed in or flow from the Twin Creek Limestone, Nugget Sandstone, Thaynes Formation, Weber Quartzite, or volcanic rocks.

The salt assemblages in Facies 2 are extremely distinctive and characterized by antarcticite, halite, tachyhydrite, dolomite, anhydrite, and carnallite, and could represent highly altered fluid compositions related to near-surface conditions or diagenetic alteration of a residual marine fluid. Most of the sample sites are from shallow wells, springs, or surface water. The possibility of a residual marine fluid does not seem likely for this facies because tritium values measured for these samples are 0.1, 11, 17, and 18 TU, indicating modern water for most of the sites. The geographic distribution of this facies (fig. 12) indicate that road salt could be causing the distinction of this facies. The samples were collected from the Park Meadows area, the Kimball Junction area, and along Interstate Highway 80.

The salt assemblages of Facies 3 are distinctly different from those of Facies 2, and are composed of anhydrite, dolomite, bischofite, halite, magnesite, and carnallite. Samples in Facies 3 were collected from the Park Meadows area, upper McLeod Creek, and the southern part of Snyderville Basin. This facies probably represents a mixing of different ground-water sources to this area, including the Thaynes Formation, the Spiro Tunnel, and the Woodside Shale adjacent to the Thaynes Formation.

The salt assemblages of Facies 4 are characterized by anhydrite, magnesite, polyhalite, bloedite, halite, blauberite, and epsomite, and are most characteristic of sulfuric weathering regimes, or contact with shales reducing enough to contain sulfides. Two samples in this facies are from the Spiro Tunnel, one is from well (D-1-3)35dba-1 completed in the Ankareh Formation, and one is from well (D-1-4)21ddd-1 completed in the Twin Creek Limestone.

Because the chemical composition of many of the consolidated rocks in the study area is similar, cluster analysis could not be used to determine the exact consolidated rocks through which water had flowed. The analysis, however, indicates that road salt is affecting the composition of water in the Park Meadows area and near Interstate Highway 80. Other contaminants also may affect water in these areas. The cluster analysis also indicated that mixing of different types of water appears to be occurring in the Park Meadows area and upper McLeod Creek.

Geochemical Modeling

Major-ion chemical data and isotopic data were used to help delineate ground-water flow paths and to compare these flow paths to measured water levels and known structural geology. Two computer models were used to aid the analyses. WATEQF (Plummer and others, 1984) was used to model the thermodynamic speciation of inorganic ions and complex species. Output from this model includes the concentration and activity of each aqueous species, and the activity product and saturation index of 101 minerals. NETPATH (Plummer and others, 1994) was used to interpret net geochemical mass-balance reactions along a hypothetical ground-water flow path and to determine if mixing of different sources could account for the chemical composition of water from some wells and springs.

Flow-path geochemical modeling requires that the chemical composition of aquifer material be known. The composition of most of the sedimentary rock in the study area has not been determined and had to be estimated on the basis of general rock description and general rock mineralogy as described in the "Hydrogeology" section of this report and Bryant (1990, sh. 1). The order of the chemicals listed below does not imply a proportion of the rock. The Keetley Volcanics primarily consist of plagioclase, hornblende, biotite, and pyroxene. The Twin Creek Limestone consists of calcite ($CaCO_3$), silica (SiO_2), gypsum ($CaSO_4$), montmorillonite with various cations, and

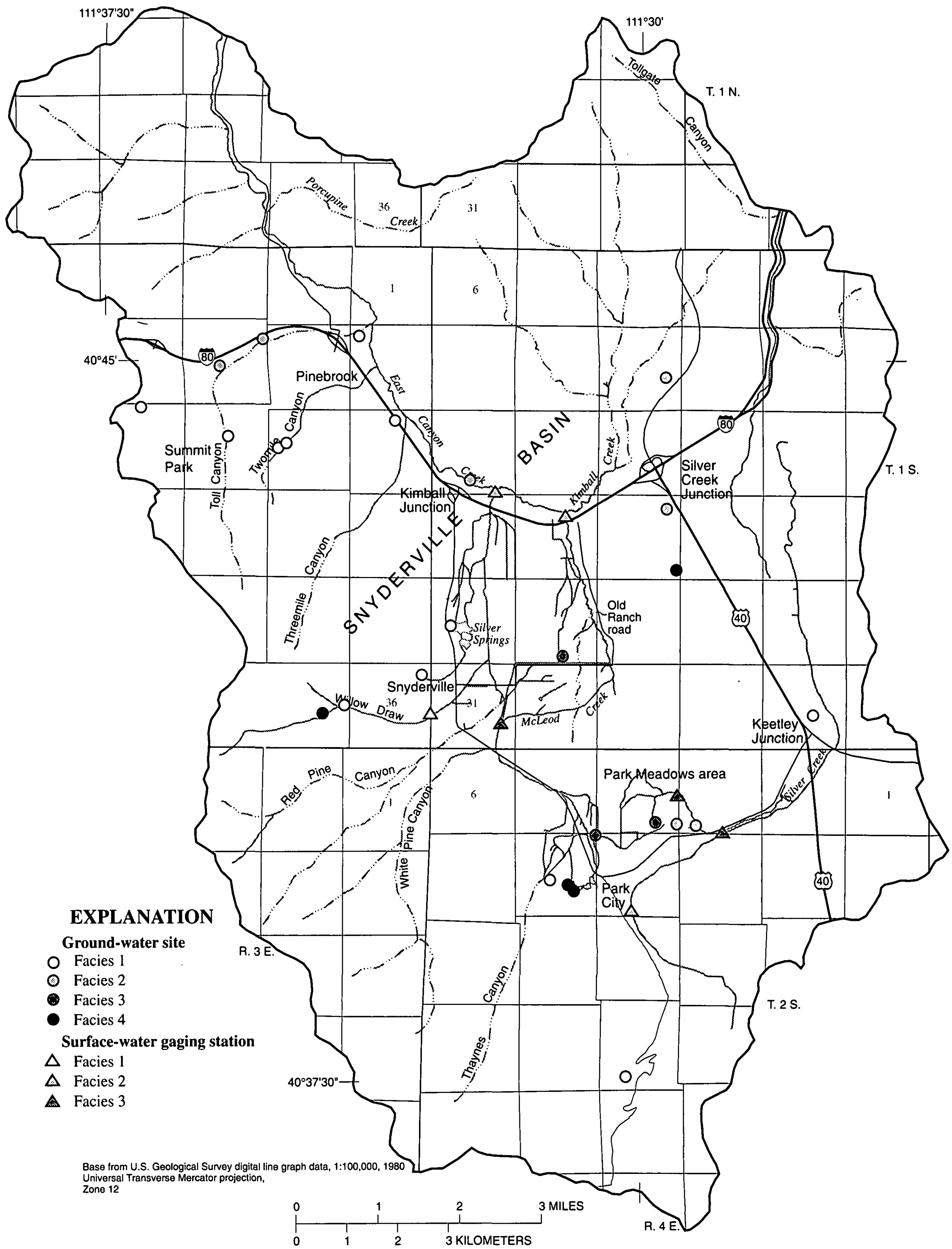


Figure 12. Location and hydrochemical facies of selected water samples in Snyderville Basin, Park City, and adjacent areas, Utah.

illite. The Nugget Sandstone consists of silica (SiO_2) and montmorillonite with various cations. The Ankareh Formation consists of calcite ($CaCO_3$), silica (SiO_2), gypsum ($CaSO_4$), montmorillonite with various cations, and illite. The Thaynes Formation consists of calcite ($CaCO_3$), silica (SiO_2), gypsum ($CaSO_4$), and montmorillonite with various cations. The Woodside Shale consists of silica (SiO_2), gypsum ($CaSO_4$), montmorillonite with various cations, and illite. The Park City Formation consists of calcite ($CaCO_3$), dolomite ($CaMg(CO_3)_2$), silica (SiO_2), and montmorillonite with various cations. The Weber Quartzite consists of calcite ($CaCO_3$), silica (SiO_2), and montmorillonite with various cations. The unconsolidated valley fill was derived from the consolidated rocks and consists of a mixture of the minerals found in the rocks.

Geochemical modeling was used to determine the amount of mixing of different sources of water being withdrawn from wells completed in the Thaynes Formation in the Park Meadows area. Water from well (D-2-4)8aaa-1 was determined to be a mixture of water from the Thaynes Formation upgradient from the well (as represented by spring (D-2-4)8cab-S1), water from the Spiro Tunnel, and water from the Woodside Shale (as represented by well (D-2-4)9aac-1). Because of uncertainties in the chemical composition of both the solid and liquid phases, an exact mixing ratio cannot be determined. Water from well (D-2-4)8aaa-1 consists of about 29 to 58 percent water from the Thaynes Formation upgradient from the well, 16 to 44 percent water from the Spiro Tunnel, and 26 percent water from the Woodside Shale. Much of the water from the Spiro Tunnel seeps into the unconsolidated valley fill and subsequently migrates into the Park Meadows area. The well withdraws water either from the unconsolidated valley fill or the Thaynes Formation. If the water is withdrawn from the unconsolidated valley fill, the well is not completed as indicated on the driller's log. If the water is withdrawn from the Thaynes Formation, the chemical analysis indicates that water from the Spiro Tunnel moves downward from the unconsolidated valley fill to the Thaynes Formation. Ground water from the Woodside Shale may move directly into the Thaynes Formation as dictated by a natural hydraulic gradient or may move upward into overlying unconsolidated valley fill and then move over the Thaynes Formation where it can move downward as a result of stress imposed by ground-water withdrawals from the Thaynes Formation.

Geochemical modeling near Snyderville indicates that water could move from the area of well (D-1-3)35daa-1, completed in the Nugget Sandstone, to the area of well (D-1-4)30cba-1, also completed in the Nugget Sandstone. This movement is consistent with the structural geology in the area, the hydraulic gradient between the two wells, and the fact that the older water is in the downgradient well. The lack of other wells suitable for water-level measurements in this area makes it difficult to accurately define the direction of ground-water flow. It is possible, therefore, that water does not flow directly between the two wells, but rather that water passes through similar rocks along the flow path to both wells. Water in well (D-1-4)30cba-1 probably is derived mainly by recharge with composite snowmelt water with additional dissolution of dolomite, calcite, sodium chloride, gypsum, and possibly sodium montmorillonite. Structural geology and water-level data indicate that water near well (D-1-3)35dba-1 completed in the Ankareh Shale should also flow toward (D-1-4)30cba-1. Geochemical modeling, however, indicates the percentage of this water in well (D-1-4)30cba-1 is only about 1 percent and that little mixing of water in the Ankareh Shale with water in the Nugget Sandstone occurs in this area. Well (D-1-3)35dba-1 is located near the top of the Willow Creek anticline, and bedding of the sedimentary rocks indicates that water could move northeast along the axis of the anticline toward Kimball Junction, or could move southeast along bedding planes, then northeast along the Dutch Draw syncline.

Because of the complex structural geology and the lack of sufficient water-level data to help delineate ground-water flow paths, geochemical modeling was not done for the Pinebrook and Summit Park areas. More supporting data would reduce the number of possible solutions to the geochemical modeling and permit more certain interpretations.

WATER-BUDGET ANALYSIS

Water budgets were determined for the total-water system, the surface-water system, and the ground-water system for six subbasins within the study area. The six subbasins are McLeod, Snyderville, Silver Creek Junction, East Canyon, Upper Silver Creek, and Lower Silver Creek (pl. 1). The subbasin boundaries are topographical divides and delineate areas of ground-water development, but are not necessarily ground-water divides. Ground water may flow between subbasins, but the amount of flow could not be determined with the data available. Many monitoring wells and production wells would be needed to determine the hydraulic gradient and aquifer characteristics near the boundaries of subbasins, and these wells do not exist. Water budgets were determined monthly from October 1994 to September 1995 and are presented as annual summaries for the 1995 water year. Monthly accounting was needed because of the extreme seasonal water-level and streamflow variations. The water budgets for the six subbasins were combined to produce the water budgets for the entire study area. The budget area, as defined for this report, is each subbasin or the entire study area.

The total-water budget, as defined for this report, includes all water that enters or leaves each budget area. Inflows include precipitation, surface-water inflow, ground-water inflow, and imported municipal wastewater. Outflows include consumptive use, surface-water outflow, ground-water outflow, and exported municipal wastewater. The individual inflows and outflows for each subbasin and the study area are listed in table 3. The distribution of precipitation between surface water and ground water and interactions between surface water and ground water are not included in the budget for the total-water system. Because much of the precipitation enters neither the surface-water system nor the ground-water system, and because the surface- and ground-water systems interact, the budgets for the surface-water system and the ground-water system do not sum to the budget for the total-water system.

The surface-water budget includes all surface water that enters or leaves each budget area. Inflows to the surface-water system include surface-water inflow, surface-water flow contributed by precipitation, surface-water contributed by the discharge of ground water, and return flow from irrigation with municipal water (pl. 1). In the East Canyon and Lower Silver Creek subbasins, surface-water inflow also includes flow from wastewater-treatment plants. Outflows from the surface-water system include surface-water outflow, surface-water that recharges the ground-water system, and consumptive use of surface water. Surface-water inflow, surface-water outflow, and consumptive use of surface water in the surface-water budget are the same as in the total-water budget. The residual is typically small in the surface-water budget, which indicates that errors in the estimates for individual budget components are small and that most inflow and outflow has been determined.

The ground-water budget includes all ground water that enters or leaves each budget area. Inflows to the ground-water system include recharge from precipitation, recharge from infiltration of streams, and recharge from irrigation (pl. 1). Inflow to the McLeod subbasin includes ground-water inflow across the south boundary of the study area. Inflow to the Upper Silver Creek subbasin includes ground-water inflow across the south boundary of the study area and from the McLeod subbasin. Outflows from the ground-water system include discharge to streams, mine tunnels, springs, and wells, and consumptive use of ground water. Outflow from the McLeod subbasin includes ground-water flow to the Upper Silver Creek subbasin. Ground-water inflow, ground-water outflow, and consumptive use of ground water in the ground-water budget are the same as in the total-water budget. The residual in the ground-water budget is typically larger than the residual in the surface-water budget and similar to the residual in the total-water budget (pl. 1 and table 3), which indicates than ground-water budget components have larger errors than surface-water budget components or that some sources of recharge or forms of discharge were not determined. Because the residual is positive, the residual could include undetermined amounts of ground-water outflow or an increase in ground-water storage.

A water budget represents the conservation of water. The difference between the inflows and the outflows during a given time equals the change in storage during the same time. Associated with each component of inflow and outflow are uncertainties caused by errors of measurement and interpretation of data (Winter, 1981, p. 82). The values of budget components in this report are rounded to indicate the degree of uncertainty associated with the component. Some small components are included to identify processes, but the amount of these components is negligible compared to the possible errors in larger components. The residual of the water budget is the net error of all the budget terms and includes inflows, outflows, and changes in storage that have not been determined. In the water budgets presented in this report, ground-water flow from the study area and the change in ground-water storage, including change in soil moisture, are not determined. In only a few areas are ground-water inflow and ground-water flow between subbasins estimated. The change in storage and ground-water inflow and outflow, therefore, are included in the residual of the total-water budget and the ground-water budget for each subbasin and the study area. To determine the change in storage and ground-water inflow and outflow would require monitoring wells and production wells that do not exist or

Table 3. Area and estimated total-water budgets for Snyderville Basin, Park City, and adjacent areas, Utah, 1995

[Inflows and outflows in acre-feet; —, not applicable; ?, could not be determined]

Budget Element	Entire study area	McLeod subbasin	Snyderville subbasin	Silver Creek Junction subbasin	East Canyon subbasin	Upper Silver Creek subbasin	Lower Silver Creek subbasin
Area (acres)	65,000	9,300	10,700	7,700	17,100	6,500	13,700
			Inflow				
Precipitation	204,000	37,000	36,000	19,000	60,000	21,000	30,000
Surface-water inflow	0	0	20,000	0	31,000	2,000	6,000
Municipal wastewater imported from other subbasins	—	—	—	—	1,300	—	1,400
Ground-water inflow from outside the study area	7,000	[1]4,000	?	?	?	[1]3,000	?
Ground-water inflow from other subbasins	—	?	?	?	?	1,000	?
Total inflow (rounded)	**211,000**	**41,000**	**56,000**	**19,000**	**92,000**	**27,000**	**37,000**
			Outflow				
Plant precipitation use and evaporation	108,000	14,000	18,000	14,000	29,000	9,000	24,000
Sublimation	2,000	600	300	0	200	400	40
Surface-water outflow	54,000	18,000	31,000	4,000	42,000	6,000	12,000
Mine-tunnel flow to the Provo River drainage	9,000	—	—	—	—	9,000	—
Consumptive use of ground water by crop and riparian areas	2,000	100	1,000	300	200	0	600
Consumptive use of surface water by crop and riparian areas	1,000	400	400	0	100	200	300
Consumptive use of municipal water by lawns and gardens	600	50	200	0	100	200	0
Ground-water outflow to other subbasins	—	1,000	?	?	?	?	?
Consumptive use from artificial snow	50	40	0	—	—	10	—
Municipal export to wastewater-treatment plant	—	1,200	500	0	0	1,000	0
Total outflow (rounded)	**177,000**	**35,000**	**51,000**	**18,000**	**72,000**	**26,000**	**37,000**
Residual[2]	34,000	6,000	5,000	1,000	20,000	1,000	0

[1] This amount of inflow is needed to produce a balanced ground-water budget in specific consolidated-rock units in this subbasin, and includes errors and residuals in other components of the ground-water budget.

[2] The residual includes the net error of inflows and outflows, change in soil moisture and ground-water storage during the 1995 water year, ground-water flow between subbasins, and ground-water outflow from the study area. A positive residual could result from an increase in soil moisture or ground-water storage, an overestimate of ground-water recharge or ground-water inflow, or an underestimate of ground-water discharge or ground-water outflow. A negative residual could result from a decrease in soil moisture or ground-water storage, an underestimate of ground-water recharge or ground-water inflow, or an overestimate of ground-water discharge or ground-water outflow.

are not available for monitoring or for pumping during aquifer tests.

The water-budget analysis was used to better understand the hydrologic system in each subbasin. The sections of this report discussing each subbasin and the study area do not necessarily repeat the inflows and outflows for each budget (table 3 and pl. 1), but discuss the hydrologic system and the processes that are most important for each area.

Methods

The methods of analysis used to estimate many water-budget components were the same for each subbasin and are discussed in the following section. Methods of analysis specific to a single subbasin are discussed in the appropriate subbasin section of this report. Methods used to determine surface-water flow are discussed in the "Streamflow" section of this report.

Precipitation and Sublimation

Most water enters the study area as precipitation, which is mostly snowfall. Precipitation is sublimated, is used by plants, becomes direct runoff to streams, and recharges the ground-water system. Precipitation is measured at three locations in the study area (pl. 1). Monthly precipitation was determined using the 1961-90 winter and summer normal precipitation contours (Utah Climate Center, 1996) to distribute the monthly precipitation at the Thaynes Canyon SNOTEL site across the study area from October 1993 to February 1994, from July 1994 to February 1995, and from July 1995 to September 1995. The amount of precipitation that fell as snow and remained snow for the month was estimated using the precipitation and snow-water-equivalent data at the Thaynes Canyon SNOTEL site and the Parleys Summit SNOTEL site and the snow-on-ground data at the Snyderville National Weather Service climate station for October 1994–February 1995, and was adjusted using the initial conditions described in the "Energy-balance snowmelt simulation" section of this report. From March to June 1994 and March to June 1995, the distribution and amount of monthly precipitation and snowmelt was determined using the simulation described. The results of the simulation were extrapolated to areas in the northern part of the study area that were not simulated.

Errors associated with estimating precipitation are large compared to errors in other components of the total-water budget. The errors include errors in measurement of precipitation at the SNOTEL sites and climate station and errors in interpreting the distribution of precipitation over the study area. The distribution varies with season, storm pattern, altitude, and other factors. Because the amount of precipitation is the largest component in the total-water budget (table 3), errors in estimating precipitation could be significant.

Sublimation directly from the snow from March to June 1995 was determined by simulation. Evaporation and sublimation also occur during snowmaking and spring melt of artificial snow. The evaporation during snowmaking in Colorado has been measured as 6 percent of the water used (Wright Water Engineers, Inc., 1986, p. 6). The combined loss during snowmaking and early-season sublimation in New Mexico was measured as 4.5 percent of the water used (Smart and Fleming, 1985, p. 11). A combined loss during snowmaking and early-season sublimation of 5 percent of the water used is assumed for the study area. Sublimation from artificial snow during spring melt was assumed to be the same percentage as sublimation from the natural snow. Snowmaking occurs in the McLeod, Snyderville, and Upper Silver Creek subbasins. In comparison to the estimate errors in other budget components, consumptive use of water by snowmaking is negligible. Part of the water used for snowmaking becomes ground-water recharge and part becomes direct runoff to surface water, but these amounts are negligible compared to estimate errors in recharge and direct runoff from precipitation and are not considered in this budget analysis.

Errors associated with estimating sublimation from snow are large compared to errors in other components in the total-water budget. The amount of sublimation varies with wind speed, vapor pressure, and other factors. Because sublimation, however, is a small part of the total-water budget (table 3), the errors are not significant.

Plant Use of Precipitation and Soil Moisture

The amount of precipitation that is intercepted, transpired, evaporated, or used as soil moisture is not available as surface water or ground water. Consumptive use of natural vegetation throughout the study area was estimated on the basis of other studies in similar climates (table 1). The consumptive use of dry meadow vegetation in the study area was assumed to be higher than reported by Tomlinson (1996a, table 5) because more summer precipitation was available. It was assumed that dry meadows used all available precipitation not to exceed the 0.25 ft/mo maximum use reported by Tomlinson (1996a, table 5). Consumptive use for crops and lawns, including nonirrigated pasture, was estimated using values determined at Park City by Utah State University (1994, table 25). Soil moisture was assumed to be negligible at the beginning of the water year. Fall, winter, and early spring precipitation was assumed to contribute to soil moisture until soil moisture was the amount needed to sustain each plant type through the drier summer months, but provide no extra for the next water year. Typically, soil moisture is probably negligible at the end of a water year because the natural vegetation has developed to use the amount of water normally available. Soil moisture at the end of the water year in areas irrigated with surface water probably fluctuates depending on how much surface-water irrigation was available during the summer.

Errors associated with estimating plant use of precipitation are large compared to errors in other components of the total-water budget. Errors include errors in measurement in the referenced research and errors in applying those measurements to this study area. Plant use varies with humidity, wind speed, slope aspect, and other factors. Because plant use of precipitation is the second largest component in the total-water budget (table 3), errors in estimating could be significant.

Direct Runoff of Precipitation

Direct runoff of rain and melting snow to streams was estimated through the use of hydrograph-separation techniques on streamflow at the six gaging stations operated by the U.S. Geological Survey in the study area during 1994 and 1995 (table 2). The flow was divided into direct runoff and ground-water contribution (base flow) for each month. Because ground-water levels fluctuate seasonally, base flow also fluctuates seasonally. Ground-water contribution to streamflow was also determined by seepage runs, and that information was used to help interpret the hydrograph-separation techniques.

Errors associated with estimating direct runoff to streams are small compared to errors in estimating precipitation and plant use of precipitation. The errors are associated with errors of measurement of streamflow at gaging stations and interpretation errors in dividing the flow between direct runoff and base flow and estimating streamflow from ungaged drainages. The interpretation errors are larger than the measurement errors. Because direct runoff of precipitation is a small part of precipitation, the errors are not significant in determining the uses of precipitation.

In addition to the determination of direct runoff to streams at each of the gaging stations, the percentage of precipitation that becomes direct runoff was estimated for specific consolidated-rock units. Direct runoff of precipitation from the Thaynes Formation was estimated to be a negligible percentage of precipitation because Thaynes Canyon is underlain by the Thaynes Formation (pl. 2) and has negligible surface runoff. Direct runoff of precipitation from the Twin Creek Limestone was estimated to be a negligible percent of precipitation because water withdrawn from a well during an aquifer test in the Summit Park area was allowed to flow over the ground surface and infiltrated rapidly into the Twin Creek Limestone (William Loughlin, Weston Engineering, oral commun., 1996). Direct runoff of precipitation from the Weber Quartzite was assumed to be 5 percent of precipitation because the Weber Quartzite underlies most of the upper reaches of Silver Creek (pl. 2) and the flow in Silver Creek near Park City is less than estimates based on drainage area. Also, flow in Silver Creek at the U.S. Geological Survey gage (table 2) peaks in March, not in May or June, which indicates that the snowmelt from the higher altitudes underlain by the Weber Quartzite does not contribute greatly to streamflow. Direct runoff of precipitation from the Nugget Sandstone was estimated to be 10 percent of precipitation because most of White Pine Canyon is underlain by Nugget Sandstone (pl. 2) and direct runoff at the gaging station in White Pine Canyon near Park City was about 10 percent of precipitation. Runoff from other units was calculated from measured runoff at gaging stations and ranged from 9 to 41 percent of precipitation.

Ground-Water Recharge From Precipitation

Precipitation that is not sublimated, is not used by vegetation, does not replenish soil moisture, and does not contribute direct runoff to streams becomes ground-water recharge. Thus, ground-water recharge was determined as the residual of precipitation minus the other uses of precipitation and includes the net error of precipitation, sublimation, plant use of precipitation, and direct runoff of precipitation.

The errors associated with estimating ground-water recharge from precipitation are large because of the large errors in estimating precipitation and plant use of precipitation. Because recharge from precipitation is the largest component of the ground-water budget, the errors could be significant.

Ground-Water and Stream Interactions

Recharge from infiltration of streams was determined by using data from gaging stations, weirs, seepage studies, and field observation. The amount of loss in a stream was calculated by the same method for streams with gaging stations or for streams with instantaneous measurements during seepage studies. If, after measuring or estimating all surface inflows and outflows between two sites on a stream, the downstream site had less measured flow than the upstream site, the difference was assumed to be infiltration of stream water into the ground-water system. For streams with gaging stations, monthly estimates of infiltration of streams were made and summed for an estimate for the 1995 water year. For streams with instantaneous measurements, the measured loss was extrapolated to the loss in the 1995 water year on the basis of data from gaged sites and the season of the measurement. For example, instantaneous measurements of McLeod Creek in October 1994 (Downhour and Brooks, 1996, table 7) were compared to gaged measurements in October 1994 and adjusted proportionally for other months to determine the total for the 1995 water year. For some small streams, observations indicated that all streamflow infiltrated to the unconsolidated valley fill near the mouths of canyons because either no flow was observed in the channel, or because no channel was observed in the unconsolidated valley fill.

Many gaging stations and weirs operated by water companies are in canyons near the contact between consolidated rock and unconsolidated valley fill. Streamflow probably infiltrates to consolidated rock upstream from the measurement sites. Because the streamflow is not measured upstream, however, the amount of infiltration cannot be determined. The amount of infiltration also is not included in the direct runoff calculated for the stream, and is therefore included in the residual that is estimated to be ground-water recharge from precipitation. In either case, the same amount of water is estimated to recharge the ground-water system in consolidated rocks, but the process of recharge from precipitation or recharge from infiltration of streams is not delineated in these areas.

Ground-water discharge directly to streams and to riparian areas that contribute to streamflow was determined by using data from gaging stations, weirs, and seepage studies. The amount of gain in a stream was calculated by the same method for streams with gaging stations or for streams with instantaneous measurements during seepage studies. If, after measuring or estimating all surface inflows and outflows between two sites on a stream, the downstream site had more measured flow than the upstream site, the difference was assumed to be ground-water discharge to the stream. For streams with gaging stations, monthly estimates of discharge were made and summed for an estimate for the 1995 water year. For streams with instantaneous measurements, the measured gain was extrapolated to the gain in the 1995 water year on the basis of data from gaged sites and the season of the measurement.

Errors associated with determining infiltration of streams to the ground-water system and ground-water discharge to streams are fairly large compared to errors in other components in the surface-water and ground-water budgets. The errors are associated with measurement errors of surface-water flow, estimate errors of unmeasured inflows and outflows, and the application of instantaneous measurements to the entire water year. The errors could be significant in both the surface-water budget and the ground-water budget. The errors could be reduced by additional surface-water gaging stations or additional instantaneous measurements made throughout the year. Repeating the same measurements on two or three consecutive days helps to understand if gains or losses are real or are part of the measurement errors. The locations of, processes of, and any changes in ground-water and stream interactions could best be understood by additional monitoring wells near streams at different depths in the unconsolidated valley fill and consolidated rocks. Accurate altitudes of streams and nearby monitoring wells would aid interpretation.

Consumptive Use, Return Flow, and Ground-Water Recharge from Irrigation

In addition to precipitation, plants also consume surface water along streams, ground water, and irrigation water. Water for consumptive use in riparian areas and nonirrigated pasture can be supplied by both surface water and ground water. If surface water was available, plants were assumed to use all surface water needed to meet consumptive-use demand that was not met by precipitation. In areas where surface water was not available, ground-water use by riparian areas and nonirrigated pasture was estimated to be 50 percent of the consumptive-use demand that was not met by precipitation. This assumes that ground-water levels during at least part of the growing season are below the optimum level for plant use. Irrigated crops and lawns were assumed to use precipitation and available irrigation water to meet the consumptive-use demand.

Recharge from irrigation occurs when precipitation plus applied water reaching the root zone exceeds consumptive-use demand of the plants. Residential land, golf courses, and irrigated crop areas were determined from a digital landuse map (Utah Department of Natural Resources, Division of Water Resources, 1992). Residential areas and irrigated crops were assumed to use available precipitation and soil moisture to satisfy consumptive use demands before using applied irrigation water.

Thirty percent of residential areas was assumed to be lawns or gardens. Water applied to lawns, gardens, parks, school grounds, and other public facilities was determined by calculating the volume of extra water supplied by municipalities during the summer months. About 2,000 acre-ft (about 30 percent of municipal supply) were applied to lawns, gardens, and public facilities during the 1995 water year. Fifty percent of the extra water was assumed to reach the root zone of plants and either became consumptive use or recharged the ground-water system. Twelve percent of the water used for lawns, gardens, and public facilities was assumed to evaporate from impermeable surfaces. The remaining 38 percent of the water used was assumed to enter stream channels, either through direct runoff or through storm drains.

Golf courses in the study area are irrigated with surface water from nearby streams. The amount of water applied to golf courses was obtained from golf-course supervisors. Golf courses have well-designed sprinkler systems, and 80 percent of the water applied to golf courses was assumed to reach the root zone (Utah State University, 1994, table 19). The remaining 20 percent was assumed to flow back to the same stream from which it was diverted.

The amount of surface-water irrigation applied to alfalfa and pasture, the only crops grown in the study area, was estimated from records of the Utah Department of Natural Resources, Division of Water Rights (written commun., 1996); from diversion records along the streams (Weber River Water Commissioner, 1995 and 1996); from other records of surface water (John Bollwinkel, Community Water Company, written commun., 1995 and 1996, and Rich Hilbert, Park City Water Department, written commun., 1995 and 1996); and from stream measurements by the U.S. Geological Survey (Downhour and Brooks, 1996, tables 6, 7, and 9). If diversion records to fields were not available, 1.23 ft of water was assumed to reach the root zone in addition to effective precipitation (Utah Department of Natural Resources, Division of Water Resources, 1996, table 8). Fifteen percent (Utah Department of Natural Resources, Division of Water Resources, 1996, p. 29) of the flow through canals was estimated to recharge the ground-water system and is included as recharge from irrigation. Fifty percent (Utah Department of Natural Resources, Division of Water Resources, 1996, p. 29) of surface-water irrigation applied to fields was assumed to flow back to the same stream from which it was diverted. The remaining 50 percent either became consumptive use by crops or recharged the ground-water system. In the surface-water budgeting process, return flow was never considered to be diverted from the stream and therefore is not included in the surface-water budget (pl. 1).

Errors associated with determining consumptive use, return flow, and recharge from irrigation are small compared to errors in other components in the surface-water and ground-water budgets. Errors are associated with the reported values of consumptive use, the application of these values throughout the study area, area estimates of land use, and estimates of the amount of water that is effectively applied. Because consumptive use, return flow, and ground-water recharge from irrigation are small components of the budgets, the errors probably are not significant.

Withdrawal from Wells

Withdrawal from wells was determined from records from the Utah Department of Natural Resources, Division of Water Rights (written commun., 1996) for public-supply wells and estimated for private domestic wells. Errors associated with withdrawal from wells are small and are associated with meter inaccuracies and estimates of private use. The errors are insignificant in the ground-water budget.

Municipal Wastewater

The amount of water discharged from municipal wastewater-treatment plants near East Canyon Creek and Silver Creek was obtained from the Snyderville Basin Sewer Improvement District (Rex Osborne, written commun., 1995 and 1996). The amount of water that becomes municipal wastewater in each subbasin was estimated to be 98 percent (Utah Department of Natural Resources, Division of Water Resources, 1996, table 16) of the average monthly winter use in each subbasin summed for the water year. Some adjustment to this estimate was required to obtain the same amount reported by the Snyderville Basin Sewer Improvement District.

In subbasins without wastewater-treatment plants, municipal wastewater is considered to be exported to other subbasins. In subbasins with wastewater-treatment plants, the municipal wastewater from upstream subbasins is considered to be imported. Municipal wastewater from subbasins with wastewater-treatment plants is not accounted for separately because it is included with measured surface-water outflow.

Errors associated with estimating the amount of municipal wastewater are small and are associated with meter inaccuracies and estimating the amount generated by each subbasin. The errors are not significant in the surface-water budget.

Residual

The total-water budget (table 3) and ground-water budget (pl. 1) for the study area and many subbasins indicate that more water entered than left the study area or subbasin during the 1995 water year. Assuming that all determined components are accurate, the residual indicates either an increase in ground-water storage, ground-water flow out of the study area, or a combination of both. Water levels were higher in most of the study area and in most formations in September 1995 than in September 1994 (Downhour and Brooks, 1996, table 3), which indicates that storage increased during the water year. Because storage-coefficient values could not be determined, the amount of increase in ground-water storage could not be determined. Monitoring wells do not exist and aquifer characteristics could not be determined to estimate the amount of flow that could be leaving the study area as ground-water outflow.

Part of the residual also could be caused by errors in estimating budget components in the total-water budget and ground-water budget. The amount of precipitation and the amount of precipitation used by plants are the components with the largest potential errors. Because ground-water recharge from precipitation is calculated with those components, the ground-water budget also is affected by estimate errors in precipitation and plant use. If precipitation was actually less than was estimated, the inflow to the total-water budget would be less and the residual in the total-water budget would be less. Ground-water recharge from precipitation also would be less, and the residual in the ground-water budget would be less. Precipitation and recharge, however, could be greater than estimated, which would result in a larger residual in the total-water and ground-water budgets. If plant use was less than estimated, then outflow from the total-water budget would be less and the residual would be greater. Recharge from precipitation would be greater because less precipitation was used by plants, and the residual in the ground-water budget also would be greater. If plant use was greater than estimated, the residuals in the total-water budget and ground-water budget would be less.

McLeod Subbasin

The McLeod subbasin contains 9,300 acres, makes up 14 percent of the study area, and received 18 percent of the precipitation in the study area (table 3) in the 1995 water year. Inflow to the subbasin was from precipitation and ground-water inflow from south of the study area. The subbasin has no surface-water inflow. Outflow from the subbasin consisted mainly of surface-water outflow to the Snyderville and Upper Silver Creek subbasins, water consumed in the subbasin, export of municipal wastewater to the East Canyon and Lower Silver Creek subbasins, and probable ground-water outflow to the Upper Silver Creek subbasin. The McLeod subbasin is the only subbasin where the amount of surface-water outflow generated in the subbasin was greater than the amount of water consumed in the subbasin. The residual of the total-water budget and the ground-water budget indicate that ground-water storage increased by about 5,000 acre-ft if all other budget components are accurate. The residual also could indicate that additional ground water is flowing to other subbasins or out of the study area. Water levels were higher in the fall of 1995 than in the fall of 1994 in two wells completed in the unconsolidated valley fill and in one well completed in the Twin Creek Limestone or Nugget Sandstone. The increases range from 10 to 25 ft.

Discharge from the Spiro Tunnel contributed about 30 percent of the surface water in the subbasin. Discharge from the Spiro Tunnel and spring (D-2-4)8dab-S1 (Theriot Springs) was obtained from the Park City Water Department (Rich Hilbert, written commun., 1995 and 1996). Some of the water from the

Spiro Tunnel and all of the water from Theriot Springs flows through the Park City Water Treatment Plant. The records of discharge from the tunnel and spring were combined with records of discharge from the water-treatment plant, also provided by the Park City Water Department, to determine the amount of spring and tunnel water that entered the water system and the amount that flowed through the plant to streams. Records provided by the Park City Water Department indicate that the Spiro Tunnel discharged about 8,000 acre-ft of water during the 1995 water year. Most of this discharge is from the Weber Quartzite. Measurements of flow in the tunnel (Rich Hilbert, written commun., 1995 and 1996) indicate that from June to December 1995, 94 percent of the discharge from the Spiro Tunnel originated at points farther than 6,600 feet into the tunnel and therefore discharged from the Park City Formation or the Weber Quartzite. Ashland and others (1996, table 7) report that when the tunnel was constructed, most of the flow was from the Weber Quartzite. The amount of water in the Weber Quartzite that is derived from precipitation in the subbasin and the amount that is derived from precipitation south of the subbasin was not determined during this study.

Discharge from spring (D-2-4)8dab-S1 (Theriot Springs) and spring (D-2-4)8cab-S1 (Sullivan Springs) contributed about 20 percent of the surface water in the subbasin. Discharge from spring (D-2-4)8cab-S1 (Sullivan Springs) was determined by monthly measurements by the U.S. Geological Survey (Downhour and Brooks, 1996, table 5). The discharge from other springs in the subbasin is not significant and was included as ground-water discharge to streams.

Ground-water discharge to streams contributed about 25 percent of the surface water in the subbasin. The Park City Water Department maintains and reads many flumes in the subbasin. The records for these flumes provided much of the data to determine ground-water and stream interactions and the amount of surface-water outflow to the Upper Silver Creek subbasin. Surface-water routing calculations indicate that streams near the mouth of Thaynes Canyon gain water during the spring and early summer and lose water during late summer, fall, and winter. Surface-water measurements during October 1994 (Downhour and Brooks, 1996, table 7) indicate that ground-water and stream interactions may change with a slight amount of recharge. Measurements on October 12, 1994, indicate that streams lost water. Rainstorms occurred between October 12 and October 25. Measurements on October 25 and 28, 1994, indicate that streams gained water. The loss and gain are greater than possible measurement errors of 20 percent. The streams are not in direct connection with consolidated rocks, but these gains and losses could indicate discharge from the Thaynes Formation and other consolidated rocks through the unconsolidated valley fill to the streams, or recharge from the streams to the unconsolidated valley fill and underlying consolidated rocks. Installation of water-level monitoring wells would help determine the seasonal gradient between consolidated rocks, unconsolidated valley fill, and streams.

Direct runoff of precipitation contributed only about 25 percent of the surface water in the subbasin, a smaller percentage than in any other subbasin. In June and July 1995, White Pine Canyon contributed about 75 percent of the direct runoff in the subbasin for those months, which indicates that much of the snowmelt at high altitudes within other parts of the subbasin was infiltrating to the ground-water system and not becoming direct runoff. Flow in McLeod Creek gradually increases from January through June as discharge from the Spiro Tunnel increases, but does not significantly increase during snowmelt in June, which also indicates infiltration of snowmelt at high altitudes. The lack of flow in Thaynes Canyon Creek indicates that substantial recharge is occurring from infiltration of precipitation in Thaynes Canyon.

Despite the large amount of infiltration of precipitation in the Thaynes Canyon area, however, ground-water discharge in the area exceeded estimated ground-water recharge from precipitation. Recharge from precipitation to the Thaynes Formation, the Weber Quartzite, and other less permeable consolidated rocks in 1995 was about 9,000 acre-ft. Information reported by Ashland and others (1996, pls. 8 and 10) was used to determine the area of the Thaynes Formation and Weber Quartzite that receives direct recharge from precipitation. This recharge was not sufficient to supply the estimated discharge from these units to the Spiro Tunnel, the two large springs, and ground-water flow to the Upper Silver Creek subbasin. Assuming no errors in budget components and no change in ground-water storage, about 4,000 acre-ft more recharge was needed to supply the discharge for the 1995 water year. This water is probably supplied by ground-water flow across the southern boundary of the subbasin and study area. Ground-water levels were higher and spring discharge was greater in September 1995 than in September 1994, which indicates an increase in ground-water storage. Assuming no errors in other budget components, the amount of ground-water flow across the southern boundary may have been greater than 4,000 acre-ft to supply enough water to increase ground-water storage. Water levels were about 10 ft higher in the fall of 1995 than in the fall of 1994 in two wells completed in the unconsolidated valley fill near Thaynes Canyon.

Greater discharge from springs in 1995 than in 1994 (fig. 9) indicates that recharge quickly causes increased ground-water levels, which increase the gradient toward and discharge from the Spiro Tunnel, two large springs, and streams. Recharge in the Thaynes Canyon area appears to contribute mostly to increased discharge, not to ground-water storage.

Ground water that flows to McLeod Creek and to the Park Meadows area probably is derived from three sources. These sources are water from the Weber Quartzite that discharges from the Spiro Tunnel, water from the Thaynes Formation that discharges from springs and that flows upward to the unconsolidated valley fill, and water from other consolidated rocks that flows upward to the unconsolidated valley fill. Monthly sampling of water for chemical analysis from McLeod Creek upstream from the confluence with White Pine Canyon and from wells in the Park Meadows area might help delineate these sources.

The water budget for the Nugget Sandstone and less permeable consolidated rocks in the White Pine Canyon area indicate a residual of about 5,000 acre-ft in 1995. Assuming all other budget components are accurate, this residual could indicate increased ground-water storage in this area, ground-water flow through the Nugget Sandstone to the Snyderville subbasin, or ground-water flow out of the study area to the Wasatch Range block. Water levels were about 25 ft higher in September 1995 than in September 1994 in well (D-2-4)6bbb-1, completed in the Twin Creek Limestone or the Nugget Sandstone in this area.

Because of its high altitude and low consumptive use, the McLeod subbasin is an important part of the hydrologic system for the entire study area. Including 800 acre-ft of municipal wastewater exported from the subbasin to the East Canyon Creek subbasin, streamflow originating in the McLeod subbasin accounted for about 40 percent of the flow leaving the study area in East Canyon Creek in 1995. Including 400 acre-ft of municipal wastewater exported from the subbasin to the Lower Silver Creek subbasin, streamflow originating in the McLeod subbasin contributed about 20 percent of the flow leaving the study area in Silver Creek. The McLeod subbasin receives about 20 percent of the recharge from precipitation for the entire study area.

Snyderville Subbasin

The Snyderville subbasin contains 10,700 acres, makes up 16 percent of the study area, and received 18 percent of the precipitation in the study area (table 3). Inflow to the subbasin was from precipitation and surface-water inflow from the McLeod and Silver Creek Junction subbasins. Outflow from the subbasin consisted mostly of water consumed in the subbasin and surface-water outflow to the East Canyon subbasin. The residuals of the total-water budget and the ground-water budget (pl. 1) indicate that ground-water storage increased by about 5,000 acre-ft if all other budget components are accurate. The residuals also could indicate that ground water is flowing to other subbasins or out of the study area. Because the water level was higher in 10 observation wells in the fall of 1995 than in the fall of 1994, was the same in 2 observation wells, and lower in 3 observation wells (Downhour and Brooks, 1996, table 3), it is difficult to determine if ground-water storage increased in the subbasin. The increases ranged from 0.8 to 25.8 ft. The decreases ranged from 7.4 to 16.4 ft.

Infiltration from streams contributed about 25 percent of the ground-water recharge in the subbasin, mainly near the consolidated-rock and unconsolidated-valley-fill contact on the west side of the subbasin. Infiltration from streams in Red Pine Canyon and Willow Draw probably contributed recharge to the Twin Creek Limestone and unconsolidated valley fill, and possibly to the Nugget Sandstone. The streams in Red Pine Canyon and Willow Draw do not enter McLeod Creek, and all flow in the streams at the contact between consolidated rock and unconsolidated valley fill is assumed to recharge the ground-water system in the unconsolidated valley fill, with the exception of consumptive use by irrigation. In Red Pine Canyon, seepage from the stream contributes water to the Twin Creek Limestone, possibly the Nugget Sandstone, and unconsolidated valley fill. This is evident in low-flow conditions when streamflow in Red Pine Canyon terminates in the area where the Twin Creek Limestone crops out. Also, weir measurements provided by Community Water Company (John Bollwinkel, written commun., 1996) indicate loss of streamflow between the upper and lower reaches of Red Pine Canyon. The annual flow at the lower weir is less than the summation of the annual flow of two branches of the creek, which indicates no additional direct runoff of precipitation between the upper and lower weirs and that some water in the stream channel infiltrates to the ground-water system between the weirs. On the basis of the weir measurements in Red Pine Canyon, streamflow probably infiltrates at altitudes higher than that of most gaging stations on other streams, including infiltration into the Nugget Sandstone and Twin Creek Limestone in Willow Draw.

Despite the large amount of infiltration of precipitation into the Twin Creek Limestone, however, ground-water discharge from the Twin Creek Limestone exceeded estimated ground-water recharge from precipitation. Recharge from precipitation to the Twin Creek Limestone in Red Pine Canyon, Willow Draw, and near Kimball Junction (pl. 2) was estimated to be about 2,000 acre-ft in 1995. The area of the Twin Creek Limestone that receives direct recharge from precipitation was determined from Ashland and others (1996, pl. 4). Discharge from the Twin Creek Limestone includes discharge to spring (D-1-3)36daa-S1 (Silver Springs) and withdrawal from public-supply well (D-1-4)19bbc-2. Discharge from Silver Springs was determined by records of public supply (David Polichette, Silver Springs Water Company, written commun., 1995 and 1996) and monthly measurements by the U.S. Geological Survey (Downhour and Brooks, 1996, table 5) to be about 2,400 acre-ft in the 1995 water year. Discharge from other springs was not significant. Discharge from the public-supply well was reported to be 140 acre-ft (Utah Department of Natural Resources, Division of Water Rights, written commun., 1996). If all budget components are accurate, discharge exceeded recharge by about 500 acre-ft in 1995. Decreased water levels in two observation wells completed in the Twin Creek Limestone (fig. 5) indicate that at least some of this difference was accounted for by water removed from ground-water storage near Kimball Junction. The remainder of the difference may be supplied by infiltration of streams in Red Pine Canyon and Willow Draw to the Twin Creek Limestone. Water also could move from the Nugget Sandstone into the Twin Creek Limestone or from the unconsolidated valley fill to the Twin Creek Limestone.

During an aquifer test done by the U.S. Geological Survey in 1985, withdrawal from the Twin Creek Limestone caused flow to cease from a spring in the unconsolidated valley fill (see "Aquifer characteristics" section of this report), which indicates that withdrawal from the Twin Creek Limestone affects ground-water levels in the unconsolidated valley fill. Discharge from Silver Springs was greater in September 1995 than in September 1994, which indicates that ground-water levels in this part of the Twin Creek Limestone were higher and that ground-water storage had increased. The increase in water levels near Silver Springs, however, does not appear to increase flow toward Kimball Junction. Recharge to the Twin Creek Limestone appears to contribute mostly to increased discharge, not to ground-water storage.

Recharge from precipitation to the Nugget Sandstone in Red Pine Canyon, Willow Draw, and near Kimball Junction (pl. 2 and Ashland and others, 1996, pl. 6) was sufficient to meet known discharge from the Nugget Sandstone. The only known discharge was testing and development of public-supply well (D-1-4)19cbd-1, small withdrawals from domestic wells, and probable ground-water flow from the Nugget Sandstone to the overlying unconsolidated valley fill. Decreased water levels in one observation well completed in the Nugget Sandstone near Kimball Junction (fig. 5), however, indicate that water was removed from storage in this consolidated rock unit. Recharge from the higher-altitude areas may not be flowing through the Nugget Sandstone toward Kimball Junction.

Natural discharge from the Twin Creek Limestone and the Nugget Sandstone in the Kimball Junction area is not known. Given the large discharge from Silver Springs, possibly the Twin Creek Limestone naturally discharges little other water. That is, annual discharge from Silver Springs may about equal annual recharge to the Twin Creek Limestone, with very little movement to Kimball Junction occurring. Ground water may flow from the Twin Creek Limestone and the Nugget Sandstone to the overlying unconsolidated valley fill, or ground water in the units may be flowing to other subbasins or out of the study area. An unnamed creek (Spring Creek) near Snyderville gained water during October 1994 (Downhour and Brooks, 1996, table 7), which could indicate that ground water from the Twin Creek Limestone discharges through the unconsolidated valley fill to the creek. The gain was greater than possible measurement errors of 20 percent. If withdrawal near Kimball Junction increases, the water removed from wells will be removed from storage until water levels decline enough to either reduce flow from the Twin Creek Limestone and the Nugget Sandstone to the unconsolidated valley fill or to induce flow from the unconsolidated valley fill to the Twin Creek Limestone and Nugget Sandstone. The extent of water-level declines and the direction of vertical ground-water flow could be estimated with well-designed aquifer tests and monitoring wells in the Twin Creek Limestone, the Nugget Sandstone, and the unconsolidated valley fill. In some places, such wells did not exist during this study; elsewhere, access to monitor and pump existing wells was not granted to the U.S. Geological Survey.

Recharge from precipitation to the Twin Creek Limestone north of Kimball Junction was estimated to be about 60 acre-ft in 1995. Natural discharge from the Twin Creek Limestone is probably to the unconsolidated valley fill. Public-supply well (D-1-4)18cda-1, completed in the Twin Creek Limestone, discharged 40 acre-ft in 1995. The source of water for the well was either removal of water from ground-water storage, the capture of almost all natural discharge (assuming that discharge approximately equals recharge), or the inducement of flow from the unconsolidated valley fill or other consolidated-rock units to the Twin Creek Limestone. Any hydraulic connections and gradients between the Twin Creek Limestone and the unconsolidated valley fill or other consolidated-rock units could not be determined because monitoring wells do not exist. Additional ground water will be removed from storage until ground-water levels decline enough to induce flow from the unconsolidated valley fill or other rock units to the Twin Creek Limestone.

Surface-water measurements (Downhour and Brooks, 1996, table 7) indicate that East Canyon Creek loses water near Kimball Junction. On October 11, 1994, the stream lost water between (D-1-4)18ddc and (D-1-4)18cbc (about 0.7 mi). The loss was greater than possible measurement errors of 20 percent. Rainstorms occurred between October 11 and October 24, 1994. On October 24, 1994, surface-water measurements also indicate that the stream lost water between (D-1-4)18ddc and (D-1-4)18cbc, but the loss was within the measurement error and may not be real loss. On July 28, 1995, the stream lost water between (D-1-4)18cbc and (D-1-3)1cdc (about 2 mi). The loss was greater than probable measurement errors of 10 percent, but less than possible measurement errors of 20 percent. Surface-water measurements also indicate a loss from this same section on September 28, 1995, but the loss was within measurement errors of 10 percent and may not be real. Surface water was not measured at intermediate locations, and the specific area of loss cannot be delineated. Stream water is infiltrating to the unconsolidated valley fill and may be moving downward to the underlying Twin Creek Limestone. The hydraulic connection and vertical gradient between the unconsolidated valley fill and the Twin Creek Limestone, however, could not be determined because monitoring wells do not exist.

The ground-water budget for the unconsolidated valley fill indicates upward flow from the consolidated rocks to the unconsolidated valley fill in most of Snyderville Basin. Recharge from precipitation to the unconsolidated valley fill was negligible in 1995. Most precipitation is consumed by plants or contributes to soil moisture. Recharge from streamflow and infiltration of irrigation water was a maximum of about 4,000 acre-ft. Ground-water discharge from the unconsolidated valley fill to streams was about 6,000 acre-ft in 1995, which is about 40 percent of the surface water originating in this subbasin. Assuming that all budget components are accurate and that no change in storage occurred, the residual of about 2,000 acre-ft may have been provided by upward flow from consolidated rock. Storage in the unconsolidated valley fill probably increased, and the amount of upward flow is probably greater than 2,000 acre-ft. Because few nonused suitable monitoring wells are completed in the unconsolidated valley fill, the amount of change in storage could not be determined. Water budgets for specific rock units indicate that most of the water for this upward flow may be provided by precipitation on the Nugget Sandstone west of the unconsolidated valley fill (pl. 2).

Infiltration from streams and ground-water discharge to streams are important processes in this subbasin, and the subbasin has the potential for much development that could affect ground-water and stream interactions. Additional monitoring wells and stream measurements would help define these processes and indicate if additional ground-water withdrawal, paving, enclosing streams in pipes, or other development are causing changes in ground-water and stream interactions. If additional withdrawals from consolidated rock reduce ground-water flow to the unconsolidated valley fill or induce ground-water flow from the unconsolidated valley fill to consolidated rocks, some of the effects of lower ground-water levels will be reduced riparian areas, possibly reduced crop production on subirrigated areas, and reduced ground-water discharge to streams. If ground-water flow is induced from the unconsolidated valley fill to consolidated rocks, the possibility of contamination of public-supply wells would increase.

Silver Creek Junction Subbasin

The Silver Creek Junction subbasin contains 7,700 acres, makes up 12 percent of the study area, and received 10 percent of the precipitation on the study area (table 3). The only known inflow was from precipitation in the subbasin. Outflow from the subbasin consisted of water consumed within the subbasin and surface-water outflow to the Snyderville subbasin. Because of low altitude and less precipitation than other areas, about 75 percent of the precipitation on this subbasin is consumed. The residuals in the total-water budget (table 3) and the ground-water budget (pl. 1) indicate that ground-water storage increased by about 1,000 acre-ft if all other budget components are accurate. Water levels were higher in two observation wells and lower in one observation well in September 1995 than in September 1994. The increases were 10 ft and 19 ft. The decrease was 1.2 ft.

Direct runoff of precipitation to streams in the subbasin contributed about 70 percent of the surface water in the subbasin. Ground-water discharge to streams contributed the remaining 30 percent of the surface water in the subbasin.

Only about 10 percent of the precipitation in this subbasin becomes ground-water recharge, but that accounts for 96 percent of the recharge for the subbasin. The only other source of recharge is infiltration from septic tanks. Recharge from septic tanks was estimated to be 70 acre-ft on the basis of an estimated domestic water use in the subbasin of about 80 acre-ft. Little water is used for lawns and gardens; most enters septic tanks and infiltrates to the ground-water system. Because recharge from precipitation is difficult to estimate and has large errors, recharge to the subbasin could be substantially different than estimated. Because water levels rose, however, the recharge is probably not significantly less than estimated. During years of less-than-normal precipitation, it is possible that all precipitation is consumed or runs off to streams and that ground-water recharge from precipitation is negligible.

Ground-water discharge to streams is the only natural discharge known. Ground-water withdrawal from wells, therefore, either removes water from storage or decreases ground-water discharge to streams. Withdrawal from areas not near streams may decrease water levels throughout a large area until water levels near the stream decline enough to reduce the discharge to streams.

Because of the low altitude and high consumptive use, surface water originating in the Silver Creek Junction subbasin contributed less than 10 percent of the streamflow leaving the study area in East Canyon Creek, even though the subbasin makes up about 17 percent of the area contributing to East Canyon Creek.

East Canyon Subbasin

The East Canyon subbasin contains 17,100 acres, makes up 26 percent of the study area, and received 29 percent of the precipitation on the study area (table 3). Inflow to the subbasin was precipitation on the subbasin, surface-water flow into the subbasin, and imported municipal wastewater from other subbasins (table 3). Surface-water inflow was estimated by summing the amount of water at upstream gages (table 2) and estimating the amount of ungaged direct runoff and ground-water discharge to McLeod Creek below the McLeod Creek near Park City, Utah, gage and upstream from this subbasin boundary. Outflow from the subbasin consisted of surface-water outflow and water consumed within the subbasin. The residuals in the total-water budget (table 3) and ground-water budget (pl. 1) indicate that ground-water storage increased by about 20,000 acre-ft during the 1995 water year if all other budget components are accurate. The water level was higher in the fall of 1995 than in the fall of 1994 in three observation wells completed in the Twin Creek Limestone and in two observation wells completed in the Thaynes Formation. Water levels were slightly lower in two observation wells probably completed in the Nugget Sandstone. The increases ranged from 1 to 31 ft, and the decreases were 0.5 and 1 ft. Ground-water storage probably increased during the 1995 water year.

The amount of precipitation in the subbasin and, therefore, the amount of ground-water recharge from precipitation, could have large errors. The northern part of the subbasin was outside the area of snowmelt simulation (see "Energy-balance snowmelt simulation" section of this report) and the precipitation on that part was estimated on the basis of snowmelt simulation for the rest of the subbasin. The northern part is typically lower in altitude than the southern part, and precipitation may have been overestimated. Ground-water may flow out of the study area from the subbasin. The area south of Interstate Highway 80 is extensively faulted (pl. 2), and water may enter those faults and flow into the deeper ground-water system in the Wasatch Range block. The area north of Interstate Highway 80 is underlain mostly by sedimentary rocks that dip north (Bryant, 1990, sh. 1) and discharge ground water only to small springs. Ground water in this area may be flowing north from the study area in consolidated rocks.

Even though this subbasin makes up about 40 percent of the drainage area for East Canyon Creek, only about 25 percent of the streamflow in East Canyon Creek is generated in this subbasin. Ground-water discharge to streams contributed about 40 percent of the surface water and is an important part of the surface-water flow in this subbasin. Ground-water recharge from infiltration from streams is insignificant (pl. 1) throughout the subbasin but may be significant in local areas near East Canyon Creek.

The interaction of ground water with East Canyon Creek is complex and varied. The following table lists measurement sites (Downhour and Brooks, 1996, table 7), date of measurement, whether East Canyon Creek gained or lost water, and if the gain or loss was greater than probable measurement errors of 10 percent. If the gain or loss was less than 10 percent of the flow in the stream, the gain or loss may or may not be actual. Because the flow in East Canyon Creek was not measured at intermediate locations, more precise locations of gains and losses could not be delineated. On the basis of these measurements, East Canyon Creek appears to be losing water to the unconsolidated valley fill between (D-1-4)18cbc and (D-1-3)1cdc and possibly between (D-1-3)1cdc and (D-1-3)2bbd. East Canyon Creek appears to be gaining water from the unconsolidated valley fill between (D-1-3)2bbd and (A-1-3)27dbc. These gains and losses may vary during annual and seasonal ground-water-level fluctuations. Ground water possibly may move downward from the unconsolidated valley fill to underlying consolidated rocks in some areas and upward from consolidated rocks to the unconsolidated valley fill in other areas. The hydraulic connection and vertical gradient between the unconsolidated valley fill and underlying consolidated rocks, however, could not be determined because suitable monitoring wells do not exist. Additional surface-water measurements throughout the year and installation of monitoring wells at various depths in the unconsolidated valley fill and underlying consolidated rocks would permit a better understanding of the interactions between ground water and East Canyon Creek.

The Twin Creek Limestone and Nugget Sandstone in the Summit Park area may be hydrologically isolated by a topographic divide and shales to the southeast. Recharge to these units may be limited to recharge from precipitation in the Summit Park area and was estimated to be 5,400 acre-ft in 1995. The area of Twin Creek Limestone and Nugget Sandstone that receives recharge from precipitation was determined from Ashland and others (1996, pl. 4 and pl. 6). Natural discharge from the Twin Creek Limestone and Nugget Sandstone appears to be small. No significant springs discharge water from these units. Ground water in the units probably discharges to the creek in Toll Canyon and to the overlying unconsolidated valley fill near East Canyon Creek. Ground water also may flow to the deeper ground-water system in the Wasatch Range. Ground-water withdrawal from wells in these units will remove water from storage until discharge to streams, the unconsolidated valley fill, or deeper systems is reduced. If low-permeability boundaries, such as faults, shale layers, or clay layers in the unconsolidated valley fill prevent the reduction of natural discharge or the inducement of infiltration of streams, withdrawal will continue to reduce ground-water storage and ground-water levels. Seasonal water-level fluctuations in well (D-1-3)9caa-1, completed in the Twin Creek Limestone, of about 60 ft indicate that the storage coefficient in this area is low and that additional withdrawals could affect water levels throughout a large area. Seasonal water-level fluctuations in well (D-1-3)15acb-1, also completed in the Twin Creek Limestone, are only about 10 ft, which indicates a higher storage coefficient, less recharge or stress on the system in this area, a possible moderation of ground-water levels near the creek in

Measurement section		Date	Gain or Loss	Greater than or within measurement error
To	From			
(D-1-4)18cbc	(A-1-3)34daa	May 1, 1995	gain	greater
(D-1-4)18cbc	(D-1-3)1cdc	July 28, 1995	loss	greater
(D-1-4)18cbc	(D-1-3)1cdc	September 28, 1995	loss	within
(D-1-3)1cdc	(D-1-3)2bbd	July 28, 1995	loss	greater
(D-1-3)1cbd	(D-1-3)2bbd	October 6, 1995	gain	within
(D-1-3)2bbd	(A-1-3)27dbc	July 28, 1995	gain	greater
(D-1-3)2bbd	(A-1-3)34daa	October 11, 1995	gain	within
(D-1-3)34daa	(A-1-3)27dbc	October 11, 1995	gain	within

Toll Canyon, or a combination of these factors. The water level in the well is close to the altitude of the stream, but the altitudes were not determined during this study. Accurate altitudes would help to determine the direction of flow between the creek and the ground-water system in this area. Additional ground-water withdrawals could reduce ground-water levels enough to reduce ground-water discharge to the stream or induce additional ground-water recharge from the stream.

Recharge from precipitation to the Thaynes Formation near Pinebrook was estimated to be 3,500 acre-ft in 1995. The area of the Thaynes Formation receiving recharge from precipitation was determined from Ashland and others (1996, pl. 8). This area has extensive faulting. Only a few unused monitoring wells exist, however, and the hydrologic connection of the Thaynes Formation across the faults could not be determined. Natural discharge from the Thaynes Formation occurs to spring (D-1-3)14bcd-S1 (Two Mile Springs) and probably to creeks and the overlying unconsolidated valley fill near East Canyon Creek. Ground water also may flow to deeper ground-water systems. Ground-water withdrawal from wells in the Thaynes Formation will remove water from storage until discharge to streams, the unconsolidated valley fill, or deeper systems is reduced. If low-permeability boundaries, such as faults, shale layers, or clay layers in the unconsolidated valley fill prevent the reduction of natural discharge or the inducement of infiltration of streams, withdrawal will continue to reduce ground-water storage and ground-water levels.

An aquifer test in February 1996 used well (D-1-3)12cca-1, completed in the Thaynes Formation, as the pumped well (see "Aquifer characteristics" section of this report). During the test, water levels in the pumped well were lowered to about 400 ft below the altitude of East Canyon Creek, and water levels in well (D-1-3)11ddb-1, also completed in the Thaynes Formation, were lowered to about 200 ft below the altitude of East Canyon Creek. These levels indicate the potential to induce flow from East Canyon Creek, but because monitoring wells in the unconsolidated valley fill did not exist, vertical gradients could not be determined. Water levels in well (D-1-3)13abb-2, also completed in the Thaynes Formation, were not affected and remained at about the same altitude as East Canyon Creek. Accurate altitudes of well (D-1-3)13abb-2 and East Canyon Creek would help determine the direction of flow between the ground-water system in the Thaynes Formation and East Canyon Creek. Water levels in well (D-1-3)13abb-2 may have been unaffected because of the short pumping time, preferred flow direction in fractures, or because faults in the area act as hydrologic boundaries.

Because the processes of natural discharge before withdrawal from wells began in this subbasin are not known, the effects of ground-water withdrawal are difficult to estimate. All withdrawal, however, must be met by a change in storage, a reduction in natural discharge, or an increase of infiltration from streams. Monitoring wells near streams, with accurately determined altitudes, would help determine the direction of flow from the ground-water system to streams. Water levels during pumping in many production wells are below the altitude of streams and could induce flow through the unconsolidated valley fill to the consolidated rocks. Withdrawal also may reduce flow from the consolidated rocks to the unconsolidated valley fill, which may reduce water levels in the valley fill and cause decreased ground-water discharge to streams and reduce the extent of riparian areas.

Upper Silver Creek Subbasin

The Upper Silver Creek subbasin contains 6,500 acres, makes up 10 percent of the study area, and received 10 percent of the precipitation on the study area (table 3). Inflow to the subbasin was precipitation on the subbasin, surface-water inflow from the McLeod subbasin, ground-water inflow from south of the study area, and ground-water inflow from the McLeod subbasin. Outflow from the subbasin consisted mainly of water consumed in the subbasin, water that flowed to the Provo River drainage in the Ontario #2 Drain Tunnel, surface-water outflow, and export of municipal wastewater to the Lower Silver Creek subbasin. The ground-water budget (pl. 1) was balanced by assuming that ground water flowed into the subbasin from south of the study area and from the McLeod subbasin. Because the inflow was calculated as the residual of the other budget components, the amount of inflow incorporates all errors in the other budget components. Water levels increased from September 1994 to September 1995 in two observation wells completed in unconsolidated valley fill overlying the Park City Formation or Woodside Shale south of the Thaynes Formation (Bryant, 1990, sh. 1), and decreased in two observation wells completed in unconsolidated valley fill overlying the Thaynes Formation. The increases were 1.1 and 3.2 ft and the decreases were 0.5 and 3 ft. These limited data indicate little change in ground-water storage. Assuming that all other budget components are accurate, ground-water inflow was required to prevent a decrease in storage.

Direct runoff of precipitation to streams was estimated using hydrograph-separation techniques for data from the gaging station on Silver Creek near Wanship. The direct-runoff component of the hydrograph was separated into Upper and Lower Silver Creek subbasins on the basis of area and season of snowmelt. Surface-water flow within the subbasin was estimated from flume readings provided by Park City Water Department (Rich Hilbert, Park City Water Department, written commun., 1996) and measurements made during this study (Downhour and Brooks, 1996, tables 6 and 7).

Ground-water recharge from infiltration of streams is negligible in the Upper Silver Creek subbasin. Mason (1989, p. 12) reported that infiltration from Silver Creek recharged the unconsolidated valley fill. Measurements made during this study, however, indicate little infiltration from Silver Creek. Water levels in the unconsolidated valley fill and consolidated rocks during measurements of Silver Creek were higher in this study than in the previous study by Mason (1989), which may have decreased the gradient and recharge from Silver Creek. Ground-water recharge from irrigation with municipal water is higher than in any other subbasin. Recharge from irrigation with municipal water is insignificant in comparison to possible errors in other budget components, but may be significant in the Park Meadows and Prospector Square areas. Runoff and recharge from municipal irrigation probably contribute to the riparian areas in Park Meadows and near Silver Creek.

The Ontario #2 Drain Tunnel discharged 10,700 acre-ft in the 1995 water year (Utah Department of Natural Resources, Division of Water Rights, written commun., 1996). Weber Quartzite in the study area and other rocks south of the study area make up about 85 percent of the area that could be contributing to the tunnel discharge. Therefore, 85 percent (about 9,000 acre-ft in the 1995 water year) of the discharge was assumed to originate in the study area or south of the study area. The remaining 15 percent of discharge was assumed to originate in the Provo River drainage. Errors in this estimate are small compared to errors in other budget estimates and are associated with measurement errors of tunnel discharge and the estimate of 85 percent of the discharge originating in the Upper Silver Creek subbasin or south of the subbasin.

In addition to the 9,000 acre-ft of discharge from the Ontario #2 Drain Tunnel, the Judge Tunnel discharged about 1,600 acre-ft in the 1995 water year (Rich Hilbert, Park City Water Department, written commun., 1995 and 1996). Recharge from precipitation to the consolidated rocks in the subbasin overlying the tunnels was about 8,000 acre-ft in 1995. The area of the consolidated rocks that receives recharge from precipitation on the subbasin was determined from Ashland and others (1996, pl. 10). Assuming that budget components are accurate and ground-water storage did not change, about 3,000 acre-ft of flow in the tunnels may be from precipitation south of the study area that flows into the study area either in consolidated rocks or in mine tunnels.

Recharge from precipitation to consolidated rocks not overlying the mine tunnels and unconsolidated valley fill in the subbasin was about 1,400 acre-ft. Recharge to the unconsolidated valley fill from irrigation with surface water and municipal water was about 400 acre-ft. Ground-water discharge from these consolidated rocks and unconsolidated valley fill was about 500 acre-ft to well (D-2-4)4dda-1, 500 acre-ft to spring (D-2-4)4dca-S1 (Dority Springs), and 2,000 acre-ft to streams. The residual of about 1,000 acre-ft more discharge than recharge is assumed to be provided by ground-water flow from the McLeod subbasin to this subbasin.

Limited water-level data (Downhour and Brooks, 1996, table 3) indicate the gradient is upward from the Woodside Shale to the overlying unconsolidated valley fill; therefore, the unconsolidated valley fill probably receives water from the Woodside Shale. The Thaynes Formation possibly contributes water to the overlying unconsolidated valley fill during the spring months but may receive recharge from the unconsolidated valley fill during late summer and fall months. The hydraulic gradient is downward in the unconsolidated valley fill near Prospector Square (Mason, 1989, p. 12). During short-term pumping, water flows downward from the unconsolidated valley fill into the Thaynes Formation and discharge to streams may be reduced (Mason, 1989, p. 33). The effects of continuous pumping for more than 7 days have not been determined during a long-term aquifer test, but water-quality data indicate that water with higher sulfate, chloride, and dissolved-solids concentrations flows downward into the Thaynes Formation. The higher sulfate concentration indicates possible downward movement of water discharging from the Spiro Tunnel into ditches and streams that flow to the Upper Silver Creek subbasin. Water-quality analyses indicate that a mixture of water from the Thaynes Formation in Thaynes Canyon, surface water from the Spiro Tunnel, and water from the Woodside Shale produces the water withdrawn from public-supply well (D-2-4)4dda-1, completed in the Thaynes Formation in Upper Silver Creek subbasin. No other public-supply wells were in the subbasin during this study.

Ground-water withdrawal from wells will be met by reductions in ground-water storage until ground-water levels are lowered enough to decrease discharge to springs and streams. Withdrawals in areas not near springs and streams in the low-altitude part of the subbasin will decrease water levels between the area of withdrawal and the springs and streams. Recharge to the Weber Quartzite occurs primarily from precipitation and could not be increased, so any water withdrawn from wells in the Weber Quartzite must be balanced by a decrease in storage or reduction in discharge. Very little natural discharge from the Weber Quartzite is known. Natural discharge before mining began is not known. Withdrawal from the Weber Quartzite may reduce ground-water storage and ground-water levels until flow from other consolidated rocks could be induced or discharge to the mine tunnels decreased. No wells exist in the Weber Quartzite to determine current water levels or hydraulic gradient.

Lower Silver Creek Subbasin

The Lower Silver Creek subbasin contains 13,700 acres, makes up 21 percent of the study area, and received 15 percent of the precipitation in the study area (table 3). Inflow to the subbasin was precipitation on the subbasin, surface-water flow from the Upper Silver Creek subbasin, and municipal wastewater imported from the McLeod and Upper Silver Creek subbasins. Outflow from the subbasin consisted mainly of water consumed in the subbasin and surface-water outflow. The residuals in the total-water budget (table 3) and the ground-water budget (pl. 1) indicate little change in ground-water storage during the 1995 water year if all other budget components are accurate. Water levels in two monitoring wells increased from September 1994 to September 1995 by about 2 ft and 7 ft. Because one well was near the south boundary and one well was near the west boundary, water-level changes throughout the subbasin are not known. Ground-water storage possibly increased and one or more of the budget components may be inaccurate.

Ground-water recharge from precipitation to the Lower Silver Creek subbasin was about 3 percent of recharge for the study area and 50 percent of recharge for this subbasin. Recharge from infiltration of irrigation water makes up about 50 percent of the recharge in this subbasin, mainly because the flow in Silver Creek is diverted to a canal along the east side of the subbasin and is used to flood irrigate fields. Much of the irrigation water infiltrates the ground along the east side but is discharged from the ground-water system to the stream in the lower altitudes.

Recharge from precipitation to the Twin Creek Limestone in this subbasin (pl. 2) was about 20 acre-ft in 1995. Well (D-1-4)21ddd-1 completed in the Twin Creek Limestone discharged 160 acre-ft in 1995 (Utah Department of Natural Resources, Division of Water Rights, written commun., 1996), indicating that additional ground water moves into the Twin Creek Limestone or that water is removed from storage. Inflow could come from overlying or underlying formations, or from the Snyderville subbasin through the Twin Creek Limestone. Flow into the rock unit is probably through fractures.

Because the only known natural discharge in this subbasin is to streams, ground-water withdrawal from wells is met by a reduction in ground-water storage until water levels decline enough to reduce ground-water discharge to streams or induce ground-water recharge from streams. Lowered water levels in the unconsolidated valley fill would decrease ground-water discharge to streams and reduce the extent of riparian areas.

Study Area

Inflow to the study area is from precipitation on the study area and estimated ground-water inflow across the south boundary of the study area through consolidated rocks, mine tunnels, or fractures that intersect mine tunnels (table 3). To determine the hydraulic gradient and amount of flow through the consolidated rocks, observation wells would be required in each formation near the boundary. To determine aquifer characteristics, a large production well and several observation wells would be required in each formation near the boundary. These wells do not exist. Ground-water inflow across the south boundary of the study area was estimated to be the residuals of the ground-water budgets in specific rock units in the McLeod and Upper Silver Creek subbasins and therefore incorporates all errors in inflows, outflows, and change in soil moisture or ground-water storage in those units. The error in this estimate could be large, but because ground-water inflow is much smaller than recharge from precipitation, the error probably is not significant in the ground-water budget.

Water leaves the study area by interception, use, and evaporation of precipitation and soil moisture by plants; surface-water flow; mine-tunnel flow; consumptive use of ground water, surface water, and irrigation water by plants; sublimation from snow; and consumptive use of artificial snow (table 3). About 50 percent of the water that entered the study area was consumed in the study area. Surface water leaves the study area through East Canyon Creek and Silver Creek. The flow in these streams includes flow from two wastewater-treatment plants and therefore includes nonconsumptive municipal use. The residuals of the total-water budget and the ground-water budget indicate that ground-water storage increased by about 33,000 acre-ft if all other budget components are accurate. Ground water also may flow out of the study area. If ground-water flow is leaving the study area, ground-water budgets indicate most of it would be from the East Canyon subbasin.

Ground-water recharge from precipitation makes up about 80 percent of the ground-water recharge in the study area. Because of the high percentage of recharge from precipitation, ground-water levels and ground-water discharge to springs and streams are highly dependent upon precipitation. The effects of precipitation are noticed in the ground-water system during the same water year. Greater- or less-than-normal precipitation during the winter and spring affects ground-water levels and discharge in the spring and summer of that same year.

The largest component of discharge from the ground-water system is discharge to streams. Ground-water discharge to streams contributes about 40 percent of the surface water in the study area. Ground-water discharge to springs and mine tunnels contributes about 25 percent of the surface water in the study area. Most spring discharge is from four springs in the Thaynes Formation and one spring in the Twin Creek Limestone (fig. 9). Several small springs also discharge ground water. Most of these small springs were measured or observed to have no discharge at least once during the study (Downhour and Brooks, 1996, table 5), and overall discharge was negligible. The discharge from some small springs is included as discharge to streams. Additional use of ground water has the potential to decrease discharge to streams and affect both the amount and quality of surface water in the study area, but much of the water used returns to the surface- or ground-water system.

The low-altitude Silver Creek Junction and Lower Silver Creek subbasins together contributed only about 15 percent of the surface water and 5 percent of the ground-water recharge from precipitation in the study area, even though they make up 33 percent of the area. In contrast, the McLeod subbasin contributed about 35 percent of the surface water and about 25 percent of the ground-water recharge from precipitation in the study area, even though it makes up only about 14 percent of the area. The other subbasins contributed surface water and ground-water recharge in approximate proportion to their area.

Water Budgets for the 1994 Water Year

Because detailed data collection did not start until May 1994, water budgets for 1994 could not be determined monthly. Instead, the annual total-water budget, surface-water budget, and ground-water budget were determined for the entire study area and are listed in table 4. Water budgets were not determined for the six subbasins. As explained in the "Climate" section of this report, precipitation in 1994 was less than normal, but was closer to normal than was the greater-than-normal precipitation in 1995.

Precipitation was determined for the water year by the same methods described in the "Methods" section of this report but was not determined monthly. Sublimation from snow was determined from the snowmelt simulation and was greater than in 1995 for reasons explained in the "Energy-balance snowmelt simulation" section of this report. Plant demand of precipitation was assumed to be the same as in 1995, but actual use was less because less precipitation was available. In 1995, plant use and runoff to streams could be supplied by precipitation. In 1994, however, precipitation did not provide enough water to meet plant demand and to allow the same percentage of precipitation to become runoff as in 1995. Determining whether plant use or runoff occurs first was beyond the scope of this study. This report assumes that plant demand must be met before runoff occurs.

Ground-water recharge from infiltration of streams was assumed to be the same as in 1995. Less surface water was available in 1994 and contribution to ground water may have been less, but ground-water levels were lower, increasing the hydraulic gradient out

Table 4. Estimated water budgets in Snyderville Basin, Park City, and adjacent areas, Utah, 1994

[All flows in acre-feet]

Total-water budget			
Inflow		**Outflow**	
Precipitation	135,000	Plant precipitation use and evaporation	95,000
Surface-water inflow	0	Sublimation	4,000
Ground-water inflow from south of the study area	[1]7,000	Surface-water outflow	22,000
		Mine-tunnel flow to the Provo River drainage	9,000
		Crop, lawn, and riparian use of ground water, surface water, and municipal water	[1]4,000
		Consumptive use from artificial snow	60
Total	**142,000**	**Total (rounded)**	**134,000**
Residual[2]			**8,000**

Surface-water budget			
Inflow		**Outflow**	
Surface-water inflow	0	Surface-water outflow	22,000
Runoff from precipitation	12,000	Ground-water recharge from streams	6,000
Ground-water discharge to streams	10,000	Ground-water recharge from irrigation with surface water	3,000
Spring flow contribution to surface water	3,000	Crop and riparian consumptive use of surface water	[1]1,000
Mine-tunnel flow contribution to surface water	6,000		
Wastewater-treatment plant contribution to surface water	2,700		
Return flow from municipal irrigation	[1]700		
Total (rounded)	**35,000**	**Total**	**32,000**
Residual[2]			**3,000**

Ground-water budget			
Recharge		**Discharge**	
Ground-water inflow	[1]7,000	Discharge to streams	10,000
Recharge from precipitation	25,000	Discharge to mine tunnel to Provo River drainage	9,000
Recharge from streams	6,000	Discharge to mine tunnels that becomes surface water	6,000
Recharge from irrigation with surface water	3,000	Discharge to springs that becomes surface water	3,000
Recharge from irrigation with municipal water	600	Crop and riparian consumptive use of ground water	[1]2,000
Recharge from septic tanks	70	Discharge to wells	2,600
Recharge from irrigation with ground water	20	Discharge to mine tunnels for municipal supply	2,400
		Discharge to springs for municipal supply	1,400
Total (rounded)	**42,000**	**Total (rounded)**	**36,000**
Residual[2]			**6,000**

[1]Assumed to be the same as in 1995.

[2]The residual includes the net error of inflows and outflows, change in soil moisture and ground-water storage during the 1995 water year, ground-water flow between subbasins, and ground-water outflow from the study area. A positive residual could indicate an increase in soil moisture or ground-water storage, ground-water flow out of the subbasin or study area, an overestimate of ground-water recharge and ground-water inflow, or an underestimate of ground-water discharge or ground-water outflow. A negative residual could indicate a decrease in soil moisture or ground-water storage, ground-water flow into the study area, an underestimate of ground-water recharge and ground-water inflow, or an overestimate of ground-water discharge or ground-water outflow.

of streams and possibly increasing infiltration from streams.

Crop, lawn, and riparian consumptive use of ground water, surface water, and municipal water were assumed to be the same in 1994 as in 1995. Return flow from irrigation with municipal water and ground-water recharge from irrigation with surface water, ground water, and municipal water were assumed to be the same as in 1995. Less precipitation available to plants would have increased the consumptive use of applied water, but factors such as less surface water, lower ground-water levels, and municipal watering restrictions would have decreased the consumptive use from these sources.

Because ground-water budgets were not computed for subbasins, ground-water inflow needed from south of the area to produce balanced budgets in the McLeod and Upper Silver Creek subbasins could not be determined. Ground-water flow into the study area across the south boundary was assumed to be the same as in 1995.

The residuals in the total-water budget and the ground-water budget indicate that ground-water storage increased by about 7,000 acre-ft if all other budget components were accurate. Water levels in the study area, however, were lower in September 1994 than in September 1993 (figs. 6 and 7), which indicates a reduction in ground-water storage during the 1994 water year. The apparently incorrect residual could be caused by errors in estimating budget components, or by not accounting for ground-water flow out of the study area to the north or to deeper flow systems in the Wasatch Range. Ground-water flow out of the study area would increase the outflow from the study area and reduce the residual in the water budgets. The surface-water budget also has a residual, which indicates errors in estimating budget components.

Comparison of 1995 to 1994 Water Budgets

A comparison of the 1995 to 1994 water budgets indicates that the hydrologic system in the study area is dependent upon the amount of annual precipitation and has low capacity for ground-water storage. Major budget components and the amount by which the components for the 1995 water year exceeded components for the 1994 water year are listed in table 5.

Consumptive use of precipitation, surface water, and ground water was not substantially greater in 1995 than in 1994. This is because most of the use is by natural vegetation, which is adapted to the area, and the maximum need for plants was met without using all of the additional water available in 1995. The use was less in 1994 because water was not available to meet the maximum required by plants. Because consumptive use remains relatively constant regardless of precipitation, when precipitation is greater than normal, consumptive use is a smaller proportion of the total-water budget, and a larger proportion of water is available to become surface water or to recharge the ground-water system. When precipitation is less than normal, consumptive use is a larger proportion of the total-water budget, and a smaller proportion of water is available to become surface water or to recharge the ground-water

Table 5. Major water-budget components for 1995 and 1994 in Snyderville Basin, Park City, and adjacent areas, Utah

Budget component	1995 amount as a percentage of 1994 amount	1995 amount in excess of 1994 amount (acre-feet)
Precipitation	150	69,000
Consumptive use of precipitation, surface water, and ground water	110	11,000
Runoff from precipitation	210	13,000
Surface-water outflow	245	32,000
Ground-water recharge from precipitation	275	44,000
Ground-water discharge to streams	230	13,000
Ground-water discharge to springs	200	4,000
Ground-water discharge to mine tunnels	105	1,000

system. Because consumptive use remains relatively constant and is a large proportion of the total-water budget, direct runoff of precipitation to streams and ground-water recharge from precipitation are not directly proportional to precipitation. Direct runoff of precipitation to streams and ground-water recharge from precipitation both increased by greater percentages than precipitation (table 5). Also, during years of less-than-normal precipitation, direct runoff and ground-water recharge will both decrease by greater percentages than does precipitation.

Surface-water outflow from the study area in 1995 was more than double the surface-water outflow in 1994. The large variation in surface-water outflow significantly affects the proportion of streamflow that is contributed by the two wastewater-treatment plants. Discharge from the East Canyon Creek treatment plant made up 9 percent of the flow leaving the study area in East Canyon Creek in 1994, and 4 percent of the flow in 1995. Discharge from the Silver Creek treatment plant made up 26 percent of the flow leaving the study area in Silver Creek in 1994, and 14 percent of the flow in 1995.

Ground-water recharge from precipitation was almost three times more in 1995 than in 1994. Ground-water discharge to streams and springs was about two times more in 1995 than in 1994. The increased discharge to springs and streams indicates that much of the additional recharge in 1995 caused increased discharge and did not remain stored in the ground-water system. Water-level fluctuations in monitoring wells (figs. 5, 6, and 7) also indicate that much of the water caused increased discharge and did not remain in storage. Although the water level in several wells was higher in the fall of 1995 than in 1994, the water level in several wells was substantially lower in the fall of 1995 than in the late spring and early summer of 1995. Recharge from precipitation raised ground-water levels, but increased discharge to springs and streams decreased ground-water levels during the summer. The water-level fluctuations and variation in discharge to springs and streams indicate that, in general, the storage coefficient is small throughout the study area. Discharge to streams occurs mainly from the unconsolidated valley fill. The large increase in discharge, therefore, indicates that the unconsolidated valley fill in the study area does not provide substantial ground-water storage. Ground-water levels and ground-water discharge are dependent upon annual precipitation and differ substantially from year to year.

Discharge to mine tunnels in 1995 was similar to discharge in 1994 (table 5). At least two factors may contribute to the small variation in annual ground-water discharge to tunnels. One factor is that mine tunnels and fractures intersecting mine tunnels could substantially increase secondary porosity and storage near the mine tunnels. The increased storage could cause smaller water-level changes than elsewhere in the study area, and the hydraulic gradient toward tunnels may not change substantially. The second factor is that discharge from mine tunnels may be somewhat controlled by bulkheads and portals or is pumped to the Ontario #2 Drain Tunnel. If discharge is controlled and cannot substantially increase, water levels in the consolidated rocks containing the tunnels would remain higher for a longer period of time than in other parts of the study area. It is likely that both of these factors, and possibly other factors that are not understood, contribute to the small variation in ground-water discharge to tunnels with variation in ground-water recharge. Because of the small variation in ground-water discharge to tunnels, the proportion of the total ground-water discharge that was discharge to tunnels was lower in 1995 than in 1994.

Although precipitation in the study area was much greater in 1995 than in 1994, most of the additional water caused additional outflow from the study area and did not remain as increased ground-water storage. About 60 percent of the extra precipitation in 1995 than in 1994 was either consumed in the study area or left the study area as surface-water outflow. The remaining 40 percent probably increased ground-water storage. Because data collection did not continue beyond September 1995, however, it is not possible to determine how rapidly the remaining water discharged from the ground-water system.

ENERGY-BALANCE SNOWMELT SIMULATION

Snowmelt runoff was simulated to estimate ground-water recharge to consolidated-rock and unconsolidated valley-fill aquifers in Snyderville Basin, Park City, and adjacent areas. A topographically distributed snowmelt model controlled by independent inputs of net radiation, meteorological parameters, and snowcover properties is used to calculate the energy and mass balance of the snowcover (Marks, 1988; Marks and Dozier, 1992). The model is topographically distributed over a digital elevation model (DEM), and the snowcover energy and mass balance is calculated at each grid cell of the DEM. The model area consists of the study area except where DEMs were not available for the part of the study area north of 40^{0} 45' latitude, and that area was not included in the snowmelt runoff simulations.

The model simulates melt in two snowcover layers, simulates runoff from the base of the snowcover, and adjusts the snowcover mass, thermal properties, and measurement heights at each time step. The modeling approach is an adaptation of the model developed by Marks (1988), and extended over a topographic grid (D. Marks, U.S. Geological Survey, written commun., 1997); it is similar to those used by Anderson (1976), Morris (1982, 1986), and Jordan (1991). The model subdivides the snowcover into two layers: a surface layer of constant thickness, and a lower layer made up of the rest of the snowcover. The surface layer is considered the active layer with its thickness set to the approximate depth of significant solar radiation penetration. All surface energy transfer occurs in this layer. Both layers are assumed to be homogeneous and are characterized by an average temperature, density, and liquid-water content.

The model assumes that energy is transferred between the surface layer and the lower layer, and between the lower layer and the soil by conduction and diffusion. At each time step, the model calculates the energy balance, the snow-surface temperature, and then adjusts the temperature and specific mass of each layer. If the calculated energy budget is negative, the cold content, or the energy required to bring the temperature of the snowcover to 32 °F, is increased, and layer temperatures decrease. If the energy budget is positive, cold content is decreased until the temperature of the snowcover is 32 °F. Additional input of energy causes the model to predict melt. If melt occurs, it is assumed to displace air in the snowcover, causing densification, and increasing the average liquid-water content of both layers. Liquid water in excess of a specified threshold becomes predicted runoff. Though meltwater is typically generated in the surface layer, mass lost to runoff is removed from the lower layer. The thickness of the surface layer remains constant until the lower layer is completely melted. At that time, the model treats the snowcover as a single layer. The physical equations solved by the model and the model structure are explained in Marks (1988), Marks and Dozier (1992), and van Heeswijk and others (1995). The reader is referred to these publications for detailed descriptions of the energy-balance snowmelt model.

Snowmelt runoff was simulated for the study area for March through June in 1994 and 1995. March through June was selected because this is when most snowmelt occurs and peak streamflow and groundwater recharge associated with snowmelt occur. The model time step was 3 hours to allow the simulation of diurnal variations of temperature and solar radiation which are important climatic factors controlling snowmelt. A model grid size of 246 ft was selected because it smoothed poor-quality 98.4 ft-data while preserving the topographic structure of the area and was computationally more efficient than using a smaller grid cell size.

Hydrometeorological Data

Data from four Natural Resource Conservation Service SNOTEL stations, two National Weather Service (NWS) cooperative network stations, the Salt Lake City Airport, a Utah Department of Air Quality (UDAQ) monitoring-network station, and a U.S Geological Survey (USGS) reference climate station at the Park City Mountain Resort were used to generate spatially distributed climate data surfaces necessary to drive the model. A data surface is a spatially distributed representation of the hydrometeorological parameter with a data value for each grid cell.

The locations of climate data stations are listed below and shown on plate 1 with the exception of Brighton, Mill D, and Salt Lake Airport which are west of the study area. The Parleys Summit SNOTEL station is just outside the border of the model DEM and was used only for comparison with the generated data surfaces in that area. The Brighton, Mill D, Thaynes Canyon, Park City Fire Station, and Snyderville stations were used to generate climate surfaces. The Salt Lake City Airport, Cottonwood Air Monitoring Station, and USGS reference station at the Park City Mountain Resort were used to develop lapse rates for estimating climate parameters.

Station	Latitude (degrees)	Longitude (degrees)	Altitude (feet)
Brighton, SNOTEL	40.599161	111.582686	8,750
Mill D, SNOTEL	40.658169	111.636673	8,960
Thaynes Canyon, SNOTEL	40.620168	111.532840	9,327
Parleys Summit, SNOTEL	40.762000	111.628500	7,500
Park City Fire Station	40.666667	111.50	6,909
Snyderville	40.70	111.533	9,088
Salt Lake City Airport	40.76667	111.96667	4,055
Cottonwood Air Monitoring	40.644667	111.84972	4,380
USGS reference site at the Park City Mountain Resort	40.6380	11.5192	8,648

Distributed hydrometeorological data surfaces were generated from point measurements at the data stations, from model simulations, and from lapse rates developed between point measurements, and then distributed over the DEM. Methods for generating data surfaces will be discussed for each model variable.

Solar and Thermal Radiation

Clear-sky solar radiation was simulated (Dozier, 1980; Dubayah and others, 1990; Marks and others, 1991) and then corrected for estimated cloud-cover effects (Hungerford and others, 1989). Clear-sky thermal radiation from the atmosphere is simulated from the altitude and the air and dew-point temperatures and is corrected for topographic effects (Marks and Dozier, 1979). Estimates of cloud cover on the basis of precipitation data were used to correct the calculated clear-sky thermal radiation for the effects of clouds.

Precipitation

Three SNOTEL stations and the Snyderville National Weather Service Cooperative station were used to create the 3-hour precipitation surfaces in a two-step process. First, daily precipitation surfaces were calculated using the detrended kriging algorithm of Garen and others (1994) and Garen (1995) with a few enhancements. The enhancements of the basic procedure used were:

1. Days were aggregated into storm periods instead of fixed length periods,
2. Least-absolute errors instead of least-squares regression were used to calculate trend lines, and
3. Negative regression lines were screened out.

The steps in the calculations are described in Garen and Marks (1996). Second, 3-hour precipitation surfaces were derived by a simple fractioning approach. The 3-hour fraction of the daily precipitation falling at the SNOTEL stations with hourly data was calculated and averaged. These fractions were then subjectively lumped and smoothed to produce a daily set of eight fractions. Lumping and smoothing were required to ensure that 24 hourly totals and daily precipitation agreed. The fractions were multiplied by daily-precipitation surfaces to produce the 3-hour surfaces.

Precipitation Density

Precipitation density for each 3-hour period was calculated as a function of the dew-point temperature for the 3-hour period or the daily-minimum temperature. Dew-point temperatures were calculated from the vapor-pressure surfaces. Precipitation density and the amount of precipitation that is snow is calculated by:

Temperature (T) (degrees Fahrenheit)	Snow (percent)	Snow density (pounds per cubic foot)
$T < 23$	100	4.65
$23 \le T < 26.6$	100	6.20
$26.6 \le T < 29.3$	100	9.30
$29.3 \le T < 31.1$	100	10.85
$31.1 \le T < 32$	75	12.40
$32 \le T < 32.9$	25	15.50
$32.9 \le T$	0	0

Temperature

The four climate stations used for calculating the precipitation surfaces and the Park City Fire Station were used to generate the temperature surfaces. The detrended kriging algorithm used to calculate precipitation surfaces also was used to distribute daily maximum and minimum temperature by interpolating among the five stations with the following modifications:

1. The days were aggregated into 5-day fixed length periods instead of storm periods,
2. Least-squares regression was used to calculate trend lines, and
3. Positive regression lines were screened out.

The 3-hour temperature surfaces were obtained by passing an average diurnal cycle through the maximum and minimum temperature. The diurnal cycle was calculated using a procedure similar to that used in the National Weather Service HYDRO-17 snow model (Anderson, 1976; Garen and Marks, 1996). The temperature of each time period was calculated as a weighted sum of the maximum and minimum temperature surrounding the time interval.

Vapor Pressure

Vapor-pressure 3-hour data surfaces were calculated using an altitude-lapse rate because vapor-pressure data were not available at the climate stations in the model area. The Cottonwood air-monitoring site and the USGS Park City reference site were used to calculate the lapse rate using 206 days of data in 1996. The data from the Cottonwood site were then used to develop vapor-pressure surfaces for the 1995 model runs. Because the Cottonwood site was installed late in 1994, data were not available for the 1994 model runs; therefore, data from the Salt Lake City Airport were used. The lapse rate was calculated as the difference between the vapor pressure at the two sites divided by the altitude difference between the sites. To calculate the lapse rates, the data from the sites were smoothed with a 7-day moving average to eliminate the local effects at the sites and yet preserve the effects of major air-mass changes. The difference between the smoothed data from the USGS reference site and the Cottonwood site was normalized to the Cottonwood station and was fit with a linear-regression model to the Cottonwood data (fig. 13). The resulting equation for the difference between the sites as a function of the Cottonwood data was then divided by the difference in altitude between the sites to derive the lapse rate (eq. 3). Thus, vapor-pressure surfaces were calculated by multiplying the difference in altitude of a grid cell and the Cottonwood site by the lapse rate and subtracting that from the daily vapor pressure at the Cottonwood site. The 3-hour surfaces were calculated by linear interpolation between the daily surfaces. Vapor-pressure surfaces were converted to dew-point-temperature surfaces for use in calculating precipitation densities and thermal radiation. Daily vapor pressures for each grid cell were calculated by:

$$lapse\ rate = (0.52CPV - 41.15)/rdz \quad (3)$$

$$vapor\ pressure = CPV - (lapse\ rate \times pdz) \quad (4)$$

where

CPV = Daily average vapor pressure at Cottonwood air-quality monitoring site (in pascals),

rdz = Park City reference site altitude minus Cottonwood site altitude (in feet), and

pdz = Grid-cell altitude minus Cottonwood site altitude (in feet).

The measured vapor pressure at the Cottonwood site and Park City reference site, and the calculated vapor pressure at the Park City reference site, are shown in figure 14. The calculated vapor pressure at the Park City reference site varied from about 200 to 600 Pa (pascals), which is the expected range of vapor pressures. The objective of these calculations was not to recreate the data exactly, but to create data surfaces that had reasonable values and that expressed variance of the data. There are two time periods of data separated by a period of no data when the instrumentation was not functioning at the Park City reference site. During the first time period, the mean and standard deviation for the measured and calculated vapor pressure were 287 Pa, 73 Pa and 285 Pa, 73 Pa, respectively. During the second time period, instrument problems early in the time period resulted in the measured vapor pressure exceeding 1,000 Pa at the reference site and exceeding the vapor pressure at the Cottonwood site. The data from this time period were not used in developing the regression. The calculated vapor pressure remained in the 200 Pa to 600 Pa range in the second time period.

Wind Speed

Wind-speed 3-hour surfaces were calculated by a procedure similar to that used to calculate vapor pressures because wind-speed data were not available at the climate sites in the model area. The ratios of 7-day moving-average daily wind speed to measured average daily wind speed at the USGS Park City reference site and at the Salt Lake City Airport were compared using a least-squares linear-regression model to develop a function to estimate USGS Park City reference-site data as a function of the Salt Lake City Airport data (fig. 15). The mean daily wind speed at a grid cell was estimated by multiplying the mean lapse rate by the calculated ratio of the 7-day moving-average daily wind speed to daily average wind speed at the Park City reference site ("Y" on fig. 15) and multiplying the result by the altitude of the grid cell (eq. 5). The mean lapse rate is the mean of the differences between the mean daily wind speeds at the USGS Park City reference site and at the Salt Lake City Airport divided by the altitude difference between the two sites. Three-hour wind speeds were estimated by linear interpolation between the mean daily wind speeds.

Mean daily wind speeds at a grid cell were calculated by:

$$wind\ speed = mlapse\ (1.40881\ x - 0.25984)\ z \quad (5)$$

where

$mlapse$ = mean lapse rate, 0.001089 (mi/hr)/ft,

x = ratio of 7-day moving-average daily wind speed to daily average wind speed at Salt Lake City Airport (dimensionless) and

z = grid cell altitude (in feet).

Because this procedure initially underestimated the wind speeds at the Salt Lake City Airport and at the USGS Park City reference station, the lapse rate was

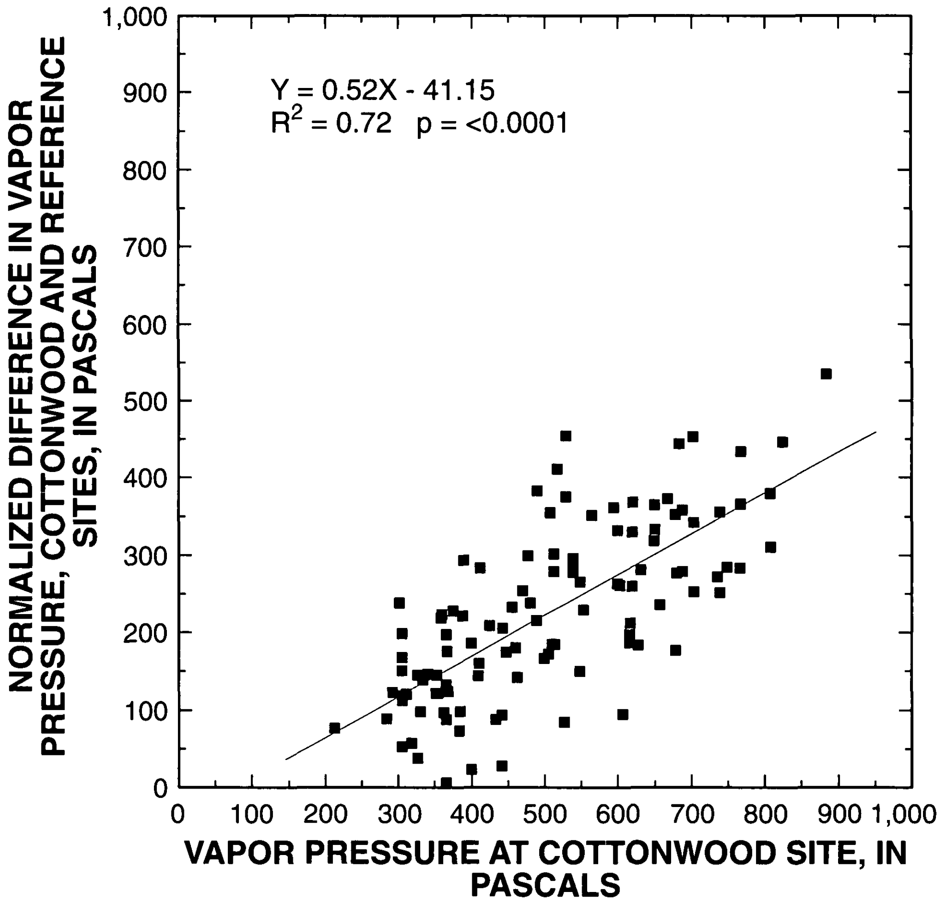


Figure 13. Linear regression of normalized difference in vapor pressure at Cottonwood site and Park City reference site with vapor pressure at Cottonwood site, Utah.

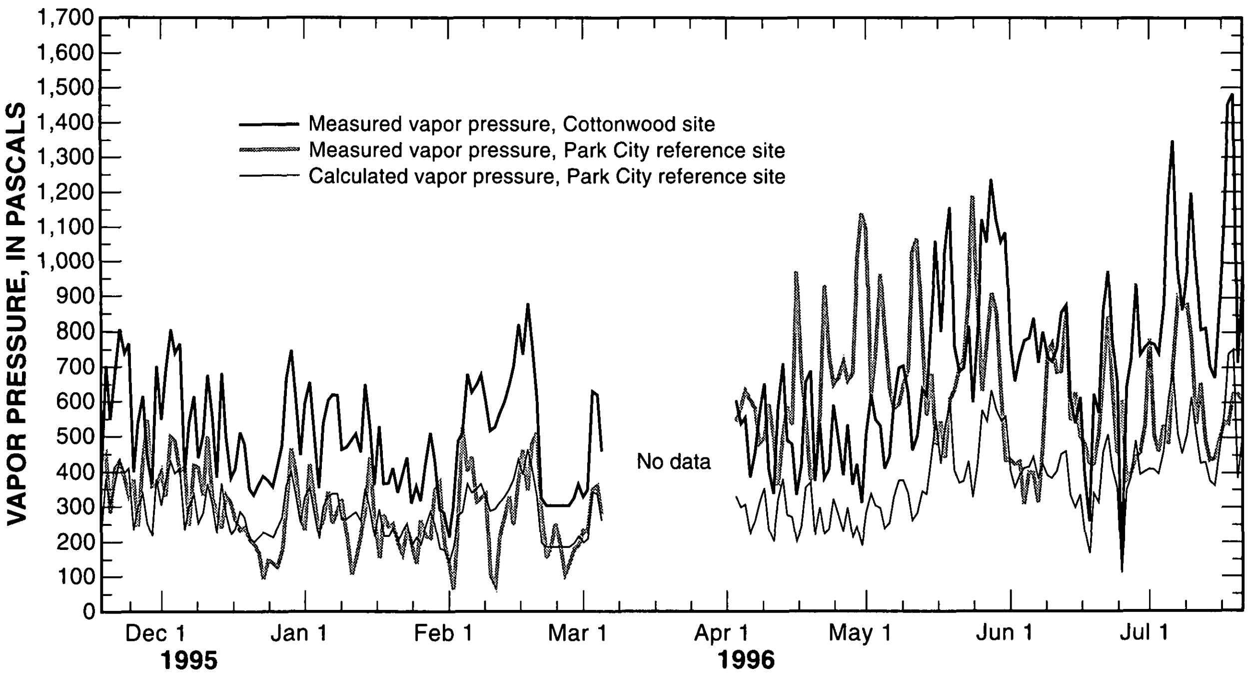


Figure 14. Measured and calculated vapor pressure at Cottonwood site and Park City reference site, Utah.

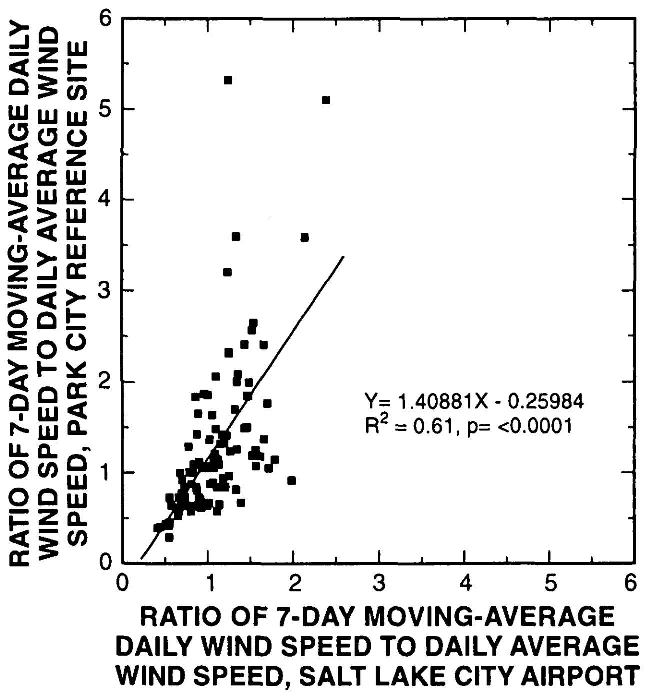


Figure 15. Linear regression of ratio of 7-day moving-average daily wind speed to daily wind speed at Park City reference site with ratio of 7-day moving-average daily wind speed to daily wind speed, Salt Lake City Airport, Utah, December 17, 1995, to March 3, 1996.

increased by trial and error from 0.000699 to 0.001089 (mi/hr)/ft until it yielded calculated mean daily wind speeds closer to measured mean daily measured wind speeds at the two stations. The measured and corrected calculated wind speeds at the Park City reference site are shown in figure 16. Again, the objective of the estimation procedures is to develop data surfaces that have reasonable values and express some of the variance of the measured data.

Snow Water Equivalent

The distributed snow water equivalent (SWE), the depth of water in inches that would result from melting the snowcover, on March 1 of 1994 and 1995 was estimated from the three SNOTEL stations using the same detrended kriging procedure as used for estimating precipitation and temperature without any enhancements. All grid cells in the model area had snowcover at these times. The SWE surface was then adjusted on the basis of the topographic aspect of the grid cell. East-facing (azimuth 45°-135°), west-facing (azimuth 225°-315°), and flat slopes were not adjusted. North-facing slopes (azimuth 0°-45° and 315°-360°) were adjusted upward by multiplying the original SWE estimate by 1.2. South-facing slopes (azimuth 135°-225°) were adjusted downward by multiplying the original SWE by 0.8. These adjustment values were selected on the basis of differences observed in the field between the snow depth and mass on different aspects. The March 1 SWE was the initial condition for the energy-balance snowmelt simulations.

Snow Density, Temperature, and Depth

Snow density, temperature, and depth were measured in snow pits dug every 4 to 6 weeks during 1995 through the melt season at the USGS Park City reference site. Mean snow densities for snow pits dug in February and March 1995 were about 18.7 lb/ft^3. Initial snow densities for the model simulations were linearly interpolated on the basis of measurements of snow density at a range of altitudes with set interpolation points of 21.8 lb/ft^3 at 4,922 ft, 18.7 lb/ft^3 at 8,203 ft, and 15.6 lb/ft^3 at 13,124 ft . Upper- and lower-layer snow temperatures were set with a similar procedure on the basis of snow temperatures from snow pits dug during February and March. Lower-layer snow temperatures were linearly interpolated on the basis of altitude between 32.0 °F at 5,577 ft, 30.2 °F at 8,203 ft, and 28.4 °F at 13,124 ft. Upper-layer snow temperatures were linearly interpolated with points of 30.2 °F at 4,922 ft, 28.4 °F at 8,203 ft, and 26.6 °F at 13,124 ft. The initial snow depth in each grid cell was the value of the density in the cell divided by the SWE for that cell.

Results

Simulated specific snow mass and depth were compared to SWE data from the three SNOTEL sites and snow-depth data from the Snyderville National Weather Service Cooperative site. Specific snow mass in pounds per square foot is equivalent to SWE at a point and is referred to as SWE. Times series of SWE and depth for grid cells were extracted from the simulated specific snow-mass and depth surfaces and were compared to the data from the climate sites. Daily specific snow-mass surfaces were compiled into digital image movies, and the spatial distribution of specific snow mass was compared to field observations of snowmelt and snow distribution to evaluate the spatial distribution of simulated snowmelt. This was a qualitative and subjective measure of model performance because it did not compare simulated spatial snowmelt to actual snow distribution observed from, for example, aerial photographs.

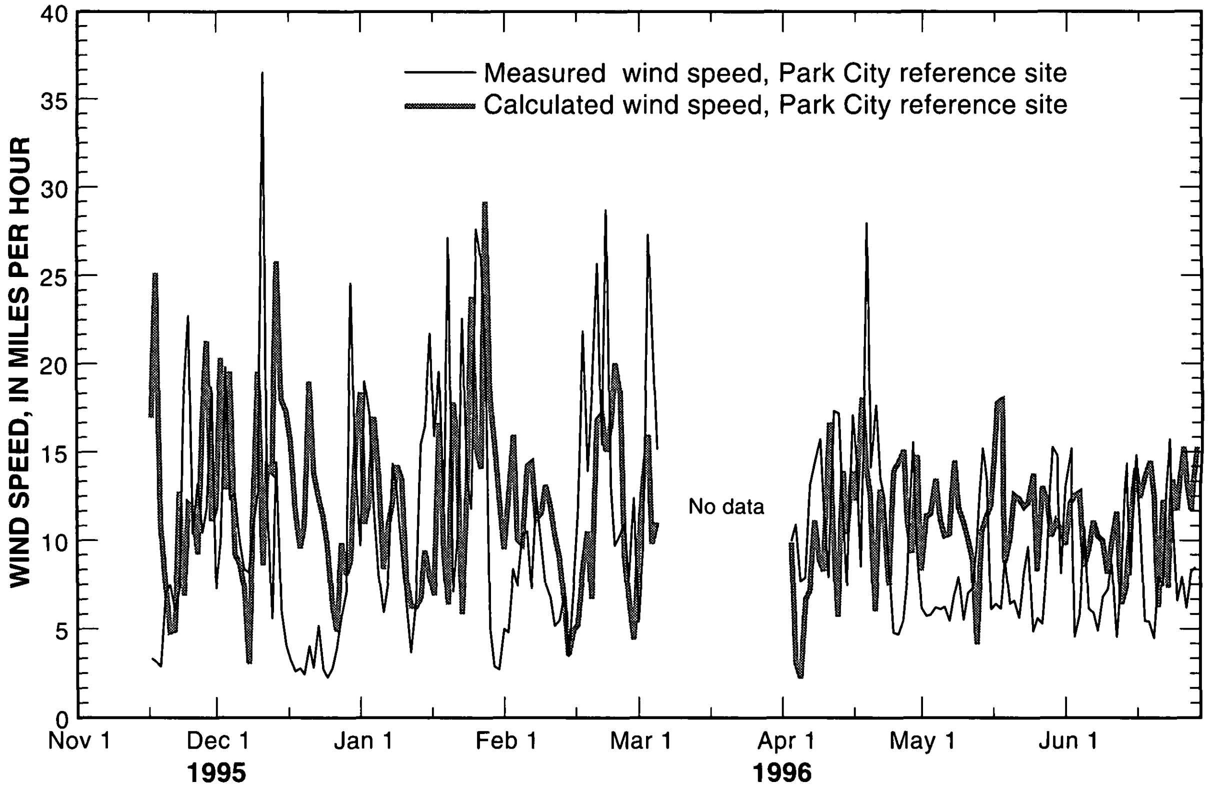


Figure 16. Measured and calculated average daily wind speed, Park City reference site, Park City, Utah.

Daily measured and simulated SWE from the Brighton, Mill D, and Thaynes Canyon SNOTEL sites is shown in figure 17 for 1994 and figure 18 for 1995. Overall, the simulated SWE compared favorably with the daily SWE at the SNOTEL sites. The difference between simulated and actual melt-off dates varies from 1 to 8 days. At the Brighton site, the simulated SWE closely tracked the actual SWE data. In 1995 at the Mill D and Thaynes Canyon sites, the model accumulated less SWE and melted it off earlier than indicated by the site data. The differences between the model performance at the SNOTEL sites may be attributable to several factors. The first factor is the accuracy of the location of SNOTEL sites. The Thaynes Canyon site was the most accurately located using global-positioning-system technology and was verified on the DEM. The other two sites were not as accurately located. When a 246-ft grid is used, if a location is off by a few grid cells, the site could be located on a different aspect and modeled snow accumulation and melt can be affected. The second factor is the effect of vegetation, which is not accounted for by the model. Shading by the vegetation canopy affects snow accumulation and melt. The third factor is the rain shadow over the crest of the Wasatch Range from west to east. The Brighton and Mill D sites on the west side of the divide, at lower altitudes, receive as much or more precipitation than the Thaynes Canyon site, which is at a higher altitude but on the east side of the divide.

The simulated spatial distribution of SWE failed to show the effects of the rain shadow. Early model runs did not melt the snow from the lower-altitude areas on the east side of the model area until mid to late May. Field observations showed that snow typically melted and was gone from these areas by early to mid-April. This was the result of not accounting for the effects of the rain shadow in the distribution of the precipitation and initial snow conditions. To account for the rain shadow in the model area, the precipitation was decreased below a specified altitude threshold by an exponential decay function. Altitude was normalized to a threshold of 8,530 ft, and precipitation at all altitudes below the threshold was decreased while precipitation above the threshold was unchanged. This is an imperfect solution but worked fairly well for the simulated area because few altitudes were below the selected threshold of 8,530 ft on the west side of the model area. A more accurate solution would be to use both altitude and distance from the crest of the Wasatch Range when

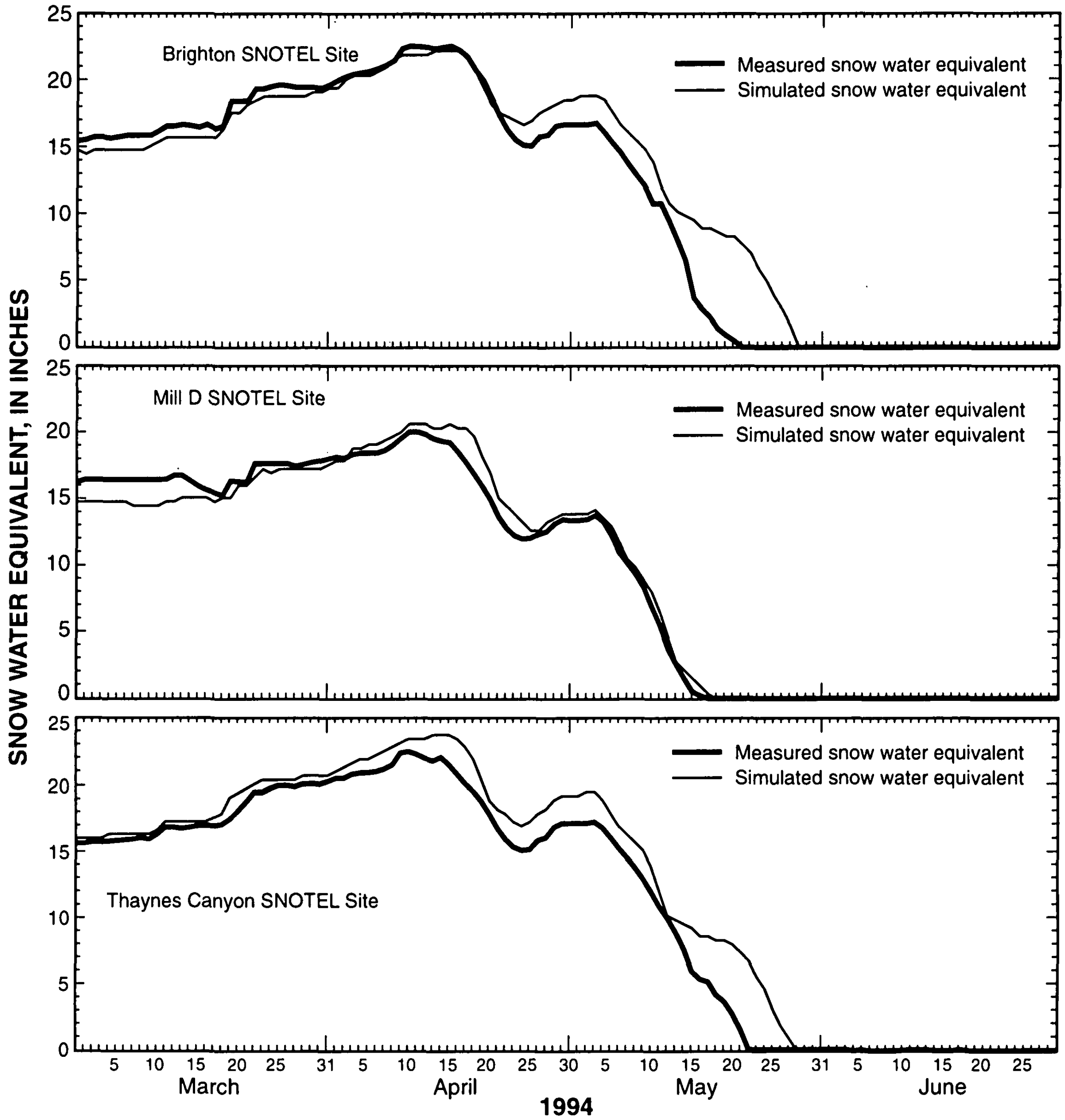


Figure 17. Measured and simulated snow water equivalent at selected SNOTEL sites, Wasatch Range, Utah, March to June 1994.

distributing precipitation. After accounting for the rain shadow, the spatial distribution of SWE showed the snow completely melting from the lower-altitude areas on the east side of the model area in early to mid-April as expected (fig. 19).

The output from the snowmelt model includes a complete set of parameters describing the snowcover energy balance and snowcover mass balance. The parameters of interest in this study are the evaporation and sublimation from the snowcover, snowmelt runoff, and snowcover mass that are equal to the snow water equivalent (SWE). Evaporation and sublimation from the snowcover is controlled by the vapor-pressure gradient between the snow surface and the atmosphere. Vapor pressures are calculated with an altitude-based lapse rate and data from low-altitude climate stations. These estimates of vapor pressure may overestimate high-altitude vapor pressures late in the melt season. Modeled evaporation and sublimation rates would be less than actual rates in this case.

Snowmelt runoff is melt water leaving the snowcover and includes rain if the snowcover-energy balance is at 32 °F. In this case, rain passes through the snowcover and becomes part of the snowmelt runoff.

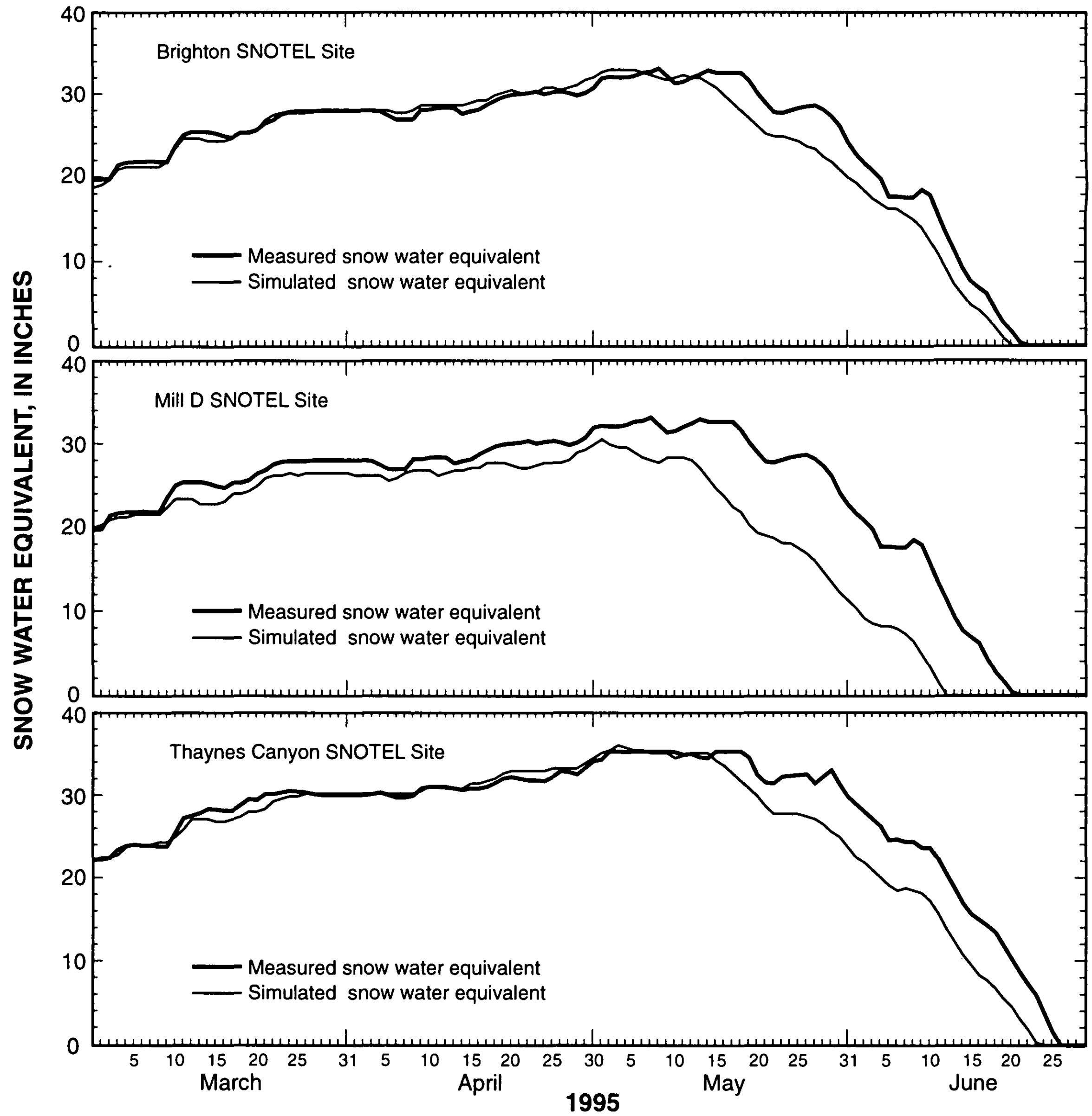


Figure 18. Measured and simulated snow water equivalent at selected SNOTEL sites, Wasatch Range, Utah, March to June 1995.

Rain falling on bare ground also is accounted for as snowmelt runoff by the model. Simulated daily snowmelt runoff and sublimation plus evaporation for the study area for March to June 1994 and 1995 are shown in figure 20. The total simulated sublimation plus evaporation was 3,339 acre-ft in 1994 and 1,937 acre-ft in 1995, and the total simulated snowmelt runoff was 41,618 acre-ft in 1994 and 73,621 acre-ft in 1995. These amounts do not include the area north of 40° 45' latitude.

The spatial distribution of snow mass was simulated by the model and daily digital images of snow mass were created. Selected snow-mass images for March 1, April 1, May 1, June 1, and June 30, the end of the simulation, for 1994 and 1995 show the differences in both accumulation of snow mass and snowmelt between the years (fig. 19). The maximum simulated snow mass occurred about April 12-15, 1994, and about May 1-5, 1995. In 1995, the higher altitudes continued to accumulate snow into early May while snow at the lower altitudes melted in a similar pattern as in 1994. The images also show the areas where snowmelt is generated throughout the melt season.

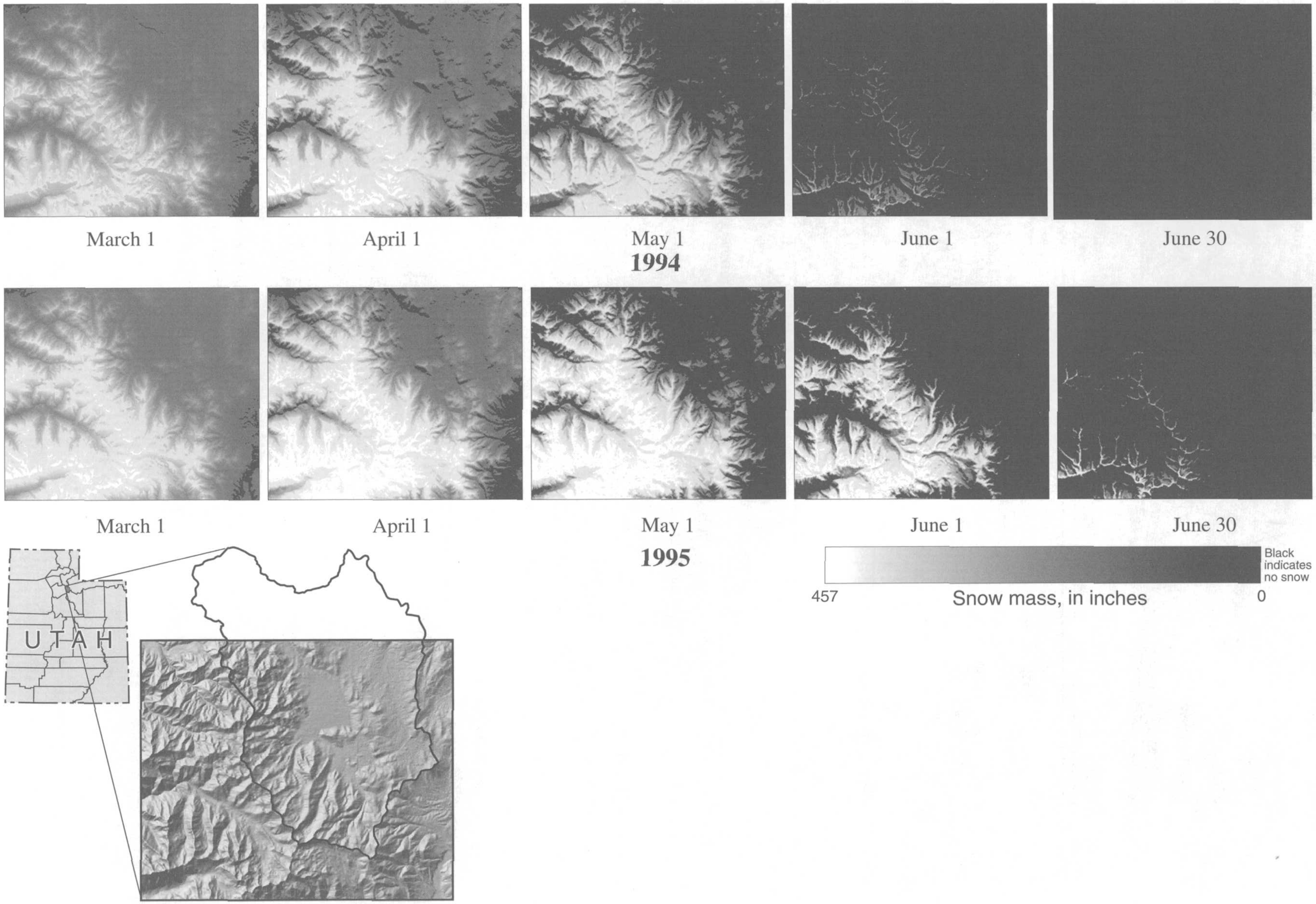


Figure 19. Simulated snow-mass distribution in Snyderville Basin, Park City, and adjacent areas, Utah, 1994 and 1995.

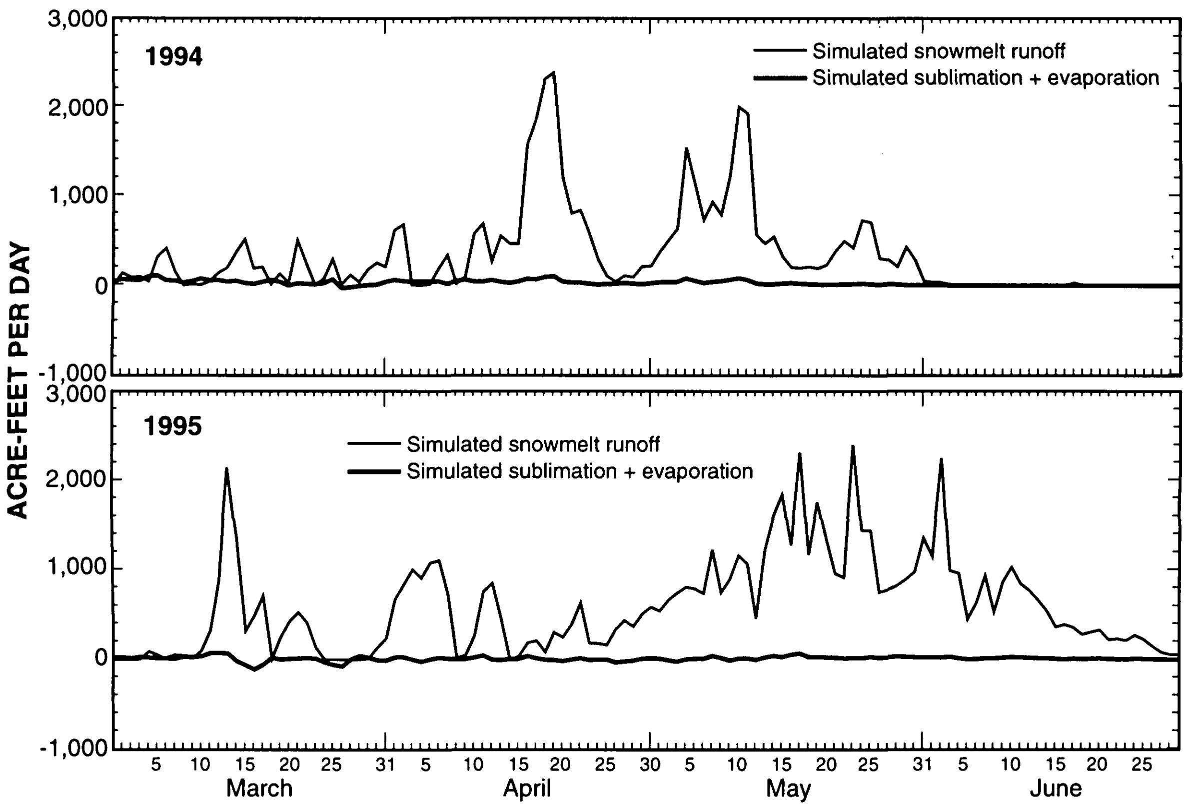


Figure 20. Simulated daily snowmelt runoff and sublimation plus evaporation, Snyderville Basin, Park City, and adjacent areas, Utah, March to June 1994 and 1995.

Simulated snowmelt runoff, evaporation, and sublimation for March to June 1994 and 1995 were used in estimating ground-water recharge as explained in the "Water-budget analysis" section of this report. Daily streamflow data were available for the McLeod subbasin (pl. 1), and a daily budget was compiled for this basin for March to June 1994 and 1995. Daily surface runoff to McLeod Creek, simulated snowmelt runoff, and the difference between daily snowmelt runoff and surface runoff are shown in figure 21. The difference is the water available for ground-water recharge and for consumptive use by vegetation and for soil moisture. Daily streamflow in McLeod Creek shows little response to snowmelt, which is atypical. The McLeod subbasin is underlain by highly fractured limestone and quartzite that allow rapid infiltration of water. Thus, most of the snowmelt runoff in the McLeod subbasin infiltrates to soil moisture or ground-water aquifers, is sublimated or evaporated, or is transpired by vegetation. The water-budget totals for the McLeod subbasin for March through June 1994 and 1995 are:

Budget element	Water-budget total (acre-feet)	
	1994	1995
Snowmelt runoff	14,500	25,100
Ground-water recharge	10,600	18,400
Evapotranspiration and soil moisture	3,500	3,500
Surface runoff from McLeod Creek	420	3,000

Evapotranspiration and soil moisture are estimated from vegetation and soil types and distributions. Surface runoff from McLeod Creek is the streamflow from overland and unsaturated flow and does not include flow contributed to the stream from ground water or mine tunnels. About 73 percent of the snowmelt runoff and spring rainfall in 1994 and 1995 is ground-water recharge. The methods used for compiling ground-water budgets are explained in the "Water-budget analysis" section of this report.

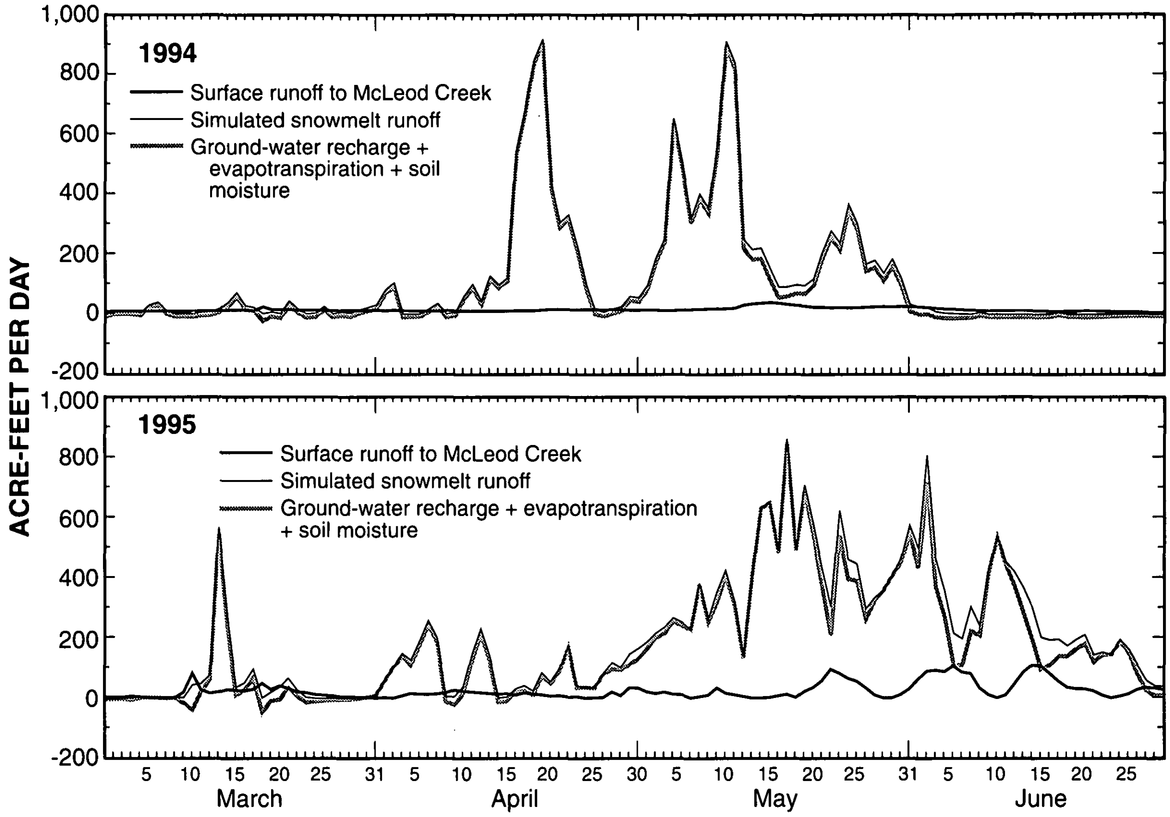


Figure 21. Surface runoff in McLeod Creek, simulated snowmelt runoff, and water available for ground-water recharge, evapotranspiration, and soil moisture, Park City, Utah, March to June 1994 and 1995.

Model Limitations

The topographically distributed energy-balance snowmelt model does not currently have a runoff routing module. Snowmelt runoff therefore must be routed through the basin by other methods. Coupling the snowmelt model with a runoff-routing model would provide a more quantitative systematic method for routing runoff through the basin.

Detailed hydrometeorological data are required to drive the model, and if hourly measurements are not available, then data sets must be created. In this application of the model, 3-hour vapor-pressure, dew-point, wind-speed, and temperature data all were developed from daily data. Daily diurnal cycles can be lost or masked in this process and care must be taken to avoid this. Data also must be spatially distributed. This process also can introduce errors if the number of data sites is limited and if gradients across the model areas are strong. Both of these conditions exist in the model area. The rain shadow across the crest of the Wasatch Range required modifications of the precipitation surfaces. Data were insufficient for some model parameters, such as solar radiation, so these parameters were simulated, which is less desirable than actual data.

FUTURE STUDIES

The effects of ground-water withdrawals on streamflow can best be determined with long-term streamflow records. Continuing data collection at existing stream-gaging stations will allow future analysis and long-term comparison of streamflow to snowpack. If the natural streamflow declines or if the relation between precipitation and streamflow changes, then ground-water withdrawals are affecting the amount of seepage to streams from unconsolidated valley fill.

As ground-water withdrawals increase, more precise estimates of recharge to and discharge from specific consolidated-rock formations may be desired. Additional stream-gaging stations established for several years in the upper reaches of tributary drainages would help determine recharge, at least qualitatively, to each of the four primary water-bearing consolidated-rock formations. Measurements of streamflow just upstream and downstream from an outcrop can determine the percentage of streamflow that recharges a specific consolidated-rock formation. Repeating this procedure for each consolidated-rock formation in more than one drainage would permit an average value to be determined. Similarly, additional stream-gaging stations in lower reaches would define more precisely where consolidated-rock formations discharge water into unconsolidated valley fill and thus, into streams.

Concurrent with streamflow monitoring, snowmelt data could be collected for snowmelt-simulation purposes. Additional simulations would have two benefits. Combined recharge estimates from streams and snowmelt simulation would provide a more detailed estimate of recharge to specific consolidated-rock units. Calculated recharge to these units and measured yearly change in water levels would provide a means to estimate storage. Additional snowmelt simulations that represent a more normal snowpack than those simulated during this study would provide better estimates of average runoff and sublimation.

Because of the location of water-bearing consolidated-rock formations and residential development, spatially distributed wells that are suitable for monitoring and are completed in known consolidated-rock formations or unconsolidated valley fill are rare. Additional monitoring wells installed with known depths and completions in specific formations and not used to withdraw ground water on a regular basis could be used to monitor water levels. Monitoring water levels and water quality would help to obtain accurate data with regard to the effects of ground-water withdrawals on consolidated-rock formations and overlying unconsolidated valley fill. This especially would be helpful in areas where declining ground-water levels from increasing ground-water withdrawals can result in decreased streamflow. Long-term water-level monitoring in several wells throughout the study area would identify areas that are impacted by additional withdrawals. Water-level declines may be more noticeable if the wells are measured at least twice a year, once in May and once in September or October.

Aquifer testing using large production wells pumped for at least 7 days with appropriately located monitoring wells would help to understand the interaction between consolidated-rock formations. If necessary, additional monitoring wells could be included to produce the best test results. With local cooperation, this testing could be combined with well-production testing and testing conducted as part of source protection.

SUMMARY AND CONCLUSIONS

Increasing residential and commercial development are placing increased demands on the ground- and surface-water resources of Snyderville Basin, Park City, and adjacent areas in the southwestern corner of Summit County, Utah. The U.S. Geological Survey, in cooperation with the Utah Department of Natural Resources, Division of Water Rights; Park City; Summit County; and the Weber Basin Water Conservancy District, completed a study in which data collected during 1993-95 were used to assess the quantity and quality of the water resources of the area.

Surface water originates in the Wasatch Range on the southern and western borders of the study area and leaves in two streams to the north. Streamflow is seasonal. During this study, 70 to 100 percent of the streamflow at gaged sites occurred during March through July. Ground-water seepage to streams is a large component of streamflow leaving the study area, especially during the remaining months. If ground-water seepage to streams is reduced because of declining water levels, streamflow could be reduced unless replenished from other sources.

The consolidated rocks and unconsolidated valley fill in the study area form a heterogeneous, anisotropic, interconnected ground-water system. The four principal water-bearing consolidated-rock formations are the Twin Creek Limestone, Nugget Sandstone, Thaynes Formation, and Weber Quartzite. Complex geology and the lack of spatially distributed water-level data make it difficult to determine the degree of connection between blocks of consolidated rock and between consolidated rock and unconsolidated valley fill.

Recharge from infiltration of snowmelt is the largest source of recharge in the study area, whereas recharge from rainfall in the summer and fall months is negligible. The rapid increase in discharge to streams and springs that results from the recharge effects of snowmelt is indicative of a ground-water system with little storage. This increase is a pressure response to the infiltration of water from snowmelt into the ground-water system and is not direct discharge of newly melted snow.

All public-supply wells in the study area are completed in consolidated rocks. Wells completed in unconsolidated valley fill typically produce sufficient water for domestic use for a single household but probably would not produce sufficient water for public supply. Withdrawal from wells can affect ground-water levels on a seasonal basis. Increased ground-water withdrawals from 1983 to 1995, however, generally have not affected ground-water levels throughout most of the study area, with two exceptions. Increased ground-water withdrawals for testing and production in 1994 and 1995 near Kimball Junction caused seasonal fluctuations and peak water levels to be lower in 1995 than in 1994 despite greater recharge in 1995. Similarly, ground-water withdrawals in the Park Meadows area resulted in water-level declines. In both areas, ground-water withdrawals may have induced downward movement of water from unconsolidated valley fill to consolidated rocks, and withdrawal from some wells affects discharge from nearby springs. No suitable monitoring wells are in either area to verify the downward movement.

The chemical composition of ground water in the study area primarily is influenced by the lithology of the consolidated rocks through which it flows. Dissolution and weathering of limestone and sandstone contribute calcium, magnesium, and bicarbonate to the water. Dissolution of gypsum in shale or gypsiferous limestone contributes calcium and sulfate to the water. Dissolution of road salt contributes sodium and chloride to much of the ground water in unconsolidated valley fill. Water from wells and springs generally has a dissolved-solids concentration that ranges from 200 to 600 mg/L. Water samples from wells, springs, and drains near Park City generally have higher dissolved-solids concentrations than ground water elsewhere in the study area. Sulfate in water discharging from the Spiro Tunnel and chloride from road salt are the primary causes of the increased dissolved-solids concentrations.

Tritium values in ground-water samples indicate that water has infiltrated into the ground-water system before and after atmospheric nuclear testing, which reached its peak during 1962-63. Chlorofluorocarbon analysis of ground-water samples indicates that water infiltrated into the ground-water system from pre-1940 to 1995. Even though ground-water levels rise within a few weeks of snowmelt, water typically takes 15 to 40 years to move through the ground-water system.

A water-budget analysis and computer simulation of snowmelt runoff were used to better understand the hydrologic system in the study area and each subbasin. Water budgets for the entire study area and six subbasins for the 1995 water year were developed. As initial conditions for development of these water budgets, total precipitation or water available must be known. As part of this process, snowmelt runoff was simulated to estimate ground-water recharge to consolidated rock and unconsolidated valley fill with a topographically distributed energy-balance snowmelt model. The model, controlled by independent climate data, favorably simulated the snow water equivalent measured at three SNOTEL sites and the spatial distribution of snowmelt over the study area.

Inflow to the study area is from precipitation in the study area and estimated ground-water inflow across the southern boundary of the study area through consolidated rocks, mine tunnels, or fractures that intersect mine tunnels. Water leaves the study area by evapotranspiration, surface-water and mine-tunnel flow out of the study area, consumptive use of ground water and surface water, and sublimation from snow. Ground water also might flow out of the study area through flowpaths deep within the mountain block. If ground water is leaving the study area, the ground-water budget analysis indicates most of it would be from the East Canyon subbasin. About 50 percent of the water that entered the study area is consumed within the study area.

Ground-water recharge from precipitation made up about 80 percent of the recharge within the study area. Because of the high percentage of recharge from precipitation, ground-water levels and discharge to springs and streams are highly dependent upon precipitation. Because precipitation was much greater than normal for the 1995 water year, the residuals of the total-water budget and the ground-water budget indicate that ground water in storage increased by about 33,000 acre-ft in 1995. Water levels were higher in most of the study area in September 1995 than in September 1994, indicating that water in storage increased during the water year.

The largest component of discharge from the ground-water system is discharge to streams. Ground-water discharge to streams contributes about 40 percent of the surface water in the study area. Ground-water discharge to springs and mine tunnels contributes about 25 percent of the surface water in the study area. Additional use of ground water has the potential to decrease discharge to streams and affect both the amount and quality of surface water in the study area. Much of the water used, however, returns to the surface- or ground-water system.

Because of its high altitude and low consumptive use, the McLeod subbasin is an important part of the hydrologic system for the entire study area. Streamflow originating in the McLeod subbasin made up about 40 percent of the flow leaving the study area in East Canyon Creek and about 20 percent of the flow leaving the study area in Silver Creek. Recharge during 1995 quickly resulted in increased ground-water levels, which increased the gradient toward and discharge from the Spiro Tunnel, two large springs, and streams. Ground-water discharge in the Thaynes Canyon area within the McLeod subbasin exceeded estimated ground-water recharge from precipitation. The additional water needed to balance the ground-water budget in this subbasin probably is supplied by ground-water inflow across the southern boundary of the study area.

Infiltration of water from streams and ground-water discharge to streams are important processes in the Snyderville subbasin. Infiltration of water from streams contributed about 25 percent of the ground-water recharge in the subbasin, and ground-water discharge to streams contributed about 40 percent to the surface-water flow originating in this subbasin. The residuals of the total-water budget and the ground-water budget in the Snyderville subbasin indicate that ground-water storage could have increased or that ground water may flow to other subbasins or out of the study area. Recharge at higher altitudes might not be flowing through the Twin Creek Limestone and the Nugget Sandstone toward Kimball Junction. Ground-water discharge from the Twin Creek Limestone exceeded estimated ground-water recharge from precipitation on the Twin Creek Limestone. Decreased water levels in two observation wells completed in the Twin Creek Limestone indicate that at least some of this difference was water removed from storage near Kimball Junction. Recharge from precipitation to the Nugget Sandstone was sufficient to meet known discharge from the Nugget Sandstone, but decreased water levels in one observation well completed in the Nugget Sandstone near Kimball Junction indicate that water also was removed from storage within this consolidated-rock unit. If ground-water withdrawals near Kimball Junction increase, water will be removed from storage until water levels decline sufficiently to either reduce flow from the Twin Creek Limestone and the Nugget Sandstone to the unconsolidated valley fill or to induce flow from the unconsolidated valley fill to the Twin Creek Limestone and Nugget Sandstone. Surface-water measurements indicate that East Canyon Creek loses water near Kimball Junction. Water that infiltrates to the unconsolidated valley fill may be moving downward to the underlying Twin Creek Limestone. The hydrologic connection and vertical gradient between the unconsolidated valley fill and the Twin Creek Limestone, however, could not be determined because monitoring wells do not exist from which to obtain definitive data.

Because of low altitude and less precipitation than in other areas, about 75 percent of the precipitation in the Silver Creek Junction subbasin is consumed. Only about 10 percent of the precipitation in this subbasin becomes ground-water recharge, but that accounts for 96 percent of the recharge for the subbasin. During years of less-than-normal precipitation, possibly all precipitation is consumed or runs off to streams and that ground-water recharge from precipitation is negligible. Ground-water discharge to streams is the only natural discharge known. Ground-water withdrawal from wells, therefore, either removes water from storage or decreases ground-water discharge to streams.

The residuals in the total-water budget and ground-water budget for the East Canyon subbasin indicate that ground-water storage could have increased or that ground water flowed out of the subbasin during the 1995 water year. The area south of Interstate Highway 80 is extensively faulted, and water may enter those faults and flow into the deeper mountain block ground-water system. The area north of Interstate Highway 80 is underlain mostly by sedimentary rocks that dip north. Ground water in this area may be flowing north out of the study area through consolidated rocks. On the basis of water levels in observation wells, however, ground-water storage probably increased during the 1995 water year. The interaction of ground water and East Canyon Creek appears to be complex and varied. Ground-water recharge from infiltration of streams is insignificant throughout most of the subbasin but may be significant in local areas along East Canyon Creek. Water levels during pumping in many production wells are below the altitude of streams and could induce flow through the unconsolidated valley fill to the consolidated rocks. Ground-water withdrawal might reduce flow from the consolidated rocks to the unconsolidated valley fill, which could reduce water levels in the valley fill and cause decreased ground-water discharge to streams or riparian areas. Because natural discharge before ground-water withdrawal from wells began is unknown, the effects of ground-water withdrawal are difficult to determine. All withdrawal, however, must be met by a change in storage, a reduction in natural discharge, or an increase of infiltration of streams. Monitoring wells near streams could help determine the direction of flow between the ground-water system and streams.

The ground-water budget in the Upper Silver Creek subbasin was balanced by assuming that ground water flowed into the subbasin from south of the study area and from the McLeod subbasin. About 3,000 acre-ft of flow in mine tunnels may be derived from recharge from precipitation south of the study area. In addition, about 1,000 acre-ft of ground-water flows from the McLeod subbasin through unconsolidated valley fill or the Thaynes Formation into this subbasin. The Thaynes Formation possibly contributes water to the overlying unconsolidated valley fill during the spring but may receive water from it during late summer and fall. The hydraulic gradient is downward in the unconsolidated valley fill near Prospector Square and water flows downward from the unconsolidated valley fill into the Thaynes Formation during short-term pumping. The downward gradient may reduce discharge to streams. Water-quality data indicate that water with higher sulfate, chloride, and dissolved-solids concentrations flows downward into the Thaynes Formation.

Recharge from infiltration of irrigation water made up about 50 percent of the recharge in the Lower Silver Creek subbasin, mainly because the flow in Silver Creek is diverted to a canal along the east side of the subbasin where much of the water infiltrates into the subsurface. Recharge from precipitation to the Twin Creek Limestone was less than discharge from the Twin Creek Limestone, indicating that additional ground water moves into the Twin Creek Limestone or that water is removed from storage. The only significant natural discharge is to streams; therefore, ground-water withdrawal from wells will be balanced by a reduction in ground-water storage until water levels decline sufficiently to reduce ground-water discharge to streams or induce ground-water recharge from streams.

A comparison of the 1995 to 1994 water budgets emphasizes that the hydrologic system in the study area is very dependent on the amount of annual precipitation and has low capacity for ground-water storage. Although precipitation on the study area was much greater in 1995 than in 1994, most of the additional water resulted in increased discharge to springs and streams rather than increased storage in the ground-water system. Water-level fluctuations in monitoring wells also indicate that much of the water caused increased discharge and did not remain in storage.

Ground-water levels and ground-water discharge are dependent upon annual precipitation and differ substantially from year to year. Water-level fluctuations and variation in discharge to springs and streams indicate that, in general, the storage coefficient is small throughout the study area. Discharge to streams is derived primarily from the unconsolidated valley fill. The large increase in discharge, therefore, indicates that the unconsolidated valley fill in the study area does not provide significant ground-water storage.

Snowmelt runoff was simulated with an energy-balance snowmelt model to estimate ground-water recharge to consolidated-rock and unconsolidated valley-fill aquifers in the study area. The simulated snow water equivalent compared favorably with the daily snow water equivalent at the SNOTEL sites indicating that the model was reasonably simulating the snow water equivalent of the snowpack and snowmelt runoff. In the McLeod subbasin for March to June of 1994 and 1995, about 70 percent of the snowmelt runoff and spring rainfall recharged the ground-water system.

The effects of increased surface-water use and ground-water withdrawals can best be determined by continuing data collection at long-term stream-gaging stations. Additional stream gages could be established to help define surface runoff and infiltration into different consolidated-rock formations. Snowmelt data could be collected for additional snowmelt simulations to help define recharge to specific consolidated-rock units. Additional monitoring wells installed with known depths and completions could help determine water-level fluctuations in specific consolidated-rock units and unconsolidated valley fill.

REFERENCES CITED

American Society of Civil Engineers, 1989, Water use by naturally occurring vegetation including an annotated bibliography: New York, 1989, 245 p.

Anderson, E.A., 1976, A point energy and mass balance model of a snow cover: NWS Technical Report 19, National Oceanic and Atmospheric Administration, Washington, DC, 150 p.

Ashland, F.X., Bishop, C.E., Lowe, Mike, and Mayes, B.H., 1996, The geology of the Snyderville Basin and its relation to ground-water conditions: Utah Geological Survey Open-File Report 337, 124 p.

Baker, C.H., Jr., 1970, Water resources of the Heber-Kamas-Park City area, north-central Utah: Utah Department of Natural Resources Technical Publication No. 27, 79 p.

Bodine, M.W., Jr., and Jones, B.F., 1986, The Salt Norm: A quantitative chemical-mineralogical characterization of natural waters: U.S. Geological Survey Water-Resources Investigations Report 86-4086, 130 p.

Boutwell, J.M., 1912, Geology and ore deposits of the Park City district, Utah: U.S. Geological Survey Professional Paper 77, 231 p.

Branson, F.A., Miller, R.F., and McQueen, I.S., 1970, Plant communities and associated soil and water factors on shale-derived soils in northeastern Montana: Ecology, v. 51, no. 3, p. 391-407.

Bromfield, C.S., 1968, General geology of the Park City region, Utah, *in* Guidebook to the geology of Utah, Park City District, Utah: Utah Geological Society Guidebook No. 22, 102 p.

Bromfield, C.S., and Crittenden, M.D., Jr., 1971, Geologic map of the Park City east quadrangle, Utah: U.S. Geological Survey Quadrangle Map GQ-832, Scale 1:24,000.

Brown, H.E., and Thompson, J.R., 1965, Summer water use by aspen, spruce, and grassland in western Colorado: Journal of Forestry, v. 63, no. 10, p. 756-760.

Bryant, Bruce, 1990, Geologic map of the Salt Lake City 30' x 60' quadrangle, north-central Utah, and Uinta County, Wyoming: U.S. Geological Survey Miscellaneous Investigation Series, Map I-1944, 2 sheets.

Cooper, L.W., Olsen, C.R., Solomon, D.K., Larsen, I.L., Cook, R.B., and Grebmeier, J.M., 1991, Stable isotopes of oxygen and natural and fallout radionuclides used for tracing runoff during snowmelt in an Arctic watershed: Water Resources Research, v. 27, no. 9, p. 2171-2179.

Craig, H., 1961, Isotopic variations in meteoric waters: Science, v. 133, p. 1702-1703.

Crittenden, M.D., Jr., Calkins, F.C., and Sharp, B.J., 1966, Geologic map of the Park City west quadrangle, Utah: U.S. Geological Survey Quadrangle Map GQ-535, Scale 1:24,000.

Croft, A.R., and Monninger, L.V., 1953, Evapotranspiration and other water losses on some aspen forest types in relation to water available for streamflow: Transactions, American Geophysical Union, v. 34, no. 4, p. 563-574.

Domenico, P.A., and Schwartz, F.W., 1990, Physical and chemical hydrogeology: New York, John Wiley and Sons, 824 p.

Downhour, P.A., and Brooks, L.E., 1996, Selected hydrologic data for Snyderville Basin, Park City, and adjacent areas, Summit County, Utah, 1967-95: U.S. Geological Survey Open-File Report 96-494, 52 p.

Dozier, J., 1980, A clear-sky spectral solar radiation model for snow-covered mountainous terrain: Water Resources Research, v. 16, p. 709-718.

Drever, J.I., 1988, The geochemistry of natural waters: Englewood Cliffs, N.J., Prentice-Hall, 437 p.

Dubayah, R., Dozier, J., and Davis, F., 1990, Topographic distribution of clear-sky radiation over the Konza Prairie, Kansas: Water Resources Research, v. 26, p. 679-691.

Faure, Gunter, 1977, Principles of isotope geology: New York, John Wiley and Sons, 464 p.

Fenneman, N.M., 1931, Physiography of the Western United States: New York, McGraw-Hill, 534 p.

Fetter, C.W., Jr., 1980, Applied hydrogeology: Columbus, Ohio, Charles E. Merrill Publishing Co., 488 p.

Forster, C., and Smith, L., 1988, Groundwater flow systems in mountainous terrain—1. Numerical modeling technique: Water Resources Research, v. 24, no. 7, p. 999-1010.

Garen, D.C., 1995, Estimation of spatially distributed values of daily precipitation in mountainous areas, *in* Mountain hydrology—Peaks and valleys in research and applications: Proceedings of Canadian Water Resources Association Conference, Vancouver, British Columbia, p. 237-242.

Garen, D.C., and Marks, D., 1996, Spatially distributed snow modelling in mountainous regions—Boise River application, *in* Application of geographic information systems in hydrology and water resources management: Proceedings of HydroGIS 96 conference, Vienna, Austria, IAHS Publication No. 235, p. 421-428.

Garen, D.C., Johnson, G.L., and Hanson, C.L., 1994, Mean areal precipitation for daily hydrologic modeling in mountainous areas: Water Resources Bulletin v. 30, no. 3, p. 481-491.

Holmes, W.F., Thompson, K.R., and Enright, M., 1986, Water resources of the Park City area, Utah, with emphasis on ground water: Utah Department of Natural Resources Technical Publication No. 85, 81 p.

Hungerford, R.D., Nemani, R.R., Running, S.W., and Coughlan, J.C., 1989, MTCLIM: A mountain microclimate simulation model: Research Paper INT-414, U.S. Department of Agriculture, Forest Service, Intermountain Research Station, Odgen, Utah.

Imlay, R.W., 1967, Twin Creek Limestone (Jurassic) in the western interior of the United States: U.S. Geological Survey Professional Paper 540, 105 p.

James M. Montgomery Engineers, 1990, Observations and results of the Middle School Well pump test: Salt Lake City, Utah, James M. Montgomery Consulting Engineers, Inc., 49 p.

Jordan, R., 1991, Special report 91-16—A one-dimensional temperature model for a snow cover—Technical documentation for SNTHERM.89: Hanover, N.H., U.S. Army Corps of Engineers Cold Regions Research and Engineering Laboratory, 49 p.

Kaufmann, M.R., 1984, A canopy model (RM-CWU) for determining transpiration of subalpine forests—II. Consumptive water use in two watersheds: Canadian Journal of Forest Research, v. 14, p. 227-232.

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

Marks, D., 1988, Climate, energy exchange, and snowmelt in Emerald Lake watershed, Sierra Nevada: Santa Barbara, University of California, Ph.D. dissertation, 158 p.

Marks, D., and Dozier, J., 1979, A clear-sky longwave radiation model for remote alpine areas: Archiv fu"r_ Meteorologie, Geophysik und Bioklimatologie, Series B, 27, p. 159-187.

Marks, D., and Dozier, J., 1992, Climate and energy exchange at the snow surface in the alpine region of the Sierra Nevada—2. Snow cover energy balance: Water Resources Research, v. 28, p. 3043-3054.

Marks, D., Dubayah, R., and Longley, R., 1991, Modeling the topographic and spectral variability of clear-sky solar radiation at regional to continental scales, *in* Remote sensing: Global Monitoring for Earth Management, Proceedings, 1991 IGARSS Symposium, Espoo, Finland, v. 3, p. 1711.

Mason, J.L., 1989, Hydrology of the Prospector Square area, Summit County, Utah: U.S. Geological Survey Water-Resources Investigation Report 88-4156, 75 p.

Mayo, A.L., Nielsen, P.J., Loucks, M., and Brimhall, W.H., 1992, The use of solute and isotopic chemistry to identify flow patterns and factors which limit acid mine drainage in the Wasatch Range, Utah: Ground Water, v. 30, no. 2, p. 243-249.

Morris, E.M., 1982, Sensitivity of the European Hydrological System snow models, *in* J.W. Glen, ed., Hydrological aspects of alpine and high-mountain areas: IAHS-AIHS Publication 138, International Association of Hydrological Sciences, Wallingford, UK, p. 221-231.

Morris, E.M., 1986, Modelling of a seasonal snowcover, *in* G. Kukla, R. G. Barry, A. Hecht, and D. Wiesnet (eds.) Snow Watch '85, Glaciological Data report GD-18, World Data Center for Glaciology, Boulder, Colo., p. 225-240.

Nielsen, P.J., and Mayo, A.L., 1989, Chemical and isotopic investigation of the cause of acid and neutral mine discharges in the central Wasatch Range, Utah, *in* Geology and hydrology of hazardous-waste, mining-waste, waste-water, and repository sites in Utah: Utah Geological Association Publication 17, p. 121-134.

Ott, R.L., 1993, An introduction to statistical methods and data analysis (4th ed.): Belmont, California, Duxbury Press, 1183 p.

Plummer, L.N., Jones, B.F., and Truesdell, A.H., 1984, WATEQF — A Fortran IV version of WATEQ, a computer program for calculating chemical equilibrium of natural waters: U.S. Geological Survey Water-Resources Investigations 76-13, 70 p.

Plummer, L.N., Michel, R.L., Thurman, E.M., and Glynn, P.D., 1993, Environmental tracers for age-dating young ground water, *in* Alley, W.M., ed., Regional Ground-Water Quality: New York, Van Nostrand Reinhold, p. 255-294.

Plummer, L.N., Prestemon, E.C., and Parkhurst, D.L., 1994, An interactive code (NETPATH) for modeling net geochemical reactions along a flow path, version 2.0: U.S. Geological Survey Water-Resources Investigations Report 94-4169, 130 p.

Smart, A.W., and Fleming, W.M., 1985, Consumptive water use in artificial snowmaking, Santa Fe Ski Area, New Mexico: New Mexico State Engineer Office Technical Report No. 45, June 1985, 14 p.

Theis, 1940, The source of water derived from wells: Civil Engineering, v. 10, no. 5, p. 277-280.

Thompson, K.R., 1983, Reconnaissance of the quality of surface water in the Weber River basin, Utah: Utah Department of Natural Resources Technical Publication No. 76, 74 p.

Tomlinson, S.A., 1996a, Evaluating evapotranspiration for six sites in Benton, Spokane, and Yakima Counties, Washington, May 1990 to September 1992: U.S. Geological Survey Water-Resources Investigations Report 96-4002, 84 p.

Tomlinson, S.A., 1996b, Comparison of Bowen-Ratio, Eddy-Correlation, and weighing-lysimeter evapotranspiration for two sparse-canopy sites in eastern

Washington: U.S. Geological Survey Water-Resources Investigations Report 96-4081, 69 p.

Utah Climate Center, 1996, 1961-90 normal precipitation contours: Logan, Utah, Utah State University.

Utah Cooperative Fish and Wildlife Research Unit, 1995, Utah Gap analysis, an environmental information system: Final Project Report 95-1, Logan, Utah, Utah State University, 46 p.

Utah Department of Natural Resources, Division of Water Rights, 1982, Water-use data for public water suppliers in Utah 1980: Utah Department of Natural Resources, Water Use Report No. 3, 94 p.

Utah Department of Natural Resources, Division of Water Rights, 1993, Water-use data for public water suppliers and self-supplied industry in Utah 1990-91: Utah Department of Natural Resources Water Use Report No. 9, 23 p.

Utah Department of Natural Resources, Division of Water Resources, 1992, A water-related land use inventory of the Weber River Basin, 51 p.

Utah Department of Natural Resources, Division of Water Resources, 1996, Water budget report of the Weber River Basin, Draft, 104 p.

Utah State University, 1994, Consumptive use of irrigated crops in Utah: Utah Agricultural Experiment Station Research Report 145, 361 p.

van Heeswijk, M., Kimball, J.S., and Marks, D., 1995, Simulation of water available for runoff in clearcut forest opening during rain-on-snow events in the western Cascade Range of Oregon and Washington: U.S. Geological Survey Water-Resources Investigations Report 95-4219, 67 p.

Weber River Water Commissioner, 1995, Weber River distribution system annual report 1994, 161 p.

Weber River Water Commissioner, 1996, Weber River distribution system annual report 1995, 161 p.

Wight, J.R., Hanson, C.L., and Cooley, K.R., 1986, Modeling evapotranspiration from sagebrush-grass rangeland: Journal of Range Management, v. 39, no. 1, p. 81-85.

Winter, T.C., 1981, Uncertainties in estimating the water balance of lakes: Water Resources Bulletin, v. 17, no. 1, p. 82-115.

Withington, C.F., 1964, Gypsum and anhydrite, *in* Mineral and Water Resources of Utah: Utah Geological and Mineral Survey Bulletin 73, 275 p.

Wright Water Engineers, Inc., 1986, The Colorado Ski Country USA water management research project—Summary of study results: Denver, Colorado, Wright Water Engineers, Inc., 18 p.

USGS
science for a changing world

TAKE Pride IN UTAH

The Utah Department of Natural Resources receives federal aid and prohibits discrimination on the basis of race, color, sex, age, national origin or disability. For information or complaints regarding discrimination, contact the Executive Director, Utah Department of Natural Resources, 1636 West North Temple #316, Salt Lake City, UT 84116-3193 or Office of Equal Opportunity, US Department of the Interior, Washington, DC 20240.

Printed with vegetable oil ink. 500 6/98

☆ U.S. GOVERNMENT PRINTING OFFICE: 1998—673-081 / 30012 Region No. 8