U.S. Department of the Interior
U.S. Geological Survey

Ground-Water Hydrology of the Upper Deschutes Basin, Oregon

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Water-Resources Investigations Report 00–4162

Prepared in cooperation with Oregon Water Resources Department;
cities of Bend, Redmond, and Sisters; Deschutes and Jefferson Counties;
The Confederated Tribes of the Warm Springs Reservation of Oregon;
and U.S. Environmental Protection Agency

Portland, Oregon: 2001

U. S. DEPARTMENT OF THE INTERIOR
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Suggested citation:

Gannett, M.W., Lite, K.E., Jr., Morgan, D.S., and Collins, C.A., 2001, Ground-water hydrology of the upper Deschutes Basin, Oregon: U.S. Geological Survey Water-Resources Investigations Report 00–4162, 78 p.

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
acre	4,047	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
acre-foot (acre-ft)	1,233	cubic meter (m^3)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
inches per year (in./yr)	0.0254	meters per year (m/yr)
feet per day (ft/d)	3.528×10^{-6}	meters per second (m/s)
gallon per minute (gal/min)	6.308×10^{-5}	cubic meters per second (m^3/s)
square feet per day (ft^2/d)	1.075×10^{-5}	square meters per second (m^2/s)
feet per year (ft/yr)	9.659×10^{-9}	meters per second (m/s)
acre-feet per year (acre-ft/yr)	3.909×10^{-5}	cubic meters per second (m^3/s)
cubic feet per day per square foot ($ft^3/d/ft^2$)	3.528×10^{-6}	cubic meters per second per square meter ($m^3/s/m^2$)
gallons per day (gal/d)	4.381×10^{-8}	cubic meters per second (m^3/s)
feet per second (ft/s)	0.3048	meter per second (m/s)

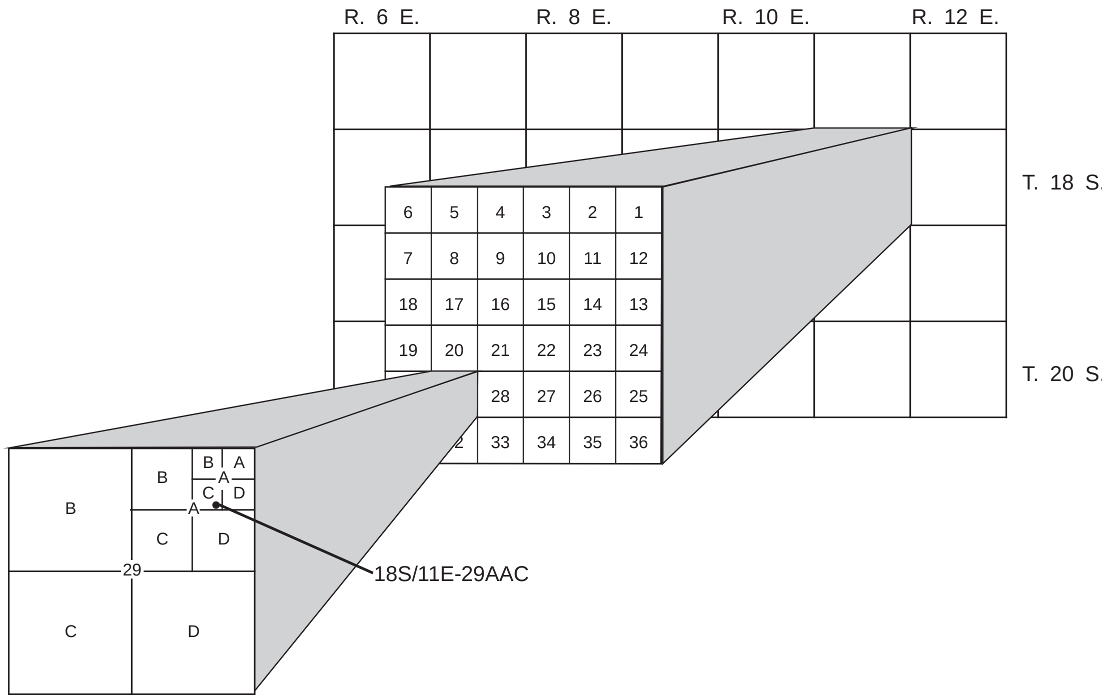
Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:

°F=1.8 °C+32

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

LOCATION SYSTEM

The system used for locating wells, springs, and surface-water sites in this report is based on the rectangular system for subdivision of public land. The State of Oregon is divided into townships of 36 square miles numbered according to their location relative to the east-west Willamette baseline and a north-south Willamette meridian. The position of a township is given by its north-south “Township” position relative to the baseline and its east-west “Range” position relative to the meridian. Each township is divided into 36 one-square-mile (640-acre) sections numbered from 1 to 36. For example, a well designated as 18S/11E-29AAC is located in Township 18 south, Range 11 east, section 29. The letters following the section number correspond to the location within the section; the first letter (A) identifies the quarter section (160 acres); the second letter (A) identifies the quarter-quarter section (40 acres); and the third letter (C) identifies the quarter-quarter-quarter section (10 acres). Therefore, well 29AAC is located in the SW quarter of the NE quarter of the NE quarter of section 29. When more than one designated well occurs in the quarter-quarter-quarter section, a serial number is included.



Well- and spring-location system.

Each well is assigned a unique 8-digit identification number known as the log-id number. The first two digits of the log-id number indicate the county code from the Federal Information Processing Standards (FIPS) code file for the county in which the well exists. The FIPS codes for the counties in the study area are as follows: 13, Crook County; 17, Deschutes County; 31, Jefferson County; and 35, Klamath County. The last 6 digits of the number correspond to the State of Oregon well-log number (a unique number assigned by the Oregon Water Resources Department to the report filed by the well driller).

MAPPING SOURCES:

Base map modified from U.S. Geological Survey 1:500,000 State base map, 1982, with digital data from U.S. Bureau of the Census, TIGER/Line (R), 1990, and U.S. Geological Survey Digital Line Graphs published at 1:100,000.

Publication projection is Lambert Conformal Conic.

Standard parallels 43°00’ and 45°30’, central meridian –120°30’.

Ground-Water Hydrology of the Upper Deschutes Basin, Oregon

By Marshall W. Gannett, Kenneth E. Lite Jr., David S. Morgan, and Charles A. Collins

Abstract

The upper Deschutes Basin is among the fastest growing regions in Oregon. The rapid population growth has been accompanied by increased demand for water. Surface streams, however, have been administratively closed to additional appropriation for many years, and surface water is not generally available to support new development. Consequently, ground water is being relied upon to satisfy the growth in water demand. Oregon water law requires that the potential effects of ground-water development on streamflow be evaluated when considering applications for new ground-water rights. Prior to this study, hydrologic understanding has been insufficient to quantitatively evaluate the connection between ground water and streamflow, and the behavior of the regional ground-water flow system in general. This report describes the results of a hydrologic investigation undertaken to provide that understanding. The investigation encompasses about 4,500 square miles of the upper Deschutes River drainage basin.

A large proportion of the precipitation in the upper Deschutes Basin falls in the Cascade Range, making it the principal ground-water recharge area for the basin. Water-balance calculations indicate that the average annual rate of ground-water recharge from precipitation (1993–95) is about 3,500 ft^3/s (cubic feet per second). Water-budget calculations indicate that in addition to recharge from precipitation, water enters the ground-water system through interbasin flow. Approximately 800 ft^3/s flows into the Metolius River drainage from the west and about 50 ft^3/s flows into the southeastern part of the study area from the Fort Rock Basin. East of the Cascade Range, there is little or no ground-water recharge from precipitation, but leaking irrigation canals are a significant source of artificial recharge north of Bend. The average annual rate of canal leakage during 1994 was estimated to be about 490 ft^3/s. Ground water flows from the Cascade Range through permeable volcanic rocks eastward out into the basin and then generally northward. About one-half the ground water flowing from the Cascade Range discharges to spring-fed streams along the margins of the range, including the upper Metolius River and its tributaries. The remaining ground water flows through the subsurface, primarily through rocks of the Deschutes Formation, and eventually discharges to streams near the confluence of the Deschutes, Crooked, and Metolius Rivers. Substantial ground-water discharge occurs along the lower 2 miles of Squaw Creek, the Deschutes River between Lower Bridge and Pelton Dam, the lower Crooked River between Osborne Canyon and the mouth, and in Lake Billy Chinook (a reservoir that inundates the confluence of the Deschutes, Crooked, and Metolius Rivers).

The large amount of ground-water discharge in the confluence area is primarily caused by geologic factors. North (downstream) of the confluence area, the upper Deschutes Basin is transected by a broad region of low-permeability rock of the John Day Formation. The Deschutes River flows north across the low-permeability region, but the permeable Deschutes Formation, through which most of the regional ground water flows, ends against this rampart of low-permeability rock. The northward-flowing ground water discharges to the streams in this area because the permeable strata through which it flows terminate, forcing the water to discharge to the surface.

Virtually all of the regional ground water in the upper Deschutes Basin discharges to surface streams south of the area where the Deschutes River enters this low-permeability terrane, at roughly the location of Pelton Dam.

The effects of ground-water withdrawal on streamflow cannot presently be measured because of measurement error and the large amount of natural variability in ground-water discharge. The summer streamflow near Madras, which is made up largely of ground-water discharge, is approximately 4,000 ft^3/s. Estimated consumptive ground-water use in the basin is about 30 ft^3/s, which is well within the range of the expected streamflow measurement error. The natural variation in ground-water discharge upstream of Madras due to climate cycles is on the order of 1,000 ft^3/s. This amount of natural variation masks the effects of present ground-water use. Even though the effects of ground-water use on streamflow cannot be measured, geologic and hydrologic analysis indicate that they are present.

Ground-water-level fluctuations in the upper Deschutes Basin are driven primarily by decadal climate cycles. Decadal water-level fluctuations exceeding 20 ft (feet) have been observed in wells at widespread locations near the margin of the Cascade Range. The magnitude of these fluctuations diminishes toward the east, with increasing distance from the Cascade Range. Annual water-level fluctuations of a few feet are common in areas of leaking irrigation canals, with larger fluctuations observed in some wells very close to canals. Annual water-level fluctuations of up to 3 ft due to ground-water pumping were observed locally. No long-term water-level declines attributable to pumping were found in the upper Deschutes Basin.

The effects of stresses to the ground-water system are diffused and attenuated with distance. This phenomenon is shown by the regional response to the end of a prolonged drought and the shift to wetter-than-normal conditions starting in 1996. Ground-water levels in the Cascade Range, the locus of ground-water recharge, stopped declining and started rising during the winter of 1996. In contrast, water levels in the Redmond area, 30 miles east of the Cascade Range, did not start to rise again until late 1997 or 1998. The full effects of stresses to the ground-water system, including pumping, may take several years to propagate across the basin.

Ground-water discharge fluctuations were analyzed using stream-gage records. Ground-water discharge from springs and seeps estimated from stream-gage records shows climate-driven decadal fluctuations following the same pattern as the water-level fluctuations. Data from 1962 to 1997 show decadal-scale variations of 22 to 74 percent in ground-water discharge along major streams that have more than 100 ft^3/s of ground-water inflow.

INTRODUCTION

Background and Study Objectives

The upper Deschutes Basin is presently one of the fastest growing population centers in the State of Oregon. The number of people in Deschutes County, the most populous county in the basin, more than tripled between 1970 and 1998 (State of Oregon, 1999). Approximately 140,000 people lived in the upper Deschutes Basin as of 1998. Growth in the region is expected to continue, and residents and government agencies are concerned about water supplies for the burgeoning population and the consequences of increased development for existing water users. Surface-water resources in the area have been closed by the State of Oregon to additional appropriation for many years. Therefore, virtually all new development in the region must rely on ground water as a source of water. Prior to this study, very little quantitative information was available on the ground-water hydrology of the basin. This lack of information made ground-water resource management decisions difficult and was generally a cause for concern.

To fill this information void, the U.S. Geological Survey (USGS) began a cooperative study in 1993 with the Oregon Water Resources Department (OWRD), the cities of Bend, Redmond, and Sisters, Deschutes and Jefferson Counties, The Confederated Tribes of the Warm Springs Reservation of Oregon, and the U.S. Environmental Protection Agency.

The objectives of this study were to provide a quantitative assessment of the regional ground-water system and provide the understanding and analytical tools for State and local government agencies, hydrologists, and local residents to make resource management decisions. This report is one in a series that presents the results of the upper Deschutes Basin ground-water study.

Purpose and Scope

The purpose of this report is to provide a comprehensive quantitative description of regional ground-water flow in the upper Deschutes Basin. The report provides an analysis of the data compiled or collected during the study, and presents a description of the regional ground-water hydrology based on that analysis.

The results of the study presented herein are based on both preexisting information and new data. Preexisting information included regional-scale maps of geology, topography, soils, vegetation, and precipitation. In addition, streamflow data were available for numerous sites for periods of time since the early 1900s. Data were also available from several weather stations that operate in the study area. In addition, surface-water diversion records were available for all major irrigation canals. Data described above were augmented by data from numerous reports and studies. Hydrologic data collected for this study included gain/loss measurements for several streams, and geologic and hydraulic-head data from about 1,500 wells that were precisely located in the field. Geophysical, lithologic, and hydrographic data were collected from a subset of these wells. Wells are unevenly distributed in the area and occur mostly in areas of privately owned land. There are few well data from the large tracts of public land that cover most of the study area. Therefore, there are large regions of the Cascade Range, Newberry Volcano, and the High Lava Plains where subsurface hydrologic information is sparse.

This study is regional in scope. It is intended to provide the most complete assessment possible of the regional ground-water hydrology of the upper Deschutes Basin given the data that were available or that could be collected within the resources of the project. This work is not intended to describe details of ground-water flow at local scales; however, it will provide a sound framework for local-scale investigations.

Study Area

The upper Deschutes Basin study area encompasses approximately 4,500 mi^2 (square miles) of the Deschutes River drainage basin in central Oregon (fig. 1). The area is drained by the Deschutes River and its major tributaries: the Little Deschutes River, Tumalo Creek, Squaw Creek, and the Metolius River from the west, and the Crooked River from the east. Land-surface elevation ranges from less than 1,300 ft near Gateway in the northern part of the study area to more than 10,000 ft above sea level in the Cascade Range.

The study-area boundaries were chosen to coincide as much as possible with natural hydrologic boundaries across which ground-water flow can be reasonably estimated or assumed to be negligible. The study area is bounded on the north by Jefferson Creek, the Metolius River, the Deschutes River, and Trout Creek; on the east by the generalized contact between the Deschutes Formation and the older, much less permeable John Day Formation; on the south by the drainage divides between the Deschutes Basin and the Fort Rock and Klamath Basins; and on the west by the Cascade Range crest.

The study area includes the major population centers in the basin, where ground-water development is most intense and resource management questions are most urgent. The major communities include Bend, Redmond, Sisters, Madras, Prineville, and La Pine. Principal industries in the region are agriculture, forest products, tourism, and service industries.

Sixty-six percent of the 4,500 mi^2 upper Deschutes Basin is publicly owned (fig. 2). Approximately 2,230 mi^2 are under the jurisdiction of the U.S. Forest Service, 730 mi^2 are under the jurisdiction of the Bureau of Land Management, and about 20 mi^2 are under the stewardship of State or County agencies. The remaining 1,520 mi^2 are in private ownership.

The highest elevations in the upper Deschutes Basin are in the western and southern parts. These regions are covered by coniferous forests, most of which have been managed for timber production. The remaining parts of the basin, which are at lower elevations, are more arid and, where not cultivated, are dominated by grassland, sagebrush, and juniper. Most of the non-forest-related agriculture occurs in the central and northern parts of the upper Deschutes Basin.

There are approximately 164,000 acres ($256\ mi^2$) of irrigated agricultural land in the study area. The largest source of irrigation water is the Deschutes River. Most water is diverted from the Deschutes River near Bend and distributed to areas to the north through several hundred miles of canals. Smaller amounts of irrigation water are diverted from Tumalo and Squaw Creeks, the Crooked River, and Ochoco Creek.

The climate in the Deschutes Basin is controlled primarily by air masses that move eastward from the Pacific Ocean, across western Oregon, and into central Oregon. The climate is moderate with cool, wet winters and warm, dry summers. Orographic processes result in large amounts of precipitation in the Cascade Range in the western part of the basin, with precipitation locally exceeding 200 in./yr (inches per year), mostly as snow, during the winter (Taylor, 1993). Precipitation rates diminish rapidly toward the east to less than 10 in./yr in the central part of the basin (fig. 3). Temperatures also vary across the basin. Records from the Oregon Climate Service show mean daily minimum and maximum temperatures at Santiam Pass in the Cascade Range (period of record 1961–85) range from 21 and 34°F (degrees Fahrenheit) in January to 43 and 73°F in July (Oregon Climate Service, 1999). Conditions are warmer at lower elevations in the central part of the basin. The mean daily minimum and maximum temperatures in Bend (period of record 1961 to 1999) range from 22 and 42°F in January to 45 and 81°F in July (Oregon Climate Service, 1999). Climate in the Deschutes Basin exhibits year-to-year and longer-term variability. This variability generally parallels regional trends in the Pacific Northwest that have been correlated with large-scale ocean-atmosphere climate variability patterns in the Pacific Basin such as the El Niño/Southern Oscillation (Redmond and Koch, 1991) and the Pacific Decadal Oscillation (Mantua and others, 1997).

Approach

The approach to this study consisted of five major elements: (1) reviewing existing geologic and hydrologic maps and literature and conceptual models of the regional flow system, (2) inventorying and field-locating wells for subsurface geological and hydraulic-head information, (3) compiling and collecting data to estimate the amounts and distribution of various components of the hydrologic budget, (4) compiling and collecting water-level fluctuation information to evaluate the dynamics of regional ground-water flow and assess the state of the system, and (5) developing a computer model to simulate the ground-water flow system. This report addresses the first four of these elements.

At the onset of this investigation there were no published reports on the quantitative regional ground-water hydrology of the basin. The only regional-scale reports prior to this study were an unpublished descriptive report written for the Oregon State Engineer (Sceva, 1960) and an assessment of the potential effects of disposal wells in the basin (Sceva, 1968). All other ground-water reports and studies were restricted to smaller geographic areas. Sceva's works presented a conceptual model of regional ground-water flow in the basin that has been largely corroborated by this study. Although no single geologic map encompassed the entire study area at a scale larger than 1:500,000, the study area was largely covered by a montage of maps at scales ranging from 1:100,000 to 1:24,000.

This study benefited from the inventory and field location of about 700 wells by the USGS in the late 1970s as part of a study that was later terminated for lack of funding. In addition, geophysical logs and periodic water-level measurements existed for a subset of those wells. To augment the 700 wells field located at the start of this investigation, an additional 800 wells were inventoried and field located. The geographic distribution of these 1,500 field-located wells (fig. 2) mirrors the distribution of wells in the basin in general. The highest density of wells occurs on private land. Water levels were measured in located wells whenever possible. Field-located wells provided information on hydraulic-head distribution and subsurface geology. Approximately 35 wells were geophysically logged and drill cuttings were collected for approximately 70 wells. One-hour specific-capacity tests were available for most wells and aquifer tests were conducted on four wells to provide additional information on hydraulic characteristics.

Water-level data from field-located wells and elevations of major springs and gaining streams were used to map hydraulic-head distribution in the region. The resulting distribution map was the basic source of information regarding the horizontal and vertical directions of ground-water flow.

Major components of the hydrologic budget were either measured or estimated. Recharge from natural precipitation was estimated by a daily mass-balance approach using the Deep Percolation Model (DPM) of Bauer and Vaccaro (1987). Recharge from canal leakage was estimated from surface-water diversion records and estimates of farm deliveries, in combination with canal seepage studies conducted by the Bureau of Reclamation (BOR). Farm deliveries and on-farm losses were derived from consumptive-use and irrigation-efficiency estimates. On-farm consumptive use was estimated from crop information derived from LANDSAT images and crop-water-use estimates from BOR AgriMet stations in the basin.

The rate and distribution of ground-water discharge to streams and springs throughout the study area were estimated using data from active and historic stream gages, gain/loss studies conducted by OWRD Central Region staff, and miscellaneous published streamflow measurements. The rate and distribution of ground-water pumping was estimated for public supply and for irrigation uses. Public-supply pumping was derived from measurements or estimates supplied by the municipalities and other public water suppliers. Irrigation pumping was estimated using information from the OWRD Water-Rights Information System (WRIS) in combination with on-farm consumptive-use estimates derived in the manner described above. Pumping by private domestic wells was estimated using well-log records and population statistics.

The dynamics of the ground-water flow system, both at a regional and local scale, were evaluated by analyzing ground-water-level fluctuations in response to both long- and short-term hydrologic phenomena such as variations in climate, individual storms, canal operation, and pumping. Periodic water-level measurements were compiled from historic data and collected from about 100 wells. The frequency of measurements and the duration of records for wells varied considerably. There were about 90 wells with quarterly water-level measurements spanning periods ranging from a few years to over 50 years. In addition, there are 16 wells in which water levels were recorded every 2 hours for periods ranging from a few months to over 4 years (Caldwell and Truini, 1997).

The chemistry of selected wells, springs, and canals in the study area was analyzed and interpreted by Caldwell (1998). This analysis provided additional insights into the regional ground-water flow system and into the interaction of ground water and surface water, including irrigation canals.

Acknowledgments

The authors gratefully acknowledge the support of the residents of the upper Deschutes Basin throughout this investigation. Particular thanks go to the hundreds of landowners who allowed access to their wells for water-level monitoring and sampling. The Public Works staff of the cities of Bend, Redmond, Sisters, and Madras were extremely helpful in providing access to their water systems as well as information on water use and disposal wells. Private water companies in the basin were also helpful, but particular thanks goes to Avion Water Company, Black Butte Ranch Corporation, Deschutes Valley Water District, and Juniper Utility for access to their wells for monitoring, testing, and sampling. Information on diversions and water use provided by the many irrigation districts in the basin was extremely helpful and is greatly appreciated. There are several individuals who have particular knowledge of certain aspects of geology or water in the basin, and who freely shared ideas and insights with the authors. This group includes (alphabetically) Larry Chitwood, Rick Conrey, Mark Ferns, Kyle Gorman, Bob Main, Dave Sherrod, and Jan Wick. Larry Chitwood provided a thoughtful and thorough review of this report. Lastly, special thanks go to Susan Prowell with the city of Bend, who handled all the logistics for years of quarterly meetings, and to all the people who showed their interest and support by attending them.

GEOLOGIC FRAMEWORK

The storage and flow of ground water are controlled to a large extent by geology. The principle geologic factors that influence ground water are the *porosity* and *permeability* of the rock or sediment through which it flows. Porosity, in general terms, is the proportion of a rock or deposit that consists of open space. In a gravel deposit, this would be the proportion of the volume of the deposit represented by the space between the individual pebbles and cobbles. Permeability is a measure of the resistance to the movement of water through the rock or deposit. Deposits with large interconnected open spaces, such as gravel, have little resistance to ground-water flow and are therefore considered highly permeable. Rocks with few, very small, or poorly connected open spaces offer considerable resistance to ground-water flow and, therefore, have low permeability. The hydraulic characteristics of geologic materials vary between rock types and within particular rock types. For example, in sedimentary deposits the permeability is a function of grain size and the range of grain sizes (the degree of sorting). Coarse, well-sorted gravel has much higher permeability than fine, silty sand deposits. The permeability of lava flows can also vary markedly depending on the degree of fracturing. The highly fractured, rubbly zones at the tops and bottoms of lava flows and in interflow zones are often highly permeable, while the dense interior parts of lava flows can have very low permeability. Weathering and secondary mineralization, which are often a function of the age of the rock, can strongly influence permeability. Sedimentary deposits or lava flows in which the original open spaces have been infilled with secondary minerals can have very low permeability.

Geologic properties that influence the movement of ground water within a flow system can also define the boundaries of the system. Terranes consisting of predominantly low-permeability materials can form the boundaries of a regional flow system.

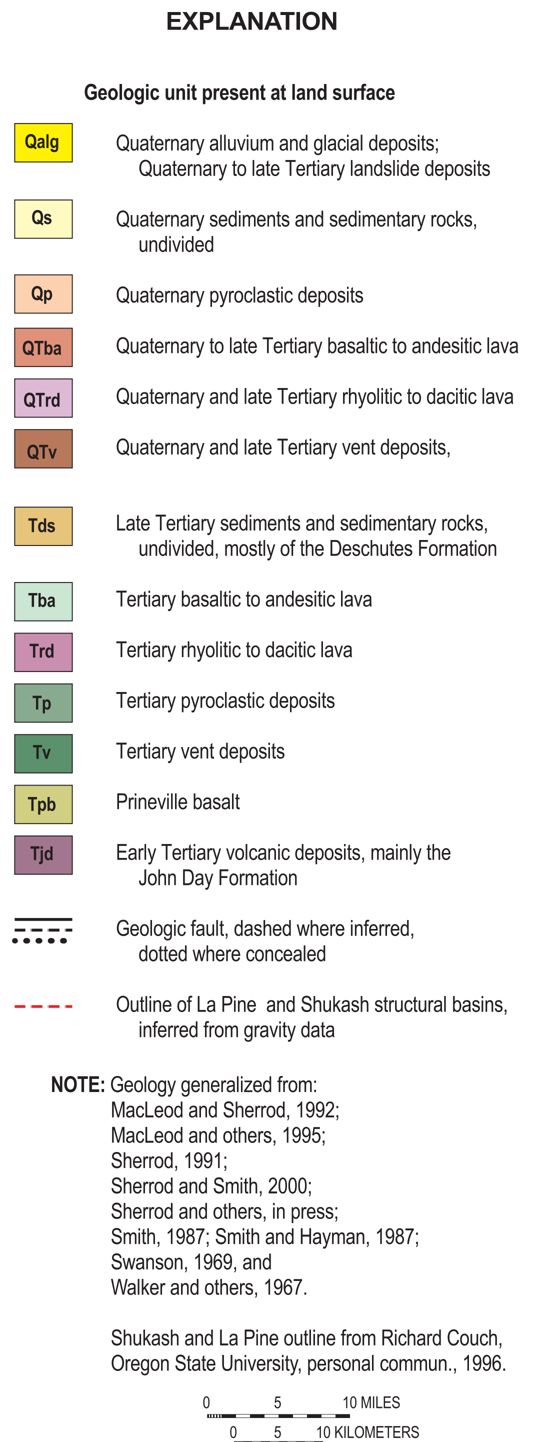
This section briefly describes the geologic framework of the regional ground-water flow system in the upper Deschutes Basin, including a brief description of the major geologic units, geologic structure, and the geologic factors controlling the flow-system boundaries.

Geologic Controls on Regional Ground-Water Flow

The upper Deschutes Basin has been a region of volcanic activity for at least 35 million years (Sherrod and others, in press), resulting in complex assemblages of volcanic vents and lava flows, pyroclastic deposits, and volcanically derived sedimentary deposits (fig. 4). Volcanic processes have created many of the present-day landforms in the basin. Glaciation and stream processes have subsequently modified the landscape in many places.

Most of the upper Deschutes Basin falls within two major geologic provinces, the Cascade Range and the Basin and Range Province (Orr and others, 1992). The processes that have operated in these provinces have overlapped and interacted in much of the upper Deschutes Basin. The Cascade Range is a north-south trending zone of compositionally diverse volcanic eruptive centers and their deposits extending from northern California to southern British Columbia. Prominent among the eruptive centers in the Deschutes Basin are large stratovolcanoes such as North, Middle, and South Sister, and Mount Jefferson, all of which exceed 10,000 ft in elevation. The Cascade Range is primarily a constructional feature, but its growth has been accompanied, at least in places, by subsidence of the range into a north-south trending graben (Allen, 1966). Green Ridge is the eastern escarpment of one of the graben-bounding faults. The Basin and Range Province is a region of crustal extension and is characterized by subparallel fault-bounded down-dropped basins separated by fault-block ranges. Individual basins and intervening ranges are typically 10 to 20 miles across. The Basin and Range Province, which encompasses much of the interior of the Western United States, extends from central Oregon south through Nevada and western Utah, and into the southern parts of California, Arizona, and New Mexico. Although the Basin and Range Province is primarily structural, faulting has been accompanied by widespread volcanism. The major stratigraphic units in the upper Deschutes Basin are described below in approximate order of their age.

The oldest rocks in the upper Deschutes Basin study area (unit Tjd in fig. 4) are part of the late Eocene to early Miocene John Day Formation and consist primarily of rhyolitic ash-flow tuffs, lava flows, tuffaceous sedimentary rocks, and vent deposits. The John Day Formation ranges in age from 22 to 39 million years and is as much as 4,000 ft thick (Smith and others, 1998). Rocks of the John Day Formation have very low permeability because the tuffaceous materials are mostly devitrified (changed to clays and other minerals) and lava flows are weathered and contain abundant secondary minerals. Because of the low permeability, ground water does not easily move through the John Day Formation, and the unit acts as a barrier to regional ground-water flow. The John Day Formation constitutes the eastern and northern boundary of the regional ground-water flow system. The John Day Formation, or equivalent rocks, are presumed to underlie much of the upper Deschutes Basin and are considered the lower boundary of the regional flow system throughout much of the study area.



The Prineville basalt (unit Tpb in figure 4) overlies the John Day Formation in the northeastern part of the study area. Radiometric techniques indicate that the Prineville basalt is 15.7 million years old (Smith, 1986). The Prineville basalt, which is up to 700 ft thick, is locally fractured, contains permeable interflow zones, and is locally an important aquifer.

The Deschutes Formation, which overlies the Prineville basalt, consists of a variety of materials deposited in an alluvial basin east of the Cascade Range, including lava flows, ignimbrites, fallout tephra, debris flows, hyperconcentrated flood deposits, and alluvium. Most of the deposits originated in the Cascade Range and were shed eastward into the basin, but some originated from intrabasin eruptive centers or were eroded from older (John Day Formation) uplands to the east. The Deschutes Formation was deposited in a rapidly filling basin with a constantly changing drainage system between about 4.0 and 7.5 million years ago (Smith, 1986). Deposition of many units within the formation was restricted to canyons and other short-lived topographic lows. Consequently, individual strata within the Deschutes Formation typically have limited geographic distribution resulting in a heterogeneous sequence. Most of the areas mapped as Tds, Tba, Tp, and Tv in figure 4 are generally recognized as part of the Deschutes Formation. Some areas so mapped in southern part of figure 4 are not generally considered part of the Deschutes Formation, but are composed of rocks similar in composition and age to the Deschutes Formation, and likely have similar hydrologic characteristics.

Strata within the Deschutes Formation were deposited in three main depositional environments (Smith, 1986). The westernmost depositional environment was a broad plain adjacent to the Cascade Range, on which a variety of materials were deposited, including flood and debris-flow deposits, ignimbrites, fallout tephra, and lava flows. The ancestral Deschutes River was another depositional environment, occurring along the eastern margin of the alluvial plain. Deposits in the ancestral Deschutes River environment include well-sorted conglomerates and coarse sandstone, fine sandstone, mudstone, and intracanyon lava flows. A third depositional environment existed along the inactive eastern margin of the basin. Here, material eroded from the highland of older rock to the east (mostly John Day Formation) was redeposited, resulting in beds of poorly sorted angular gravel and sand, reworked pyroclastic debris, and fine-grained sediment.

The Deschutes Formation is the principal aquifer unit in the upper Deschutes Basin. The unit ranges in thickness from zero where it contacts the underlying John Day Formation or Prineville basalt to over 2,000 ft at its westernmost exposure at Green Ridge. Permeable zones occur throughout the Deschutes Formation. The lava flows, vent deposits, and sand and gravel layers in the Cascade Range-adjacent alluvial plain facies and the ancestral Deschutes River facies are locally highly permeable. Two sequences of lava flows in the Deschutes Formation, the Opal Springs basalt, which is up to 120 ft thick, and the Pelton basalt, which may locally exceed 400 ft in thickness, are notable aquifers and locally discharge large amounts of water where exposed in the canyons of the Deschutes and Crooked Rivers. The inactive margin facies is less permeable because of poor sorting and a high degree of weathering.

Rhyolite and rhyodacite domes (unit Trd in figure 4) occur in the north-central part of the study area and are locally interbedded with the Deschutes Formation. These materials form Cline Buttes and also crop out in the area between the Deschutes River and Squaw Creek north of Lower Bridge. These rocks are locally highly fractured and permeable. Numerous springs discharge from permeable zones in this unit where it is exposed in the canyon of the Deschutes River near Steelhead Falls (Ferns and others, 1996).

The Cascade Range and volcanic deposits of similar age elsewhere in the basin overlie the Deschutes Formation and constitute the next major composite stratigraphic unit. These deposits include units Qp, QTba, QTrd, and QTv in figure 4. This composite unit, which is likely several thousand feet thick, is composed of lava flows, domes, vent deposits, pyroclastic deposits, and volcanic sediments. Most are Quaternary in age (younger than 1.6 million years old). This unit includes the entire Cascade Range and Newberry Volcano to the east. Much of this material is highly permeable, especially the upper several hundred feet. Permeability of the unit is greatly reduced at depth beneath the Cascade Range, however, due to hydrothermal alteration and secondary mineralization (Blackwell and others, 1990; Blackwell, 1992; Ingebritsen and others, 1992). Temperature gradient data (Swanberg and others, 1988) and hydrothermal mineralization studies (Keith and Barger, 1988, 1999) suggest a similar loss of permeability at depth beneath Newberry Volcano. The top of the region at depth beneath the Cascade Range and Newberry Volcano where permeability is reduced by several orders of magnitude due to hydrothermal mineralization is considered, for the purposes of this study, to be the base of the regional ground-water flow system in these areas.

The Cascade Range and volcanic deposits of similar age are highly permeable at shallow depths. The near-surface deposits are often highly fractured or otherwise porous and largely lack secondary mineralization. The Cascade Range is the principal ground-water recharge area for the upper Deschutes Basin, and these deposits are the principal avenue by which most ground water moves from the recharge area out into the basin. Because there are very few wells in the Cascade Range and on Newberry Volcano, there is little information on the distribution of hydraulic head or subsurface conditions.

The youngest units in the upper Deschutes Basin are Quaternary sedimentary deposits. These deposits include alluvium along modern flood plains, landslide deposits, and glacial drift and outwash (unit Qalg on figure 4). Undifferentiated Quaternary sedimentary deposits resulting from a variety of depositional processes are mapped as Qs in figure 4. Many of the Quaternary sedimentary deposits in the basin are too thin or discontinuous to affect regional ground-water flow. However, glacial deposits, particularly outwash deposits, are sufficiently thick and widespread to be significant. Glacial deposits, generally porous and permeable, are an important source of ground water along the margin of the Cascade Range, for example in the area around the city of Sisters. Alluvial sand and gravel deposits also form an important aquifer in the La Pine subbasin (fig. 4).

Geologic structure, principally faults and fault zones, can influence ground-water flow. Fault zones can act either as barriers to or conduits for ground-water flow, depending on the nature of the material in and between the individual fault planes. Faults most commonly affect ground-water flow by juxtaposing rocks of contrasting permeability or by affecting the patterns of deposition. Structural basins caused by faulting can act as depositional centers for large thicknesses of sediment or lava that may influence regional ground-water flow. Faults do not always influence ground-water flow; there are regions in the upper Deschutes Basin where ground-water flow appears unaffected by the presence of faults.

There are four prominent fault zones in the upper Deschutes Basin (fig. 4). Green Ridge, north of Black Butte, is a prominent north-south trending escarpment caused by faulting along the margin of the Cascade graben. The region to the west of Green Ridge has dropped as much as 3,000 ft (Conrey, 1985). This fault movement has juxtaposed rock materials of contrasting permeability, and subsidence west of the fault system has created a depositional basin for accumulation of volcanic and glacial materials from the Cascade Range. A large amount of ground water discharges to the Metolius River along the western side of the Green Ridge escarpment. It is possible that the ground-water discharge occurs because the Green Ridge fault zone acts as a barrier to the eastward flow of ground water from the Cascade Range. It is also possible that discharge occurs because the western side of the escarpment is a regional topographic low.

The Sisters fault zone is a north-northwest trending zone of normal faults that extends from the north flank of Newberry Volcano to the south end of Green Ridge near Black Butte. Escarpments of some faults along the Sisters fault zone have impounded lava flows from the Cascade Range and prevented flow into lower-elevation areas toward the northeast. Escarpments along the Sisters fault zone also have caused local accumulation of glacial sediments. Although the Sisters fault zone affects the occurrence of shallow ground water by controlling the deposition of glacial sediment, it does not appear to affect ground-water flow at depth.

The Brothers fault zone is a major northwest-trending zone of normal faults that extends from southeastern Oregon to the north flank of Newberry Volcano. Faults along this zone are covered by lava flows from Newberry Volcano and do not appear to offset those flows. The influence of the Brothers fault zone on regional ground-water flow is unknown.

The Walker Rim fault zone is a major northeast-trending zone that extends from Chemult to the south flank of Newberry Volcano. The region to the west has dropped as much as 2,500 ft (feet). The influence of this fault zone on ground-water flow is unknown.

The La Pine and Shukash structural basins (fig. 4) are complex graben structures extending from Newberry Volcano to the crest of the Cascade Range. Much of what is known of these features is from interpretations of gravity data by Couch and Foote (1985, and written commun., 1996). The La Pine graben is a present-day landform, and well data shows that it has accumulated over 1,000 ft of sediment, much of which is fine grained. The Shukash basin, in contrast, has no surface expression, is mostly covered by younger volcanic and glacial deposits, and its existence is inferred largely from gravity data. The sediment thickness at the center of the basin is inferred to be about 2,500 ft. The nature of sediment fill is poorly known, but where exposed or drilled, the sediment in the Shukash basin is similar to that of the La Pine basin. The fine-grained sediment fill in the La Pine and Shukash basins has low permeability. The presence of large springs on the margins of the La Pine and Shukash basins may be due to the juxtaposition of permeable Cascade Range volcanic rocks with the low-permeability basin-fill deposits. The faults bounding both of these grabens are largely obscured by younger volcanic deposits.

Hydraulic Characteristics of Subsurface Materials

As described in the preceding section, geologic materials possess certain hydraulic characteristics that control the movement and storage of ground water. This section describes quantitative terms that represent those characteristics and presents estimates or ranges of values of those terms for various materials in the upper Deschutes Basin. A more thorough discussion of the terms used to describe the hydraulic characteristics of aquifers and aquifer materials can be found in any basic ground-water hydrology text such as Freeze and Cherry (1979), Fetter (1980), or Heath (1983).

The term *permeability* was introduced in the last section as a measure of the resistance to fluid flow offered by a particular rock type. Permeability is an intrinsic property of the rock type, and is independent of the fluid properties. In ground-water studies, the term *hydraulic conductivity* is used more commonly than permeability. The hydraulic conductivity term includes both the properties of the rock (the intrinsic permeability) and the properties of the water, such as viscosity and density. Hydraulic conductivity is defined as the volume of water per unit time that will pass through a unit area of an aquifer material in response to a unit hydraulic-head gradient. Hydraulic conductivity has the units of volume per unit time (such as cubic feet per day) per unit area (such as square feet), which simplifies by division to length per unit time (such as feet per day). Hydraulic-conductivity values for aquifer materials commonly span several orders of magnitude from less than 0.1 ft/d (feet per day) for fine sand and silt to over 1,000 ft/d for well-sorted sand and gravel.

When discussing aquifers instead of rock types, the hydraulic conductivity is often multiplied by the aquifer thickness to derive a term known as *transmissivity.* Transmissivity is defined as the volume of water per unit time that will flow through a unit width of an aquifer perpendicular to the flow direction in response to a unit hydraulic-head gradient. Transmissivity has units of volume per unit time (such as cubic feet per day) per unit aquifer width (such as feet) which simplifies to length squared per unit time (such as square feet per day).

The storage characteristics of an aquifer are described by the *storage coefficient*. The storage coefficient is defined as the volume of water an aquifer releases from, or takes into, storage per unit area of aquifer per unit change in head. The volume of water has units of length cubed (such as cubic feet), the area has units of length squared (such as square feet), and the head change has units of length (such as feet). Thus, the storage coefficient is dimensionless. Storage coefficients typically span several orders of magnitude from 10^{-4} for aquifers with overlying confining units, to 0.1 for unconfined aquifers.

Aquifer Tests

The hydraulic characteristics of subsurface materials in the basin have been estimated using data from aquifer tests, some of which were conducted as part of this study, and specific-capacity tests conducted by drillers upon completion of new wells. An aquifer test consists of pumping a well at a constant rate and measuring the change in water level (the drawdown) with time. The data collected allow generation of a curve showing the change in drawdown as a function of time. Similar data are collected after the pumping is stopped, allowing generation of a curve showing the water-level recovery as a function of time. These data are collected not only from the pumped well, but from nearby wells (called observation wells) in which the water level may be affected by the pumping. Analysis of the drawdown and recovery curves in the pumped well and observation wells provides estimates of the transmissivity and storage coefficient of the aquifer.

Four aquifer tests were conducted as part of this study (fig. 5). Each involved pumping a large-capacity public-supply well and observing drawdown and recovery in nearby nonpumped wells. In addition, results from seven aquifer tests conducted by private consultants were available. A common problem encountered in many of the tests was the inability to stress the aquifer sufficiently to induce an interpretable effect in the observation wells. In other words, the aquifer transmissivity is so large in some places that pumping a well in excess of 1,000 gal/min (gallons per minute) may produce only a few hundredths of a foot of drawdown in an observation well just a few hundred feet from the pumped well.

Aquifer tests were conducted for this study on wells belonging to the cities of Madras, Redmond, and Bend, as well as Juniper Utilities, a privately owned water utility. Each of the tests is summarized in table 1 and described in the following paragraphs. The location of the tested wells is shown in figure 5.

The city of Madras test involved pumping City Well No. 2 at 351 gal/min for 3 days and monitoring the response in the pumped well and in an observation well 250 ft from the pumped well. The pumped well produces from a layer of sand and gravel at the base of a sequence of lava flows. The producing sediments are part of the inactive-margin facies of the Deschutes Formation (fig. 5). Both the pumped well and the observation well showed good responses to the pumping, with maximum drawdowns of 36.20 and 17.67 ft respectively. The drawdown and recovery curves were typical of a confined aquifer (Lohman, 1979). The test yielded a transmissivity estimate of 1,700 to 2,500 ft^2/d (square feet per day) and a storage coefficient estimate of 0.0001 to 0.0002.

The city of Redmond test consisted of pumping City Well No. 3 at 1,141 gal/min for 3 days and monitoring the response in the pumped well and an observation well 350 ft from the pumped well. The well produces from a combination of lava flows and sand and gravel layers in the Cascades-adjacent alluvial plain or ancestral Deschutes River facies of the Deschutes Formation. Interpretation of the results of this test was complicated by the very small response in the observation well. Total drawdown in the observation well after 3 days of pumping was only 0.16 ft, which is close to the range of observed pre-test water-level fluctuations caused by external influences such as barometric pressure changes and earth tides. Drawdown in the pumping well (11.67 ft) was dominated by well losses (excessive drawdown in the well bore due to well inefficiency) so only the recovery data from the pumped well was usable. The drawdown and recovery curves resulting from this test were not typical of a confined aquifer. The drawdown followed the typical Theis curve (Lohman, 1979) near the beginning of the test, but later deviated from the curve, indicating that drawdown was less than would be expected for a confined aquifer. The exact cause of this behavior is unknown, but similar behavior is observed in aquifers where drainage of water from overlying strata cause a delayed-yield response (Neuman, 1975). Analysis of the test results yielded a transmissivity estimate of 2.0×10^5 ft^2/d to 3.0×10^5 ft^2/d, and a storage coefficient estimate of 0.05.

Table 1. Summary of selected aquifer tests in the upper Deschutes Basin, Oregon

[OWRD well no., Oregon Water Resources Department well number; gpm, gallons per minute; Analysis method: T, Theis non-equilibrium curve matching; SL, straight-line method; SC, Theis solution using specific-capacity data (Lohman, 1979). Test conducted by: S, study team; C, private consultant. ft^2/d, square feet per day; ft, feet; ft/d, feet per day; ---, value not determined; ∗, time-averaged pumping rate]

Well number	OWRD well no.	Well name	Discharge (gpm)	Duration (hours)	Draw-down	Distance from pumped well	Analysis method	Test conducted by	Transmissivity (ft^2/d)	Storage coefficient	Aquifer material	Open interval length (ft)	Hydraulic conductivity (ft/d)
11S/13E-01BCA2	JEFF0427	City of Madras Well 2	351	72	17.67	250	T, SL	S	1.7×10^3 to 2.5×10^3	0.0001 to 0.0002	Sand & gravel	16	110 to 160
14S/10E-30DDB2	DESC1835	Cascade Meadows	1,000	4.5	29.3	0	SC	C[1]	5×10^4	---	Mafic lava	88	600
14S/12E-13DC	DESC8593	City of Redmond MW-8	11.4	24	8.25	0	S	C[2]	1.8×10^2	---	Basalt	13	14
15S/10E-05BBB	DESC2999	Tollgate	1,210	48	4.5	0	SC	C[3]	1×10^4	---	Glacial outwash & mafic lava	100	100
15S/13E-20CDC	DESC0407	City of Redmond Well 4	∗2,751	72	1.8	844	T	C[4]	6×10^3	.1	Sand & gravel	100	60
15S/13E-22CBA2	DESC3951	City of Redmond Well 3	1,141	72	.16	350	T, SL	S	2.0×10^5 to 3.0×10^5	.05	Sand & gravel, lava	132	1.5×10^3 to 2.3×10^3
17S/11E-11BAs	DESC1304	Awbrey Butte	225	24	.65	0	SC	C[5]	1×10^5	---	Basalt with minor cinders	105	100
18S/9E-20BDA	DESC5215	Mount Bachelor	110	18	39	0	SC	C[6]	5×10^2	---	Basaltic cinders	56	9
18S/12E-BDB1	DESC5576	Brooks-Scanlon No. 1	1,270	17	7.61	0	SC	C[7]	5×10^4	---	Lava & tuff	253	1.5×10^2
18S/12E-07DBD2	DESC9108	City of Bend Rock Bluff	722	24	.06	210	---	S	---	---	Basaltic cinders	395	---
18S/12E-16BCC3	DESC5613	Juniper Utility	1,300	3	1.14	35	---	S	---	---	Lava & cinders	122	---

Sources for tests by private consultants:

[1] Century West Engineering Corporation, 1990, Regional hydrogeologic investigation for the proposed Shadow Mountain R.V. Park, Sisters, Oregon.
[2] Cascade Earth Sciences Limited, 1994, City of Redmond, Oregon, wastewater treatment system expansion: hydrogeologic characterization report.
[3] Century West Engineering Corporation, 1990, Preliminary pump test data and groundwater material for the proposed Shadow Mountain R.V. Park, Sisters, Oregon.
[4] Century West Engineering Corporation, 1985, Assessment of water availability, Redmond, Oregon.
[5] W and H Pacific, 1994, Brooks Resources Corporation groundwater appropriation claim of beneficial use and site report, Permit No. G-1106.
[6] Century West Engineering Corporation, 1984, Water-supply study for Mount Bachelor.
[7] CH2M, 1964, A report on an engineering study of the municipal water system, city of Bend, Oregon.

The city of Bend test involved pumping one of the wells at the city's Rock Bluff well field south of town at 722 gal/min for a period of 24 hours. This well produces from basaltic lava and cinders of the Deschutes Formation, which is predominantly lava at this location. The response was measured in a nearly identical observation well 210 ft from the pumped well. There was no access to the pumped well for water-level measurements. The drawdown in the observation well was less than 0.06 ft, which is well within the range of water-level fluctuations caused by external influences such as barometric pressure changes and earth tides. The small drawdown due to pumping could not be satisfactorily separated from the water-level fluctuations due to external influences, and no quantitative analysis was possible. The small drawdown in this well, however, suggests a large transmissivity of a magnitude similar to that estimated from the city of Redmond well test.

The fourth aquifer test conducted for this study involved pumping a production well belonging to Juniper Utilities, south of Bend, at 1,300 gal/min for just over 3 hours. This well produces from basaltic lava with minor interbedded cinders which are likely correlative to the Deschutes Formation. Drawdown and recovery were measured in an observation well 35 ft from the pumped well and open to the same water-bearing strata. There was no access for water-level measurements in the pumped well. The drawdown in the observation well, which totaled 1.14 ft after 3 hours, did not follow the Theis curve for a confined aquifer (Lohman, 1979). The drawdown departed from the Theis curve about 7 minutes into the test in a manner indicating that drawdown was less than would be expected for a confined system. After about 50 minutes the water level stabilized and drawdown did not increase for the duration of the test, indicating that the cone of depression encountered a source of recharge equal to the well discharge. The likely source of recharge was leakage from large (hundreds of cubic feet per second) unlined irrigation canals within 3,000 ft of the pumped well. Analysis of recovery data also indicated the aquifer received recharge during the test. The short duration of this test and the atypical response in the observation well precluded a reliable estimation of hydraulic parameters. The relatively small total drawdown in the observation well suggests a large transmissivity.

Results from seven additional aquifer tests conducted by consultants are summarized in table 1. Most of these tests were affected by one or more problems such as insufficient response in observation wells, measurement errors, variable pumping rates, effects of well losses in the pumping well, and recharge effects. Time-drawdown data from five of the tests were not suitable for type-curve analysis, but the tests did allow calculation of the specific capacity of the wells. Specific capacity is a general measure of well performance and is calculated by dividing the rate of pumping by the amount of drawdown and typically has units of gallons per minute per foot of drawdown. Transmissivities were estimated from specific-capacity data using an iterative technique based on the Jacob modified nonequilibrium formula (Ferris and others, 1962, p. 98; Vorhis, 1979).

Transmissivity estimates from aquifer tests are affected by well construction and the thickness of the aquifer open to the well. In order to allow meaningful comparisons between aquifer tests, transmissivity estimates can be normalized by dividing them by the length of the open interval below the water table in the pumped well to derive an estimated hydraulic conductivity. Hydraulic-conductivity values so calculated are included in table 1. Hydraulic-conductivity estimates derived from aquifer tests vary more than two orders of magnitude, from less than 10 to nearly 1,900 ft/d. The variation in hydraulic conductivity of subsurface materials is undoubtedly much greater than indicated by the tests. Production zones in wells are not a true sample of the range in hydraulic conductivities in the subsurface because the wells are selectively open to the most permeable strata and less permeable zones are not represented.

Hydraulic-conductivity values from the available tests do not correlate well with rock type. Tests yield a wide range of values from both volcanic and sedimentary aquifers. This is not surprising because hydraulic conductivities of both types of materials can range over several orders of magnitude (Freeze and Cherry, 1979, table 2.2). The small number of tests precludes determination of the spatial distribution of hydraulic conductivity. The highest hydraulic-conductivity values, however, are associated with Deschutes Formation materials, including basaltic lava and vent deposits, and sand and gravel deposits likely belonging to the ancestral Deschutes River channel facies described by Smith (1986).

Well-Yield Tests

Another source of information on subsurface hydraulic characteristics are the well-yield tests conducted by drillers and reported on the well logs submitted on completion of all new wells. Well-yield tests generally consist of a single drawdown measurement taken after a well has been pumped at a specified rate for a specified length of time, typically 1 hour. Well-yield tests allow determination of a well's specific capacity, which can be used to estimate transmissivity as described previously. Specific capacity is only a semiquantitative measure of well performance in that it can vary with pumping rate. Specific-capacity values can be used to calculate only rough estimates of the aquifer transmissivity and provide no information on the aquifer storage characteristics. Although transmissivity values calculated from specific-capacity tests are only approximate, they can be used to evaluate the relative differences in hydraulic characteristics between different geographic areas if data are available from a sufficient number of wells.

Well-yield tests were evaluated from 1,501 field-located water wells (raw data are in Caldwell and Truini, 1997). Of these tests, 390 were air-lift tests, in which the water is blown out of the well using compressed air, precluding measurement of drawdown and calculation of specific capacity. An additional 152 tests had information that was incomplete in some other way. Of the 959 remaining yield tests, 453 had pumping (or bailing) rates that did not sufficiently stress the aquifer to produce a measurable effect in the well, and zero drawdown is indicated on the well log.

This precludes calculation of a specific capacity because if drawdown is zero then specific capacity is infinite, a physical impossibility. Eliminating wells with drawdown shown as zero from the data set would have selectively removed wells representing the most transmissive areas. To avoid biasing the data in this manner, wells with zero drawdown were arbitrarily assigned a drawdown of 1 ft, which is the limit of precision to which most drillers report water levels, and probably the limit to which it is measured during bailer tests. Statistics for specific capacities derived from well-yield tests in the study area and from various subareas within the study area are shown in table 2.

A map showing the geographic distribution of transmissivity estimates derived from well-yield tests can be used to help understand spatial variations in aquifer characteristics. When creating such maps, it is important to include only wells with comparable construction. Certain wells, such as high-yield municipal and irrigation wells are constructed to be very efficient, and consequently have higher specific capacities than small-yield household wells in the same aquifer. Therefore, it is desirable to use only wells with comparable construction when creating maps showing transmissivities estimated from specific-capacity data.

The geographic distribution of transmissivities estimated from specific capacities of 623 household wells is shown in figure 5. Although a wide range of transmissivity values occurs throughout the areas represented, some subtle patterns are apparent.

Table 2. Statistics for transmissivities (square feet per day) estimated from specific-capacity data for subareas in the upper Deschutes Basin, Oregon

[*, includes wells outside the listed subareas]

Area	Minimum	25th Percentile	Median	75th Percentile	Maximum	Number of wells
La Pine Subbasin Alluvium	7.1	342	901	1,953	114,297	175
Deschutes Formation West	11.4	617	1,917	3,587	1,458,724	382
Deschutes Formation East	12.6	1,099	2,337	4,063	221,887	209
Inactive Margin	1.1	46.2	796	2,225	59,683	92
All located wells*	1.1	518	1,821	3,660	1,458,724	959

The La Pine subbasin, the area just north of Bend, Jefferson County, and the eastern margin of the study area show the highest incidence of wells with low transmissivity values. The areas east of Bend, between the Crooked and Deschutes Rivers near Redmond, and west of Sisters show the highest incidence of high transmissivity wells. This distribution is consistent with the results of aquifer tests and with the regional geology. The areas where transmissivities appear to be slightly higher coincide with regions of coarse-grained sedimentary deposits, such as the glacial outwash west of Sisters and the ancestral Deschutes River channel deposits in the Redmond area. The areas where transmissivities appear lower coincide, at least in part, with regions where fine-grained materials predominate, such as the La Pine subbasin, or regions where older rock or sediments derived from older rock predominate, such as the eastern and northern parts of the upper Deschutes Basin.

The aquifer tests described above provide information on aquifer characteristics at specific locations, and taken as a group provide a general picture of the minimum range of conditions and of geographic variations in the areas represented. The specific-capacity values from well-yield tests provide a rough picture of the geographic distribution of transmissivity. The aquifer-test and specific-capacity data described in this section, however, represent only a small part of the flow system. There are large geographic areas in the upper basin, such as the Cascade Range and Newberry Volcano area, where there are virtually no data. Moreover, in areas of the upper Deschutes Basin where wells are plentiful, most wells penetrate only the upper part of the saturated zone and may not be representative of the deep parts of the flow system.

GROUND-WATER RECHARGE

The Deschutes Basin ground-water flow system is recharged by infiltration of precipitation (rainfall and snowmelt), leakage from canals, infiltration of applied irrigation water that percolates below the root zone (on-farm losses), and leakage from streams. Recharge from all of these processes is discussed in this section. The amounts of recharge from each of the processes cannot be simply summed to determine the net recharge for the upper Deschutes Basin because some water cycles into and out of the ground-water system twice. For example, the water that recharges the ground-water system through canal leakage originates as streamflow, a large percentage of which originates as springflow in the Cascade Range. The ground water supplying the springs originates from infiltration of precipitation in the Cascade Range.

Infiltration of Precipitation

Recharge from precipitation occurs where rainfall or snowmelt infiltrates and percolates through the soil zone and, eventually, reaches the saturated part of the ground-water flow system. Recharge is the quantity of water remaining after runoff and evapotranspiration take place.

The spatial and temporal distribution of ground-water recharge to the upper Deschutes Basin from infiltration of precipitation were estimated for water years 1962–97 using a water-balance model. The model, referred to as the Deep Percolation Model, or DPM, was developed by Bauer and Vaccaro (1987) for a regional analysis of the Columbia Plateau aquifer system in eastern Washington. The DPM is based on well-established empirical relations that quantify processes such as interception and evaporation, snow accumulation and melt, plant transpiration, and runoff. The DPM has been successfully applied to estimate regional recharge for studies of the Goose Lake Basin in Oregon and California (Morgan, 1988), the Portland Basin in Oregon and Washington (Snyder and others, 1994), and several other areas in Oregon and Washington. A detailed description of the application of the DPM to the Deschutes Basin, including the data input, can be found in Boyd (1996). The following sections provide a summary of the methodology and results.

The DPM was applied to the entire upper Deschutes Basin by subdividing the basin into 3,471 equal-sized grid cells with dimensions of 6,000 ft by 6,000 ft (fig. 6). The DPM computed a daily water balance at each cell using input data describing the location, elevation, slope, aspect, mean annual precipitation, land cover, and soil characteristics of each cell. Daily data (precipitation, maximum and minimum temperature, solar radiation) from six weather stations (table 3) in the basin were used to compute daily moisture input and potential evapotranspiration at each cell. The six climate stations used were selected because they had the longest periods of record with the fewest occurrences of missing data among stations in the basin. Climate data were obtained from the Oregon Climate Service (1999).

The DPM requires that several types of data be specified for each cell: long-term average annual precipitation, land-surface elevation, slope, aspect, land-cover type, and soil type. Long-term average annual precipitation at each cell was derived from a statewide distribution for the 1961–90 period estimated by the Oregon Climate Service using the PRISM model (Daly and Nielson, 1992). PRISM uses digital topographic data to account for orographic effects on precipitation. The DPM uses the ratio of the long-term annual average precipitation at the cell to the long-term average at each climate station to interpolate daily precipitation values at each cell.

The mean elevation, slope, and aspect of each cell were calculated from 90-meter digital elevation data using a geographic information system (GIS). Elevation was used with temperature lapse rates to interpolate daily temperature values at each cell from the nearest climate stations. Slope at each cell was used to compute runoff and aspect was used to estimate incident solar radiation in the calculation of potential evapotranspiration.

Land-cover data from the Oregon Gap Analysis Program (J. Kagan, Oregon Natural Heritage Program, written commun., 1992) was used to specify four land-cover types in the model: forest, sage and juniper, grass, and surface water. These types covered 61, 36, 2, and 1 percent of the basin, respectively. Recharge from irrigated croplands was not estimated using DPM; estimates of recharge to these areas from canal leakage and on-farm losses are described later in this section. For each land-cover type, the maximum plant rooting depth, foliar cover fraction, and interception storage capacity were specified based on literature values (Boyd, 1996).

A statewide soil database (STATSGO) (U.S. Department of Agriculture, 1991) was used to specify soil type and associated parameters at each cell. A cluster analysis was used to aggregate the 26 general soil types found within the basin into 10 hydrologic soil types (Boyd, 1996). For each hydrologic soil type, thickness, texture, field capacity, specific yield, horizontal hydraulic conductivity, and vertical hydraulic conductivity were specified.

The DPM was used to compute daily water balances at each cell from January 1961 through November 1997. The daily recharge values were used to compute mean monthly and annual recharge values.

The distribution of mean annual recharge for water years 1993–95 (fig. 6) illustrates the strong relation between precipitation (fig. 3) and recharge. Recharge for the 1993–95 period was calculated to correspond to the calibration period for a steady-state numerical ground-water flow model. Computed recharge from precipitation ranged from less than 1 in./yr in the lower elevations, where annual precipitation is less than 12 inches, to more than 130 inches in the high Cascade Range, where soils are thin and precipitation locally exceeds 200 inches. The mean recharge for the basin during the 1993–95 water years was 10.6 in./yr; converted to a mean annual value for the 4,500 mi^2 basin, this is the equivalent of about 3,500 ft^3/s (cubic feet per second).

Table 3. Weather stations used for estimation of recharge from infiltration of precipitation with the Deep Percolation Model
[ID, identification; X, data collected]

Station name	Station ID	Elevation, in feet	Precipitation data	Temperature data	Solar-radiation data
Bend	0694	3,650	X	X	
Brothers	1067	4,640		X	
Madras	5139	2,230	X		
Prineville	6883	2,840	X	X	
Redmond	7062	3,060	X	X	X
Wickiup Dam	9316	4,360	X	X	

Between 1962 and 1997, estimated recharge ranged from less than 3 inches in the drought years of 1977 and 1994 to nearly 23 inches in 1982 (fig. 7). The mean for the 26-year period was 11.4 in./yr, which converts to an annual rate of about 3,800 ft^3/s. The estimated evapotranspiration for the basin is relatively constant from year to year because the effects of above or below normal precipitation are dampened by storage in the soil moisture zone. Runoff is a relatively small component of the total water budget in the Deschutes Basin due to high infiltration rates of the permeable volcanic soils. The Deschutes and Metolius Rivers are noted for their extraordinarily constant flows that are sustained primarily by ground-water inflow. Recharge averages about 35–40 percent of annual precipitation within the basin, but ranges from less than 5 percent at low elevations, where potential evapotranspiration greatly exceeds precipitation, to as much as 70 percent at higher elevations, where annual precipitation may be several times greater than potential evapotranspiration.

Manga (1997) developed a physically based model using the Boussinesq equation (Boussinesq, 1904) to estimate recharge rates within the contributing areas of four spring-dominated streams tributary to the Deschutes River above Benham Falls. Results agreed well with those from the DPM for the area. Within the inferred contributing areas to all four streams, mean DPM recharge was 29 in./yr (1962–97) and mean recharge estimated by Manga was 28 in./yr (1939–91). Manga's estimated recharge averages 56 percent of precipitation within the contributing area of the four streams, while the DPM recharge was approximately 45 percent of precipitation within the same area.

About 84 percent of recharge from infiltration of precipitation occurs in the Deschutes Basin between November and April (fig. 8). According to the DPM, recharge rates peak in December and again in March–April. The December recharge peak results from deep percolation of precipitation after heavy fall rains and early winter snowfall and melt have saturated soils. After January, precipitation is reduced, but snowmelt sustains recharge at higher elevations through April. By May, increasing evapotranspiration begins to deplete soil moisture storage and reduce recharge rates to nearly zero.

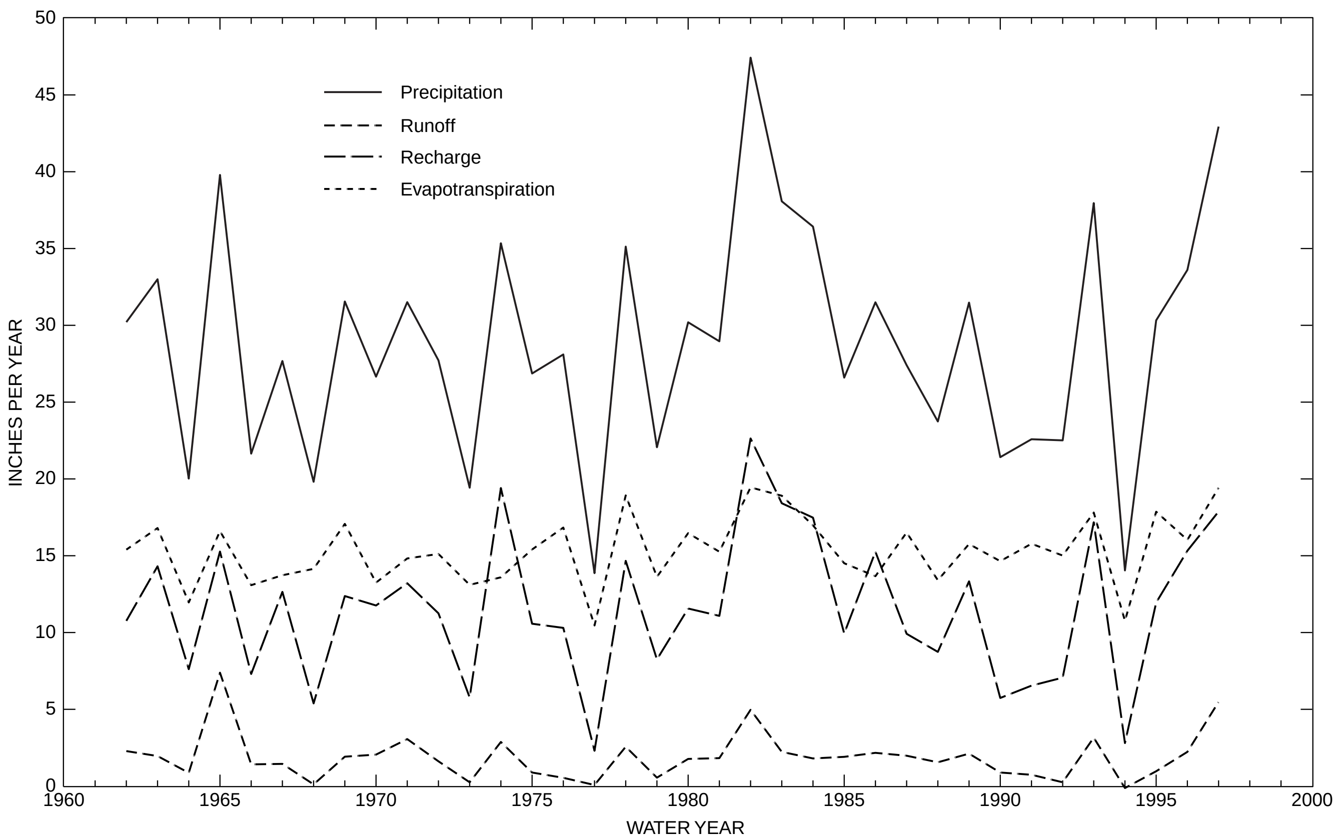


Figure 7. Annual mean components of the basinwide water budget, estimated using the Deep Percolation Model for water years 1962–97.

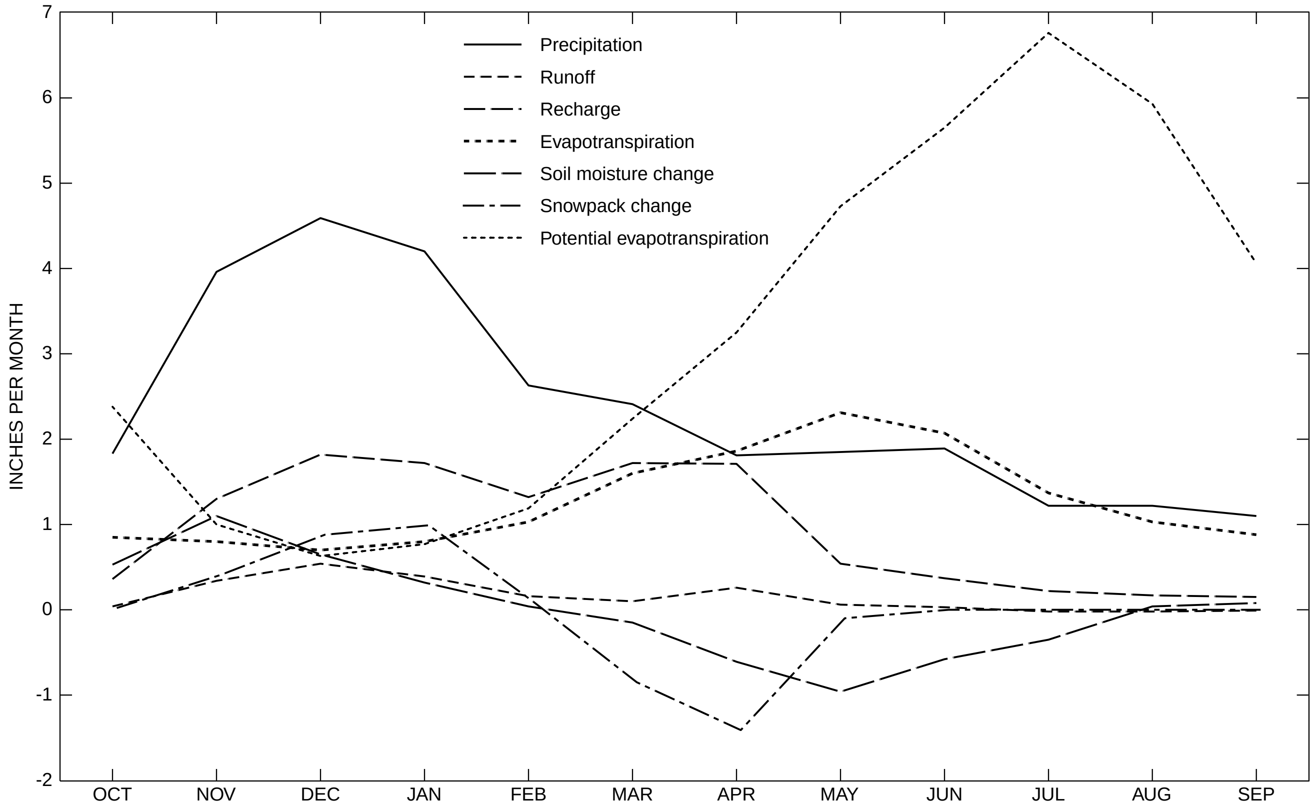


Figure 8. Mean monthly components of the basinwide water budget, estimated using the Deep Percolation Model for water years 1962–97.

Canal Leakage

There are approximately 720 miles of canals and laterals that carry water diverted from the Deschutes and Crooked Rivers to more than 160,000 acres of irrigated lands in the basin. Many of the canals are cut into young basaltic lava that is blocky and highly fractured; these canals lose large quantities of water. Most of the leakage percolates to the water table and is a significant source of ground-water recharge in the irrigated parts of the basin (fig. 9).

Canal leakage was estimated for the 1994 irrigation season (May–September) using several sources of information, including: (1) diversions into canals measured at gaging stations operated by the OWRD, (2) estimates of irrigated acreage and crop-water applications from satellite imagery, (3) estimates of canal leakage rates from ponding experiments and surveys of canal-bottom geology by BOR (Bureau of Reclamation, 1991a, 1991b), and (4) estimates of irrigation efficiency by BOR (Bureau of Reclamation, 1993).

The 1994 canal leakage volume was calculated as the residual of the volume of water diverted into canals minus the volume of water delivered to farms. The areal distribution of canal leakage in the main canals and laterals was estimated on the basis of information on canal-bottom geology and ponding experiments.

To determine the on-farm deliveries from each canal in 1994, it was necessary to estimate the irrigated acres within each canal service area, the amount of water actually needed for the crops to grow (crop-water requirement), and the average irrigation efficiency within the canal service area. The actual crop-water application is equal to the crop-water requirement divided by the irrigation efficiency. For example, if the crop-water requirement were 2.0 ft/yr (feet per year) and the irrigation efficiency were 0.50, the crop-water application would be 4.0 ft/yr.

Satellite imagery was used to map 164,000 acres of irrigated croplands in the basin in 1994 and classify them according to the relative magnitude of crop-water requirements. The three classifications used were low, medium, and high water requirement crops. Of the total irrigated acreage, low water requirement crops made up 33,000 acres, medium water requirement crops made up 24,000 acres, and high water requirement crops made up 107,000 acres.

Water-rights information from the OWRD was used to determine that ground water was the source of irrigation to approximately 13,000 acres, with surface water supplying the remaining 151,000 acres.

The water requirement for each crop classification was estimated based on tables for the region (Cuenca and others, 1992; Bureau of Reclamation, 1995). County crop census data (Oregon State University, Extension Service, written commun., 1996) was used to weight the crop-water requirements to reflect the variability of crops grown in different parts of the basin. Climatic variability was accounted for by dividing the study area into northern and southern regions and applying appropriate crop-water requirements to irrigated lands in each region. The boundary between the regions coincides with the Deschutes–Jefferson County line (fig. 1). The low water requirement crop classification contained mostly fallow land; therefore, the water requirement was assumed to be zero for these areas. In 1994, medium water requirement crops were assumed to need 1.5 acre-feet per acre in the northern region and 1.7 ft in the southern region, while high water requirement crops were assumed to need 2.7 ft in the northern region and 2.4 ft in the southern region.

Irrigation efficiency depends primarily on the method used to apply the irrigation water. Sprinkler irrigation is the most efficient method and typically results in efficiencies of 75 to 90 percent. Flood irrigation is the least efficient and efficiencies of 35 to 50 percent are typical (U.S. Department of Agriculture, 1993). Irrigation efficiencies for each canal service area were estimated based on BOR studies in the basin (Bureau of Reclamation, 1993) and from interviews of local irrigation district and extension service personnel.

The total irrigation-water deliveries to farms within each canal service area, I_c, in acre-feet per year, were calculated:

$$I_c = (A_h \times C_h / E_c) + (A_m \times C_m / E_c)$$

where,

A_h and A_m are the areas of high and medium water-use crops, in acres,

C_h and C_m are the crop-water requirements for high and medium water-use crops, in feet per year, and

E_c is the average irrigation efficiency for the canal service area, in decimal percent.

Total 1994 diversions, irrigated acreage, on-farm deliveries, and canal leakage are listed for each major canal in table 4.

Canal leakage rates vary greatly within the study area depending on the geology of the canal bottom, the degree to which cracks and voids have been filled by sediment, and the wetted perimeter of the canal. The estimated total leakage within each canal service area (table 4) was apportioned among the canal and laterals on the basis of information available from studies by the BOR (Bureau of Reclamation, 1991a, 1991b, 1993). The BOR conducted ponding experiments in several canal reaches and determined leakage rates ranging from 0.64 to 4.20 $ft^3/d/ft^2$. This information was extrapolated using geologic mapping of the canal bottoms to estimate leakage rates for most of the main canals and laterals in the study area (fig. 9). The wetted area of each canal reach was calculated from the average width, depth, and length of the canal. Leakage rates were multiplied by wetted area to obtain estimates of leakage from each canal reach within a canal service area. If the total leakage did not match the total estimated as the residual of diversions minus on-farm deliveries, then the leakage rates were adjusted until the totals matched.

In 1994, 356,600 acre-ft, or 490 ft^3/s, leaked through canal bottoms to become ground-water recharge (table 4). This amounted to 46 percent of the 770,400 acre-ft (1,060 ft^3/s) diverted into canals in the upper Deschutes Basin. Canal leakage for the period 1905–97 was estimated for the basin assuming that the same proportion (46 percent) of diversions would be lost each year (fig. 10). Canal leakage peaked in the late 1950s when mean annual diversions were approximately 940,000 acre-ft (1,300 ft^3/s) and nearly 435,000 acre-ft (600 ft^3/s) was lost to ground-water recharge.

Figure 9 shows the distribution of canal leakage in the basin for 1993–95. The highest rates of leakage occur in reaches of the North Unit and Pilot Butte canals immediately east and north of Bend. In these reaches, canals are cut through highly fractured, blocky basalt and were estimated to lose an average of more than 20 $ft^3/s/mi$ (cubic feet per second per mile) during 1993–95.

Table 4. Canal diversions, irrigated acreage, on-farm deliveries, and canal leakage, by major canal service area, upper Deschutes Basin, Oregon, 1994

[All values in acre-feet unless otherwise noted; ft/yr, feet per year; --- not applicable.]

Canal	A Canal diver-sions	B Irrigated area[1] (acres)	C Mean crop-water requirement (ft/yr)	D Crop-water needs (B × C)	E Mean irrigation efficiency (percent)	F Estimated deliveries (D / E)	G Canal losses (A–F)
Arnold	26,570	2,310	2.25	5,200	0.50	10,400	16,170
Central Oregon	181,500	22,500	2.37	53,330	.43	124,020	57,480
North Unit	196,700	45,000	2.03	91,350	.94	97,180	99,520
Lone Pine	10,640	2,390	2.13	5,090	.89	5,720	4,920
Ochoco	75,000	16,600	2.12	35,190	.66	53,320	21,680
Peoples	6,500	1,540	2.21	3,400	.66	5,150	1,350
Pilot Butte	165,800	14,800	2.36	34,930	.43	81,230	84,570
Squaw Creek	26,400	5,450	1.50	8,180	.62	13,190	13,210
Tumalo	42,600	4,890	2.31	11,300	.60	18,830	23,770
Swalley	38,700	2,450	2.33	5,710	.51	11,200	27,500
Total	770,410	117,930	---	253,680	---	420,240	350,170
Average	---	---	2.15	---	.60	---	---

[1] Includes only high and medium water-use crops.

On-Farm Losses

Applied irrigation water can be lost to evaporation (from droplets, wetted canopy, soil and water surfaces), wind drift, runoff, and deep percolation. All of these losses are considered on-farm losses; however, the contribution of deep-percolation losses to ground-water recharge was the part of the loss of direct interest to this study. On-farm losses are directly correlated with irrigation efficiency. Irrigation efficiency is the ratio of the depth of irrigation water used by the plant to the depth of irrigation water applied, expressed as a percentage. As shown in table 4, estimated mean irrigation efficiencies in the study area vary from 43 percent in areas where flooding is the primary method of application to 94 percent where sprinklers are the primary method.

Literature values were used to estimate losses to evaporation, wind drift, and runoff. The percentage of applied irrigation water lost to these sources is highly variable and dependent on individual water-management practices and soil and climatic conditions. A maximum of 20 percent was assumed to be lost to these sources throughout the study area (U.S. Department of Agriculture, 1993). For example, where the irrigation efficiency is 60 percent (60 percent of the applied water is used by the plant), of the remaining 40 percent of applied water, 20 percent is assumed to be lost to evaporation, wind drift, and runoff, while 20 percent is assumed to be lost to deep percolation. In areas of sprinkler irrigation with efficiencies of 94 percent, only 6 percent of applied water is lost (mostly to evaporation and wind drift), and no water is assumed to be lost to deep percolation.

Mean annual recharge (1993–95) from deep percolation of on-farm losses was only about 49,000 acre-ft (68 ft^3/s) (fig. 9). The service area for the North Unit canal is almost entirely irrigated by sprinkler; therefore, no recharge from on-farm losses were estimated in this area. In other areas, where a mixture of flood and sprinkler irrigation is used, up to 5 in./yr of recharge occurs from on-farm losses. Areas where flood irrigation is the predominant irrigation method receive recharge of up to 10 in./yr from on-farm losses.

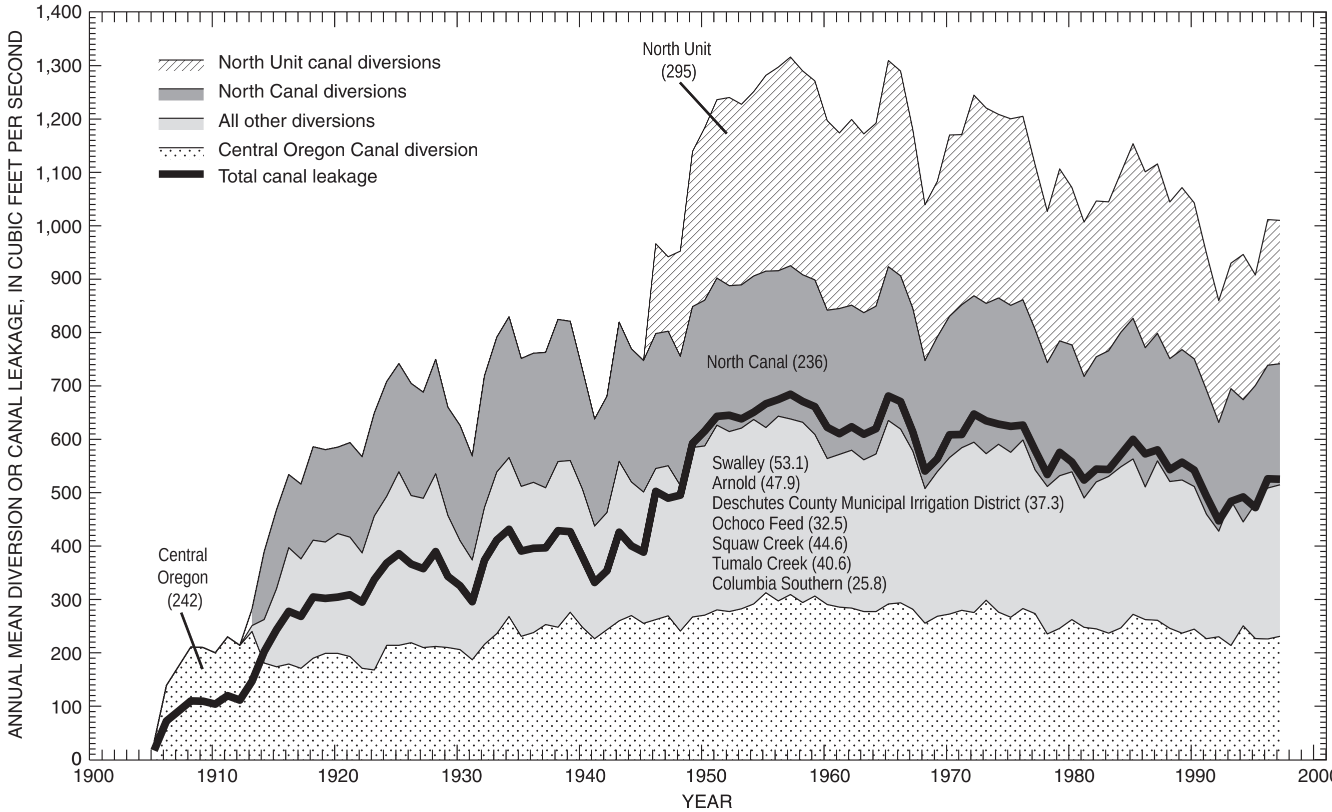


Figure 10. Annual canal diversions and estimated annual mean canal leakage in the upper Deschutes Basin, Oregon, 1905–97. (Mean annual discharge, in cubic feet per second, is shown in parentheses for the period of record for each diversion.)

Stream Leakage

Where the elevation of a stream is above that of the water table in adjacent aquifers, water can leak from the stream to the underlying strata and recharge the ground-water system. Such streams are termed *losing streams*. Conversely, in areas where the stream elevation is below that of adjacent aquifers, ground water can discharge to streams, increasing streamflow. Such streams are termed *gaining streams*.

In this study, ground-water flow from and to streams was estimated using data from a variety of sources. The primary sources of information were sets of streamflow measurements known as *seepage runs*. A seepage run consists of a series of streamflow measurements taken a few to several miles apart along a stream over a short enough period that temporal variations in streamflow are minimal. Tributary inflow and diversions are measured as well. Any temporal changes in streamflow occurring during the measurement period also are measured or otherwise accounted for. Seepage runs provide a snapshot of the rate and distribution of ground-water inflow to, or leakage from, a stream; single seepage runs, however, do not provide information on temporal variations in stream gains and losses. Seepage runs were conducted along all major streams in the upper Deschutes Basin by OWRD, and multiple runs were conducted on certain streams. Data from the seepage runs were provided by Kyle Gorman, OWRD (written commun., 1994, 1995, 1996) and are presented in table 5.

The methods used to measure streamflow have an inherent error of plus or minus 5 percent under good measurement conditions. Therefore, streamflow variations of less than 5 percent measured between two points during a seepage run may represent measurement error and not an actual gain or loss. However, if the sum of such small gains or losses along a reach exceeds the likely measurement error, it is reasonable to assume there is an actual gain or loss.

Data from stream-gaging stations also were useful in estimating the amount of ground water discharging to or leaking from streams. Because stream gages operate continuously, they can provide information on temporal changes in gains and losses.

Most stream-gage data used in this section and the following section on ground-water discharge were from the USGS National Water Information System (NWIS). Additional data were obtained from published compilations (U.S. Geological Survey, 1958; Oregon Water Resources Department, 1965). The locations of gaging stations used in this report are shown in figure 11, and the station numbers and names are listed in table 6. Some statistical summaries were taken from Moffatt and others (1990). Data from OWRD gages and irrigation diversions were provided by the OWRD (Kyle Gorman, written commun., 1998, 1999, 2000). Estimated stream gains and losses are presented in table 7 and shown graphically along with selected stream-gage locations in figure 12. Unless otherwise noted, the gain and loss rates in table 7 are assumed to represent average conditions.

In the upper Deschutes Basin, losing streams are much less common than gaining streams (fig. 12). The conditions required for losing streams, a water-table elevation below the stream elevation, occur much less commonly than the conditions required for gaining streams.

The rates of water loss from losing streams are usually much less than the rates of ground-water inflow to gaining reaches (fig. 12) because of differences in the ways water enters and leaves streams. In the upper Deschutes Basin, water typically enters streams from springs issuing from highly fractured lava or coarse sedimentary deposits like sands and gravels. These springs commonly occur above river level (Ferns and others, 1996), and there is no mechanism by which the fractures or other openings through which the water emerges can be effectively blocked. The fractures and openings through which water leaks from losing streams, in contrast, are much more easily blocked and sealed. Streams typically carry sediment suspended in the water column and along the bottom. Over long periods of time, these materials can infiltrate the openings and essentially seal them, greatly reducing the permeability of the streambed. This process is likely particularly important in streams, such as those in most of the Deschutes Basin, that flow in canyons and do not meander and, therefore, do not periodically establish new channels. Irrigation canals lose more water than streams over a given length. This is because canals are much younger features and have had much less time to be sealed by sediment, and possibly because canal water typically carries very little suspended sediment. Even though the amount of water lost from streams to the ground-water system is only a fraction of the amount that flows from the ground-water system to streams, stream leakage is still an important source of recharge in certain areas.

Leakage from streams, lakes, and reservoirs recharges the ground-water system in some areas in the southern part of the basin. Some of the high lakes, such as Hosmer Lake and Elk Lake (fig. 1) are essentially ground-water fed, and their leakage represents little, if any, net ground-water recharge. Others, such as Sparks and Devils Lakes, are fed at least in part by perennial streams. The net ground-water recharge from these lakes is unknown, but much of it likely emerges as springflow in the Deschutes River and tributaries above Crane Prairie Reservoir.

Crane Prairie Reservoir also loses water through leakage to the ground-water system. This is the only reservoir in the southern part of the basin for which sufficient gages have been operated to allow a good estimate of seepage losses. The average loss from Crane Prairie Reservoir between 1939 and 1950 was computed to be 60,000 acre-ft/yr, or about 83 ft^3/s (U.S. Geological Survey, 1958). A more detailed analysis indicated that the leakage ranges from about 30 to 135 ft^3/s, depending on the stage of the reservoir (Robert F. Main, OWRD, written commun., 1999). Some of this loss probably returns to the Deschutes River through springs within about 3 or 4 miles below Crane Prairie Dam, along what is now an arm of Wickiup Reservoir. It is probable, however, that some of this water contributes to the regional ground-water flow system.

The water budget of Wickiup Reservoir is not as well understood as that of Crane Prairie Reservoir. Although the major streams entering Wickiup Reservoir are gaged, there is substantial spring flow into the western parts of the reservoir along the Deschutes River and Davis Creek. A comparison of annual mean gaged inflow and outflow from Wickiup Reservoir from 1939 to 1991 showed that annual mean net spring flow into the reservoir from the west ranged from 308 to 730 ft^3/s and averaged 486 ft^3/s. This value does not include evaporation, which is considered negligible. This inflow rate varies with climatic conditions and apparently with the stage-dependent losses from Crane Prairie Reservoir (Bellinger, 1994). Although there is net inflow to the reservoir, there is seepage from the reservoir as well. Sinkholes develop periodically, into which large amounts of water drain.

Table 5. Gain/loss measurements of major streams obtained from Oregon Water Resources Department seepage runs, upper Deschutes Basin, Oregon

[Q, discharge in cubic feet per second at the measurement site; Trib Q, discharge in cubic feet per second of a tributary entering the stream of interest; Seepage, calculated seepage in cubic feet per second between the measurement site and the upstream measurement site (corrected for tributary inflow), negative value indicates a loss of water to the aquifer; USGS, U.S. Geological Survey; RM, river mile]

Site	River mile	Date	Q	Trib. Q	Seepage	Date	Q	Trib. Q	Seepage	Comments
					Little Deschutes River					
Cow Camp	86.0	10/10/95	19.7	--	--	--	--	--	--	
Highway 58	80.0	10/10/95	31.5	--	11.8	--	--	--	--	
Forest Service Road	71.0	10/10/95	28.9	--	-2.6	--	--	--	--	
Above Gilchrist Pond	64.5	10/10/95	26.9	--	-2.0	10/9/96	32.4	--	--	
Above Crescent Creek	57.5	10/10/95	15.9	--	-11.0	10/9/96	18.0	--	-14.4	
Wagon Trail Ranch	43.0	10/10/95	50.4	32.2	2.3	--	--	--	--	Tributary: Crescent Creek
Road Crossing (off 6th St. West of La Pine)	33.5	10/10/95	53.3	--	2.9	--	--	--	--	
Little Deschutes near La Pine (USGS gaging station 14063000)	26.7	10/10/95	53.3	--	0.0	10/9/96	54.7	--	--	
State Recreation Road	13.7	10/10/95	55.9	--	2.6	10/9/96	57.7	--	3.0	
South Century Drive	5.5	10/10/95	59.2	--	3.3	10/9/96	59.9	--	2.2	
Above Mouth at Crosswater	1.4	10/10/95	57.4	--	-1.8	10/9/96	57.0	--	-2.9	
					Crescent Creek					
Below Crescent Lake	30	10/10/95	5	--	--	--	--	--	--	
Highway 58 crossing	18.5	10/10/95	33.7	10	18.7	--	--	--	--	Tributary Q estimated
Above Mouth	2.2	10/10/95	32.2	--	-1.5	--	--	--	--	
					Fall River					
Near headwaters	9.0	2/21/96	18.3	--	18.3	--	--	--	--	
Near La Pine (USGS gaging station 14057500)	5.0	2/21/96	119	--	100.7	--	--	--	--	
Near mouth	0.4	2/21/96	112	--	-7.0	--	--	--	--	
					Upper Deschutes River					
Below Wickiup Dam (USGS gaging station 14056500)	226.7	2/21/96	128	--	--	--	--	--	--	
Pringle Falls	217.6	2/21/96	126	--	-2.0	--	--	--	--	
La Pine State Park	208.6	2/21/96	132	--	6.0	--	--	--	--	
General Patch Bridge	199.7	2/21/96	268	112	24.0	--	--	--	--	
Harper’s Bridge	191.7	2/21/96	758	544	-54.0	--	--	--	--	Tributary Q measured at RM 26.8
Benham Falls (USGS gaging station 14064500)	181.4	2/21/96	1,047	--	289.0	--	--	--	--	Includes inflow from Spring River
					Indian Ford Creek					
Black Butte Ranch Weir	10.7	2/5/92	5.86	--	5.86	4/23/92	6.42	--	6.42	
Willows Ranch Diversion	8.0	2/5/92	4.31	--	-1.55	4/23/92	4.62	--	-1.8	
Willows Ranch	3.6	2/5/92	4.43	--	0.1	4/23/92	3.15	--	-1.5	
Camp Polk Road	2.1	2/5/92	3.02	--	-1.4	4/23/92	2.94	--	-0.2	
Sisters Airport	1.3	2/5/92	0.01	--	-3.0	4/23/92	0.47	--	-2.5	
Barclay Road	0.8	2/5/92	0.00	--	0.0	4/23/92	0.00	--	-0.5	
Confluence with Squaw Creek	0.0	2/5/92	0.00	--	0.0	4/23/92	0.00	--	0.0	

Table 5. Gain/loss measurements of major streams obtained from Oregon Water Resources Department seepage runs, upper Deschutes Basin, Oregon–Continued

[Q, discharge in cubic feet per second at the measurement site; Trib Q, discharge in cubic feet per second of a tributary entering the stream of interest; Seepage, calculated seepage in cubic feet per second between the measurement site and the upstream measurement site (corrected for tributary inflow), negative value indicates a loss of water to the aquifer; USGS, U.S. Geological Survey; RM, river mile]

Site	River mile	Date	Q	Trib. Q	Seepage	Date	Q	Trib. Q	Seepage	Comments
Lower Squaw Creek										
Brooks-Scanlon Logging Road	22	4/13/94	0.00	--	--	--	--	--	--	
Near Highway 20 Crossing	19	4/13/94	0.00	--	--	--	--	--	--	
Camp Polk Road Crossing	16.5	4/13/94	6.61	--	6.6	--	--	--	--	
Near Henkle Butte	14.7	4/13/94	7.30	--	0.7	--	--	--	--	
Below Squaw Creek Estates	9.9	4/13/94	6.46	--	-0.8	--	--	--	--	
Forest Road 6360 Crossing	5.5	4/13/94	6.61	--	0.2	--	--	--	--	
Above Alder Springs	1.7	4/13/94	7.46	--	0.9	--	--	--	--	
Below Alder Springs	1.05	4/13/94	26.50	--	19.0	--	--	--	--	
Middle Deschutes River										
Deschutes below Bend (USGS gaging station 14070500)	164.3	5/11/92	29.8	--	--	5/16/94	29.0	--	--	
Below mouth of Tumalo Creek	160.2	5/11/92	37.1	7.3	0.0	5/16/94	31.2	2.2	0.0	Tributary Q determined from gain
Tumalo State Park	158.8	5/11/92	39.7	--	2.6	5/16/94	42.3	--	11.1	
Deschutes River Ranch	154.5	5/11/92	43.8	--	4.1	5/16/94	43.7	--	1.4	
Above Eagle Crest	146.8	5/11/92	--	--	--	5/16/94	34.8	--	-8.9	
Cline Falls State Park	145.3	5/11/92	56.0	--	12.2	5/16/94	45.1	--	10.3	
Crestridge Estates	143.2	5/11/92	--	--	--	5/16/94	35.2	--	-9.9	
Tethero Road crossing	141.2	5/11/92	32.4	--	-23.6	5/16/94	36.8	--	1.6	
Odin Falls Ranch	138.0	5/11/92	--	--	--	5/16/94	35.5	--	-1.3	
Below Odin Falls Ranch	137.5	5/11/92	--	--	--	5/16/94	44.4	--	8.9	
Lower Bridge Road crossing	134.0	5/11/92	52.2	--	19.8	5/16/94	46.0	--	1.6	
NW Riffle Road	130.5	5/11/92	67.7	--	15.5	5/16/94	61.4	--	15.4	
Below fishing point near mine cabin	128.7	5/11/92	216	--	148.3	5/16/94	217	--	155.6	
Half mile below Steelhead Falls	127.2	5/11/92	223	--	7.0	5/16/94	--	--	--	
River Place below pump house	126.1	5/11/92	207	--	-16.0	5/16/94	--	--	--	
Sundown Canyon Road	124.9	5/11/92	226	--	19.0	5/16/94	213	--	-4	
Scout Camp Trail Road	123.3	5/11/92	365	--	139.0	5/16/94	341	--	128.0	
Below mouth of Squaw Creek	123.0	5/11/92	468	103	0.0	5/16/94	442	101	0.0	Tributary Q determined from gain
Deschutes River near Culver (USGS gaging station 14076500)	120.0	5/11/92	480	--	12.0	5/16/94	467	--	25.0	
Lower Crooked River										
Crooked River near Terrebonne	27.6	6/22/94	10.5	--	--	--	--	--	--	
Trail Crossing	20.4	6/22/94	23.6	--	13.1	--	--	--	--	
At Osborne Canyon	13.8	6/22/94	93.5	--	69.9	--	--	--	--	
Crooked River below Opal Springs (USGS gaging station 14087400)	6.7	6/22/94	1,100	--	1,006.5	--	--	--	--	

Table 6.Station numbers, names, and mean annual flow for selected gaging stations in the upper Deschutes Basin, Oregon
[All data are from Moffatt and others (1990) unless noted; OWRD, Oregon Water Resources Department]

Station number	Station name	Mean annual flow	Period of record
14050000	Deschutes River below Snow Creek, near La Pine	151	1938 to 1987
14050500	Cultus River above Cultus Creek, near La Pine	63	1923 to 1987
14051000	Cultus Creek above Crane Prairie Reservoir, near La Pine	22	1924 to 1962
14052000	Deer Creek above Crane Prairie Reservoir, near La Pine	7.5	1924 to 1987
14052500	Quinn River near La Pine	24	1938 to 1987
14054500	Browns Creek near La Pine	38	1923 to 1987
14055100	Davis Creek (OWRD gage data)[1]	191	1939 to 1942
14055500	Odell Creek near Crescent	82	1913 to 1976
14055600	Odell Creek (OWRD gage data, gage several miles downstream of gage 14055500)[2]	126	1970 to 1990
14056500	Deschutes River below Wickiup Reservoir, near La Pine	754	1943 to 1987
14057500	Fall River near La Pine	150	1938 to 1987
14061000	Big Marsh Creek near Hoey Ranch, near Crescent	72	1912 to 1958
14063000	Little Deschutes River near La Pine	208	1924 to 1987
14063800	Deschutes River at Peters Ranch (OWRD gage data)[1]	1,210	1944 to 1953
14064000	Deschutes River at Camp Abbott Bridge (OWRD gage data)[1]	1,478	1944 to 1953
14064500	Deschutes River at Benham Falls, near Bend	1,480	1944 to 1987
14066000	Deschutes River below Lava Island, near Bend	1,380	1943 to 1965
14070500	Deschutes River below Bend	377	1957 to 1987
14073001	Tumalo Creek near Bend	101	1924 to 1987
14075000	Squaw Creek near Sisters	105	1906 to 1987
14076500	Deschutes River near Culver	929	1953 to 1987
14087400	Crooked River below Opal Springs, near Culver	1,610	1962 to 1987
14087500	Crooked River near Culver	1,560	1920 to 1960
14088000	Lake Creek near Sisters	52	1918 to 1987
14088500	Metolius River at Allingham Ranger Station, near Sisters[3]	376	1911 to 1912
14090350	Jefferson Creek near Camp Sherman[4]	94.9	1984 to 1999
14090400	Whitewater River near Camp Sherman[4]	86.6	1983 to 1999
14091500	Metolius River near Grandview	1,500	1912 to 1987
14092500	Deschutes River near Madras	4,750	1964 to 1987

[1] Oregon Water Resources Department (1965).
[2] Kyle Gorman, OWRD, written commun. (1999).
[3] U.S. Geological Survey (1958).
[4] Hubbard and others (2000).

Table 7. Estimated stream gains and losses due to ground-water exchange, upper Deschutes Basin, Oregon

[RM, river mile; ft^3/s, cubic feet per second; OWRD, Oregon Water Resources Department; USGS, U.S. Geological Survey; Res., reservoir; NW, northwest; OSU, Oregon State University]

Stream Name	Reach (river mile)	Estimated gain (+) or loss (-) (ft^3/s)	Data source	Remarks
Little Deschutes River	Entire drainage above Highway 58, RM 80	31.5	OWRD measurements 10/95	Includes Hemlock Creek
Little Deschutes River	Highway 58 to above Crescent Creek, RM 80 to RM 57.5	-15.6	OWRD measurements 10/95	
Little Deschutes River	Above Crescent Creek to Crosswater, RM 57.5 to RM 1.4	9.3	OWRD measurements 10/95	
Big Marsh Creek	Drainage above gage near Mouth, RM 0.5	21	Statistical summary of gage 14061000 (Moffatt and others, 1990)	Mean September discharge, 1924–87
Crescent Creek	Crescent Lake outlet to Highway 58, RM 30 to RM 18.5	18.7	OWRD measurements 10/95	Includes inflow from Cold Springs Creek
Crescent Creek	Highway 58 to above mouth, RM 18.5 to RM 2.2	-1.5	OWRD measurements 10/95	
Paulina Creek	Paulina Lake outlet to Road 21, RM 13 to RM 5.2	-1.7 to -6.1	Morgan and others (1997)	
Odell Lake	Above gage at lake outlet	41	Statistical summary of gage 14055500 (Moffatt and others, 1990)	Mean September discharge, 1913–76
Odell Creek	Odell Lake outlet to OWRD gage	41	USGS and OWRD gage data for stations 14055500 and 14055600	Period of overlapping records from 1970 to 1976
Davis Creek	Upstream from RM 3	191	OWRD gage data for station 14055100 (1938–43), OWRD miscellaneous measurements 1978–94	Entire flow from springs
Cultus River	Above Cultus Creek	63	Statistical summary of gage 14050500 (Moffatt and others 1990)	Mean annual flow of spring-fed stream
Deschutes River	Above Crane Prairie Res.	151	Statistical summary of gage 14050000 (Moffatt and others, 1990)	Mean annual flow of spring-fed stream
Deschutes River	Crane Prairie Res. to Wickiup Res.	229	USGS Water-Supply Paper 1318	Based on gage data 1926–32
Deschutes River	Wickiup Res. to La Pine State Park, RM 226.7 to RM 208.6	4	OWRD measurements 2/96	
Fall River	Headwaters springs to gage at RM 5	119	OWRD measurements 2/96	Entire flow from springs
Fall River	Gage to mouth, RM 5 to RM 0.4	-7	OWRD measurements 2/96	
Deschutes River	Near Fall River, RM 208.6 to RM 199.7	24	OWRD measurements 2/96	

Table 7. Estimated stream gains and losses due to ground-water exchange, upper Deschutes Basin, Oregon—Continued

[RM, river mile; ft^3/s, cubic feet per second; OWRD, Oregon Water Resources Department; USGS, U.S. Geological Survey; Res., reservoir; NW, northwest; OSU, Oregon State University]

Stream Name	Reach (river mile)	Estimated gain (+) or loss (-) (ft^3/s)	Data source	Remarks
Deschutes River	Peters Ranch to Camp Abbott Bridge, RM 191.5 to RM 189	271	OWRD gage data for stations 14063800 and 14064000	Period of record 1945 to 1953, includes flow of Spring River
Deschutes River	Camp Abbott Bridge to Benham Falls, RM 189 to RM 181.4	-24	OWRD and USGS gage data for stations 14064000 and 14064500	Period of record 1945 to 1953
Deschutes River	Benham Falls to below Bend, RM 181.4 to 164.4	-89	USGS and OWRD gage data for the river and diversions	Period of record 1945 to 1995
Tumalo Creek	Entire drainage above gage at RM 3.1	68	Statistical summary of gage 14073001 (Moffatt and others, 1990)	Mean August–September discharge, 1924 to 1987, does not include City of Bend diversion
Tumalo Creek	RM 2.3 to RM 0.3	-0.24	OWRD measurements 8/94	Measurements made at very low flow conditions
Deschutes River	Below Bend to Odin Falls Ranch, RM 164.4 to RM 138.0	6.5	OWRD measurements 5/94	Gain may decrease during summer and fall, may be irrigation return flow
Deschutes River	Odin Falls Ranch to NW Riffle Road, RM 138.0 to RM 130.5	17.0	OWRD measurements 5/94	Gain may be irrigation return flow
Deschutes River	Riffle Road to Culver gage, RM 130.5 to RM 120.0	305	OWRD measurements 5/94	
Indian Ford Creek	Upstream from RM 10.7 at Black Butte Ranch	6.1	OWRD measurements 2/94–4/94	
Indian Ford Creek	Black Butte Ranch to mouth, RM 10.7 to RM 0.8	-6.1	OWRD measurements 2/94–4/94	Stream dry at RM 0.8
Squaw Creek	Entire drainage above gage at RM 26.8	65	Statistical summary of gage 14075000 (Moffatt and others 1990)	Mean September discharge 1906 to 1987, may be influenced by glacial melt
Squaw Creek	Near McKinney Butte, RM 19 to RM 16.5	6.6	OWRD measurements 4/94	Discharge from springs
Squaw Creek	McKinney Butte to Alder Springs, RM 16.5 to RM 1.7	0.85	OWRD measurements 4/94	
Squaw Creek	Alder Springs to confluence RM 1.7 to RM 0	94	OWRD measurements 4/94 and 5/94 (at mouth)	
Crooked River	Terrebonne to Trail Crossing, RM 27.6 to 20.4	13.1	OWRD measurements 6/94	
Crooked River	Trail Crossing to Osborne Canyon, RM 20.4 to 13.8	70	OWRD measurements 6/94	

Table 7. Estimated stream gains and losses due to ground-water exchange, upper Deschutes Basin, Oregon—Continued

[RM, river mile; ft^3/s, cubic feet per second; OWRD, Oregon Water Resources Department; USGS, U.S. Geological Survey; Res., reservoir; NW, northwest; OSU, Oregon State University]

Stream Name	Reach (river mile)	Estimated gain (+) or loss (-) (ft^3/s)	Data source	Remarks
Crooked River	Osborne Canyon to Opal Springs gage, RM 13.8 to RM 6.7	1,006	OWRD measurements 6/94	
Metolius River	Headwaters to Allingham Ranger Station (gage site) at RM 38.1	257	Alexander and others (1987, p. 215); data for gage 14088500 (USGS Water-Supply Paper 1318)	Estimated by using data from different time periods: 9/85 measurements and gage data from 1911 to 1917.
Suttle Lake	Lake above gage at outlet	31	Statistical summary of gage 14088000 (Moffatt and others, 1990)	Mean September flow, 1918 to 1987
Lake Creek	Suttle Lake to mouth	36	Alexander and others (1987, p. 215) and Lake Creek gage data	9/85 measurements, includes about 0.5 mile of Metolius River
First Creek	Upper part of drainage	1.5	OWRD measurements 1992–94	
Jack Creek	Upper part of drainage	46	OWRD measurements 1992–94	
Canyon Creek	Entire drainage above about RM 0.5	60	OWRD measurements 1992–94	
Abbot Creek	Entire drainage above about RM 1.5	12	OWRD measurements 1992–94	
Candle Creek	Entire drainage above about RM 2	73	OWRD measurements 1992–94	
Metolius River	Allingham Ranger Station (RM 38.1) to gage near Grandview (RM 13.6) including lower parts of tributary drainages	724	OWRD measurements 1992–94, USGS gage data	Gain based on OWRD measurements and mean August discharges for gages. Most inflow occurs above Jefferson Creek (RM 28.8)
Lake Billy Chinook	Reservoir encompassing confluence of the Deschutes, Crooked, and Metolius Rivers	379–461	USGS gage data and Bolke and Laenen (1989)	Figure uncertain due to bank storage effects. Evaporation is estimated at 20 to 25 ft^3/s based on 1997 evaporation data from OSU Agricultural Experiment Station at Madras.
Deschutes River	Round Butte Dam to Lake Simtustus, RM 109.9 to RM 108.5	200	Bolke and Laenen (1989)	
Lake Simtustus	Reservoir on the Deschutes River extending from RM 108.5 to RM 102.8	51	Bolke and Laenen (1989)	Figure does not account for evaporation
Deschutes River	Pelton Dam to Regulator Dam, RM 102.8 to RM 100.1	80	Bolke and Laenen (1989)	
Deschutes River	Regulator Dam to below Campbell Creek, RM 100.1 to RM 97.2	53	Bolke and Laenen (1989)	
Deschutes River	Below Campbell Creek to Dry Creek, RM 97.2 to RM 91.8	28	Bolke and Laenen (1989)	No significant gains from this point downstream in study area

Sinkholes apparently have been less of a problem since the early 1960s (Bellinger, 1994). The average rate of seepage from Wickiup Reservoir is unknown, but it is probably not more than a few tens of cubic feet per second.

Seepage runs indicate some losses along the Little Deschutes River as it flows through the La Pine subbasin (table 5). Most of the measured losses are small, 1 to 3 ft^3/s, and are within the measurement error of the 30 to 60 ft^3/s streamflow rates. Measured losses between Gilchrist and Crescent Creek, ranging from 11 to 14.4 ft^3/s, are sufficiently large with respect to measurement error to be considered meaningful. The Little Deschutes River crosses lava flows of Crescent Butte Volcano along this reach and it is likely that water is being lost into permeable lava. Much of this water likely returns to the river in gaining reaches not far downstream. A seepage run on Crescent Creek, a tributary to the Little Deschutes River, indicated a 1.5 ft^3/s loss in the lower 18 miles. This loss is small compared to the flow, approximately 33 ft^3/s, and is within the measurement error.

Paulina Creek, a tributary to the Little Deschutes River that flows down the west flank of Newberry Volcano, had measured net losses of approximately 2 to 6 ft^3/s between river mile 13, at its source at the outlet of Paulina Lake, and river mile 5.2, where it flows onto the floor of the La Pine subbasin (Morgan and others, 1997). This loss accounted for roughly 20 to 40 percent of the flow of Paulina Creek at the times the seepage runs were made.

Seepage runs indicate that, with the exception of the reservoirs discussed previously, the Deschutes River has no significant losing reaches upstream of its confluence with the Little Deschutes River. Downstream from the confluence, gaging-station data indicate significant losses occur along the reach extending from the community of Sunriver downstream to Bend. Comparison of flow measured at a gage operated from 1945 to 1953 at the Camp Abbott Bridge with the flow at the Benham Falls gage about 10 miles downstream indicates that this reach of the river lost an average of about 24 ft^3/s during that period (Oregon Water Resources Department, 1965). The loss, as calculated using monthly mean flow, is variable and weakly correlated with flow (correlation coefficient = 0.40).

The Deschutes River loses an average 83 ft^3/s between Benham Falls and the gage site below Lava Island about 7.5 miles downstream, based on the period of record from 1945 to 1965. The loss in flow along this reach ranged from −10 ft^3/s (a slight gain) to 236 ft^3/s and is fairly well correlated with flow (correlation coefficient = 0.74) (fig. 13).

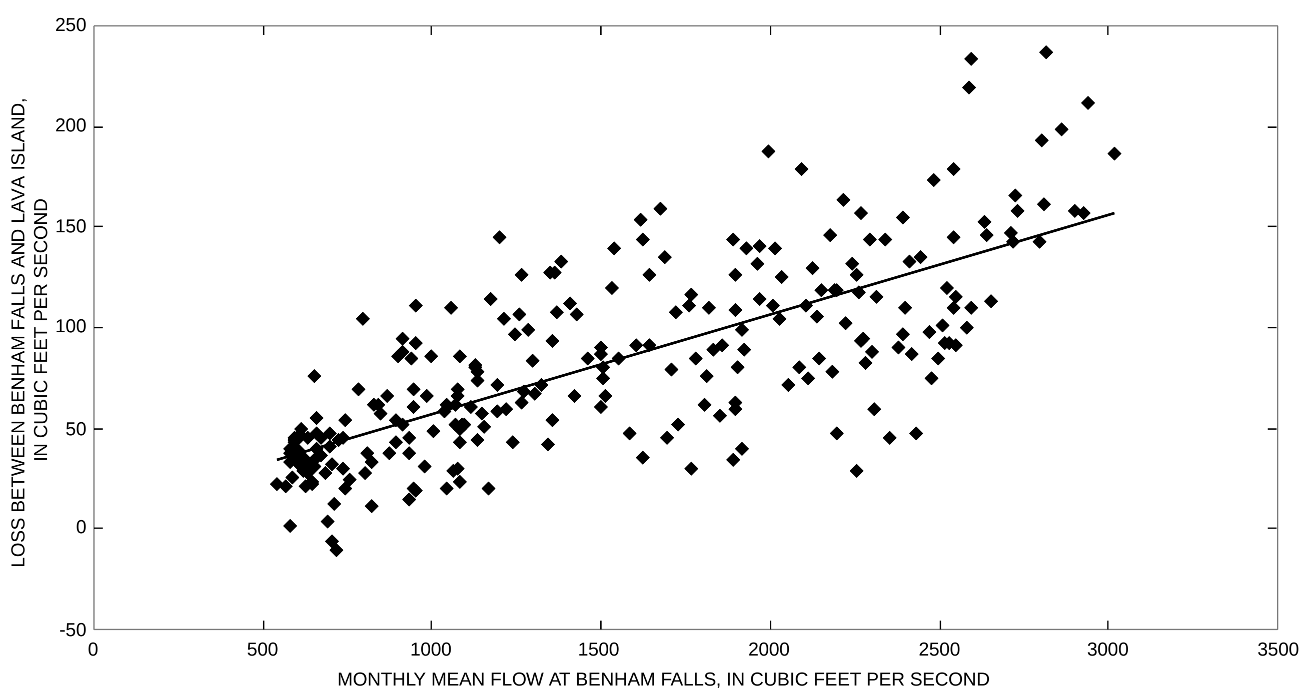


Figure 13. Relation between monthly mean losses along the Deschutes River between Benham Falls and Lava Island and flow at Benham Falls.

The wide range of these values is likely due to measurement error of the stream gages and of the gage on a diversion used in the loss calculation. The rate of leakage in this reach far exceeds that of any other losing stream reach in the upper Deschutes Basin. The leakage in this area is likely into very young, highly permeable lava flows from Lava Butte that diverted the river and now form much of the east bank and some of the falls along this reach. Stream losses between Camp Abbott Bridge and Lava Island far exceed losses anywhere else in the upper Deschutes Basin and are an important source of recharge.

USGS and OWRD stream-gage data from 1945 to 1965 indicate that average stream losses between the gage below Lava Island and the gage below Bend are small, about 4.0 ft^3/s. The difference in flow along this reach ranged from a 68 ft^3/s gain to a 72 ft^3/s loss, and shows no correlation with flow. The wide range in values is likely due to measurement error of the stream gages and of the gages on five diversions used in the calculations.

Calculated losses along the two reaches of the Deschutes River described above, which total 87 ft^3/s, are based on a period of record from 1945 to 1965. Losses along the two separate reaches after 1965 cannot be calculated because the gage below Lava Island ceased operation. Losses can be calculated, however, for the entire reach from Benham Falls to Bend for a much longer period. The average loss between Benham Falls and Bend, based on monthly mean flows from 1945 to 1995, is 89 ft^3/s. This agrees favorably with the sum of losses calculated for the subreaches for the shorter period of record.

Information on stream losses along the Deschutes River from Bend downstream to Lower Bridge is from OWRD seepage runs (Kyle Gorman, OWRD, written commun., 1995) (table 5); gage data are insufficient for evaluating losses along this reach. Seepage runs indicate that there are two areas between Bend and Lower Bridge where the Deschutes may lose a small amount of water (table 5). These areas are between river miles 154.5 and 146.8, near Awbrey Falls, and between river miles 145.3 and 143.2, near Cline Falls. Losses in both these areas are about 10 ft^3/s, and were measured when flows ranged from 30 to 50 ft^3/s. Not far downstream from both of these losing reaches, the river gains comparable amounts of water, implying that water lost from the river along this section apparently returns to the surface not far downstream. These seepage runs were done during periods of very low streamflow and may not reflect losses at higher flow rates. However, gage data from upstream between Lava Island and Bend suggest that losses may not be flow dependent along this reach. There are no significant losses from the Deschutes River downstream of Lower Bridge.

Stream losses also were measured along Indian Ford Creek (table 5). A series of seepage measurements taken by OWRD during the winter months of 1992 indicate that Indian Ford Creek lost its entire flow (approximately 6 ft^3/s) between the Black Butte Ranch springs, where it originates, and its confluence with Squaw Creek.

No other streams measured in the upper Deschutes Basin showed significant losses. The lower sections of Tumalo and Squaw Creeks showed only minor losses of less than 1 ft^3/s when measured during low flow conditions. Possible losses during higher flow conditions are not known.

Drainage Wells

Storm runoff in urban areas of the upper Deschutes Basin is often disposed of through drainage wells. Drainage wells in this report include both drilled disposal wells and larger diameter, but shallower, drywells, which are usually dug. Runoff disposed of in drainage wells is routed directly to permeable rock beneath the land surface, bypassing the soil zone from which a certain amount of the water would normally be returned to the atmosphere through evaporation or transpiration by plants. Once routed to permeable rock beneath the soil, the runoff percolates downward to recharge the ground-water system.

Although runoff disposed of through drainage wells represents a source of ground-water recharge, the volume of water is very small relative to other sources of recharge in urban areas, such as canal leakage, and minuscule compared to the entire ground-water flow budget. To illustrate this, estimates of the amount of ground-water recharge through drainage wells in Bend and Redmond are presented in this section.

Engineering maps provided by the city of Bend in 1994 show approximately 1,175 drainage wells used for street drains in the city. This number does not include drainage wells on private property, but their distribution is taken to represent the area over which runoff is handled in this manner. There are 163 quarter-quarter sections (40-acre tracts) with at least 1 and as many as 30 drainage wells. The quarter-quarter sections with at least one drainage well compose a total area of just over 10 mi^2. To estimate the amount of ground-water recharge from drainage wells, it is necessary to estimate the fraction of the total precipitation that is routed to them.

Runoff routed to drainage wells is that which falls on impervious surfaces and cannot infiltrate the soil naturally. Roofs, driveways, parking lots, and streets are examples of impervious surfaces. The amount of impervious surface relative to the total land area varies with land-use type. Commercial areas, with large roofed structures and expansive parking lots, can be 85 percent impervious (Snyder and others, 1994). Impervious surfaces in residential areas, in contrast, range from 20 percent of the land area, for large lots where yards are big relative to structures and driveways, to 65 percent for small lots (Soil Conservation Service, 1975). A value of 35 percent impervious surface was used for calculations for Bend, based on mapped impervious areas for dominantly residential areas in Portland, Oregon, and Vancouver, Washington (Laenen, 1980, table 1).

Not all of the precipitation that falls on impervious surfaces runs off to drainage wells. A certain amount is evaporated from wetted surfaces and undrained areas such as puddles, and from detention structures. This is known as detention-storage loss. In estimating recharge from drainage wells in the Portland Basin, Snyder and others (1994), using the work of Laenen (1980), estimated that about 25 percent of the precipitation was evaporated in this manner, leaving about 75 percent to run off to drainage wells. Because this value was derived using conditions in western Oregon, it may be low for the Bend area, where conditions are much dryer. A detention-storage loss of 25 percent is used herein with the assumption that if it is too conservative, recharge from drainage wells may be slightly overestimated.

Average recharge from drainage wells in Bend was estimated assuming that runoff from all impervious surfaces in any quarter-quarter section (40-acre tract) with at least one drainage well was disposed of through drainage wells. There are 163 quarter-quarter sections meeting this criteria, with an aggregate area of 10.19 mi^2. Average precipitation in Bend is 11.70 in./yr (period of record 1961 to 1990) (Oregon Climate Service, 1999). Using these figures and assuming that 35 percent of the area is impervious surface and that 25 percent of the precipitation is lost through evaporation, the runoff routed to dry wells is approximately 73 million ft^3/yr, or about 2.3 ft^3/s. This is not a significant source of recharge when compared to canal and stream leakage, which can exceed 20 ft^3/s/mi near Bend.

Similar calculations were done for Redmond using maps provided by the city and aerial photographs taken in 1995. A public-facilities map indicates there are about 30 quarter-quarter sections within Redmond in which there is at least one drainage well, with an aggregate area of 1.88 mi^2. Analysis of 1995 aerial photographs suggests that there may be new residential areas not included in this total, but these represent only a small increase in the total area and are not included in the following calculation. Using the same values as in the analysis for Bend to represent the percentage of impervious area and evaporative losses and an average annual precipitation of 7.83 inches (1961–90), total runoff to drainage wells in Redmond is estimated to be approximately 9 million ft^3/yr, or about 0.28 ft^3/s. As with Bend, this is not a significant source of recharge.

Similar calculations were not carried out for other urban areas in the upper Deschutes Basin. Examples from Bend and Redmond, the most urbanized areas, illustrate that runoff to drainage wells is not an important volumetric component of ground-water recharge.

Although runoff to drainage wells is not volumetrically substantial, it may be significant in terms of water quality. Urban runoff can contain contaminants such as household pesticides and fertilizers, and automotive petroleum products. Runoff routed directly to drainage wells has a direct pathway to the ground-water system, bypassing the soil zone, where natural processes such as filtration, adsorption, and biodegradation may serve to reduce levels of some contaminants.

Interbasin Flow

The final source of recharge to the upper Deschutes Basin regional ground-water system is subsurface flow from adjoining basins. In general, the lateral boundaries of the upper Deschutes Basin study area are considered to be no-flow boundaries. There are, however, two areas where inflow from adjacent areas is probable: along the Cascade Range crest in the Metolius River drainage and in the southeastern part of the study area northeast of Newberry Volcano.

The western boundary of the study area coincides with the topographic crest of the Cascade Range. It is generally considered a no-flow boundary because the ground-water divide is assumed to follow the distribution of precipitation, which generally follows the topography. The isohyetal map of Taylor (1993) shows that in the area of the Metolius River subbasin, the region of highest precipitation occurs west of the topographic crest of the Cascade Range, suggesting that the ground-water divide is also to the west of the topographic divide and that there is likely ground-water flow eastward across the topographic divide. This interbasin flow is also indicated by the hydrologic budget of the Metolius River subbasin. Average ground-water discharge to the Metolius River in the study area above the gage near Grandview is approximately 1,300 ft^3/s. The mean annual recharge from precipitation in the Metolius River subbasin above this point in the study area is estimated to be only about 500 ft^3/s. The difference, 800 ft^3/s, almost certainly comes from subsurface flow from an adjacent basin. The most plausible source for this additional water is the upper Santiam and North Santiam River Basins to the west.

South of Bear Creek Butte, through Millican and the China Hat area, the eastern study-area boundary does not coincide with either a topographic divide or a geologic contact. The region east of this area was not included in the study area because of the lack of subsurface hydrologic information, very low recharge, and distance from the areas of primary concern. Hydraulic-head data, however, indicate there is some flow across this boundary into the study area from the southeast. This flux was estimated using a variety of methods.

The part of the Deschutes Basin east of this boundary is very dry (10 to 15 in./yr precipitation) and has a poorly developed drainage system with no perennial streams. The divide between this part of the Deschutes Basin and the Fort Rock and Christmas Lake Basins to the south is poorly defined and interbasin flow is likely. Miller (1986) states that flow to the Deschutes Basin from the Fort Rock Basin “probably exceeds 10,000 acre-ft/yr,” which equals about 14 ft^3/s. Estimates based on the Darcy equation, using measured head gradients and estimated hydraulic conductivity and aquifer thickness, suggest that the flux into the study area may be as high as 100 ft^3/s. Additional estimates were derived using a water-budget approach. The probable area contributing to the boundary flux was defined using hydraulic-head maps from the Deschutes Basin and the Fort Rock Basin (Miller, 1986). Flux rates were calculated using a range of recharge values from Newcomb (1953), Miller (1986), and McFarland and Ryals (1991). Assuming a contributing area of 648 mi^2 and recharge estimates ranging from 0.5 to 3.0 in./yr, the boundary flux could range from 25 to 145 ft^3/s. If recharge is assumed to be 1.0 in./yr in the contributing area for this boundary flux, the estimated flux rate is about 50 ft^3/s.

GROUND-WATER DISCHARGE

Ground water discharges from aquifers to streams, to wells, and through evapotranspiration. Discharge to streams is the principal avenue by which water leaves the ground-water system. Discharge can occur to discrete springs or as diffuse seepage through streambeds. Pumping by wells is another avenue by which ground water leaves the ground-water system. In the Deschutes Basin, discharge to wells represents a small fraction of the total ground-water discharge. Evapotranspiration by plants is the third mechanism considered in this report. Most plant water requirements are met by water percolating downward through the soil before it enters the ground-water system. In some areas where the water table is sufficiently shallow to be within the rooting depth of plants, transpiration can occur directly from the ground-water system. This process represents a very small fraction of the total ground-water discharge in the basin. Each of these mechanisms is discussed in more detail in the following sections.

Ground-Water Discharge to Streams

Discharge to streams is the main avenue by which water leaves the ground-water system and is one of the major components of the hydrologic budget. Ground water discharges to streams in areas where the stream elevation is lower than the elevation of the water table in adjacent aquifers. Considerable amounts of ground water can discharge to the streams in this way from regional aquifers with large recharge areas. Streams in which the flow increases due to ground-water discharge are termed *gaining streams*. The amount of ground water discharging to streams or leaking from streams varies geographically and with time.

Understanding the rates and distribution of ground-water discharge to streams is critical to understanding the ground-water hydrology of an area. The amount and location of ground-water discharge can be determined by measuring streamflow at points along a stream and accounting for tributary inflow and diversions between the points as well as temporal changes in flow. In general, increases in flow from point to point downstream that are not due to tributary inflow are caused by ground water discharging to the stream. Discharge can occur either at discrete locations such as springs or as diffused seepage through the streambed.

Stream-gage data can be particularly useful for estimating ground-water discharge. Gages on spring-fed streams, such as Fall River, measure ground-water discharge directly. Data from pairs of gages operated concurrently along a stream can be compared to estimate ground-water inflow between the gages as long as tributary inflow and diversions can be accounted for. Late summer and early fall flows in some streams are essentially entirely ground-water discharge (base flow). Therefore, annual low flows at certain stream gages can provide reasonable estimates of ground-water discharge.

Estimates of ground-water discharge to major streams in the upper Deschutes Basin are provided in table 7. These estimates are based on seepage runs and stream-gage data as well as other miscellaneous measurements. Unless otherwise noted, the values in table 7 represent approximate long-term average conditions.

Geographic Distribution of Ground-Water Discharge to Streams

There are three main settings in the upper Deschutes Basin where substantial amounts of ground-water discharge to streams: the southern part of the basin in and near the margin of the Cascade Range, the Metolius Basin adjacent to the Cascade Range, and the area surrounding the confluence of the Deschutes, Crooked, and Metolius Rivers extending downstream to about Pelton Dam (fig. 12). This latter area is referred to as the "confluence area" in this report.

Ground water constitutes a large proportion of the flow in many streams in and along the margin of the Cascade Range in the southern part of the basin (table 7). Ground water constitutes virtually the entire flow of some of these streams, such as Fall River. Such streams are recognized by the presence of source springs, lack of tributary streams, and flows that are very constant relative to other streams. Hydrographs of mean monthly flows (fig. 14) illustrate the differences between streams in which ground water is a the dominant source and those in which surface runoff is the dominant source. Fall, Cultus, and Quinn Rivers, and Browns Creek all show relatively little variation in flow throughout the year indicating that

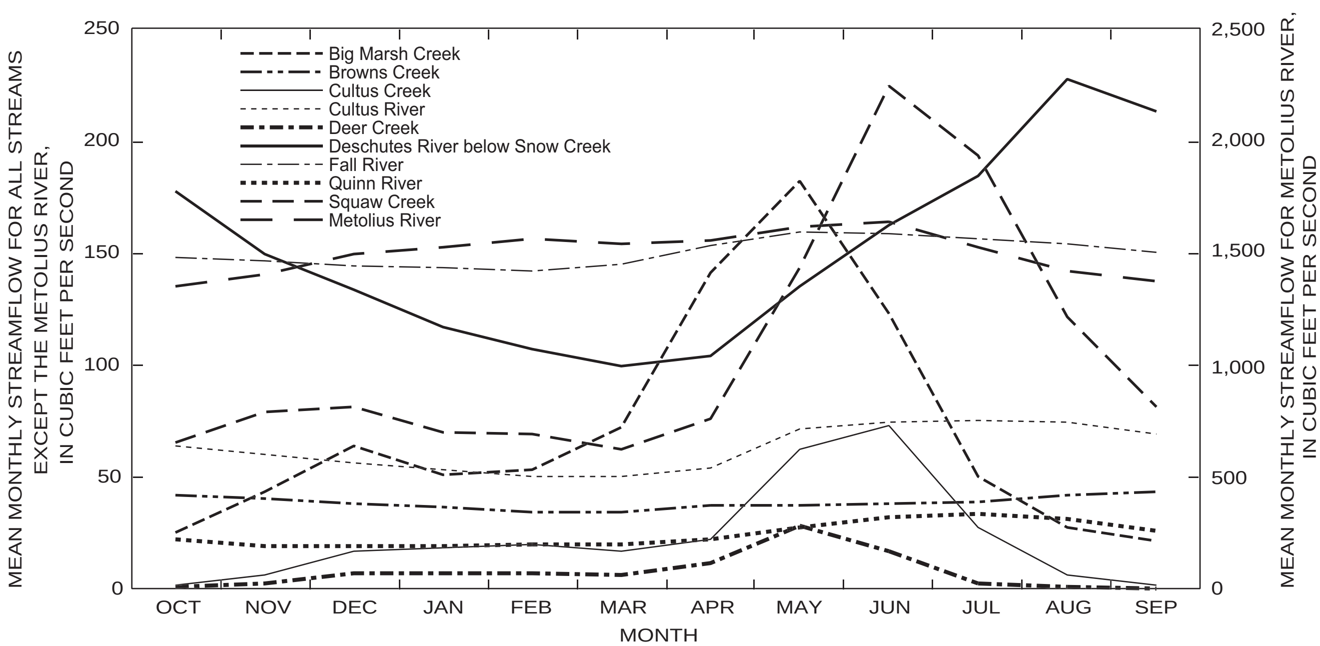


Figure 14. Mean monthly flows of selected nonregulated streams in the upper Deschutes Basin, Oregon.

they are not greatly affected by surface runoff and that ground water provides most of their flow. In contrast, Squaw, Big Marsh, Cultus, and Deer Creeks, and the Deschutes River (measured at the gage below Snow Creek just above Crane Prairie Reservoir) all show substantial increases in flow during spring due to runoff, indicating that their flow is dominated, or at least affected, by surface runoff.

Some of these runoff-dominated streams, such as the Deschutes River, have substantial flow even during the driest months of the year, indicating that ground-water discharge constitutes an important part of the flow. Others, such as Cultus and Deer Creeks, nearly cease to flow in the driest months of the year, indicating that ground-water discharge is only a minor part of their total flow. Temporal variations in ground-water discharge are discussed in more detail in a later section of the report.

The Metolius River drainage is the second region of significant ground-water discharge in and along the margin of the Cascade Range (fig. 12, table 7). The Metolius River drainage comprises numerous streams emanating from the Cascade Range, many of which are spring fed and others that are probably runoff dominated. The only long-term stream gage on the Metolius River is low in the drainage just above Lake Billy Chinook (this gage is officially referred to as being near Grandview, an abandoned town site). Although this gage represents a large drainage area that encompasses both spring-fed and runoff-dominated streams, it warrants analysis because of the large volume of ground water that discharges in the Metolius River drainage. Two tributary streams, Jefferson Creek and Whitewater River, carry glacial runoff from Mt. Jefferson and have late-season flows not entirely attributable to ground-water discharge.

A hydrograph of the monthly mean flow of the Metolius River near Grandview from 1922 to 1997 (fig. 15) clearly shows transient runoff events caused by spring snowmelt and large storms. During the late summer, however, when surface runoff is minimal, the flow of the Metolius is largely ground-water discharge. These late-summer flows are relatively large, reflecting the large amount of ground-water discharge. The lowest mean monthly flow occurs during October. The mean October flow of the Metolius River near Grandview for the period 1912–87 was 1,350 ft^3/s (Moffatt and others, 1990). This amount includes the flow of Jefferson Creek and Whitewater River, which may include late-season glacial melt, but the contribution from these streams is relatively small. The mean October flow of Jefferson Creek was 77 ft^3/s during

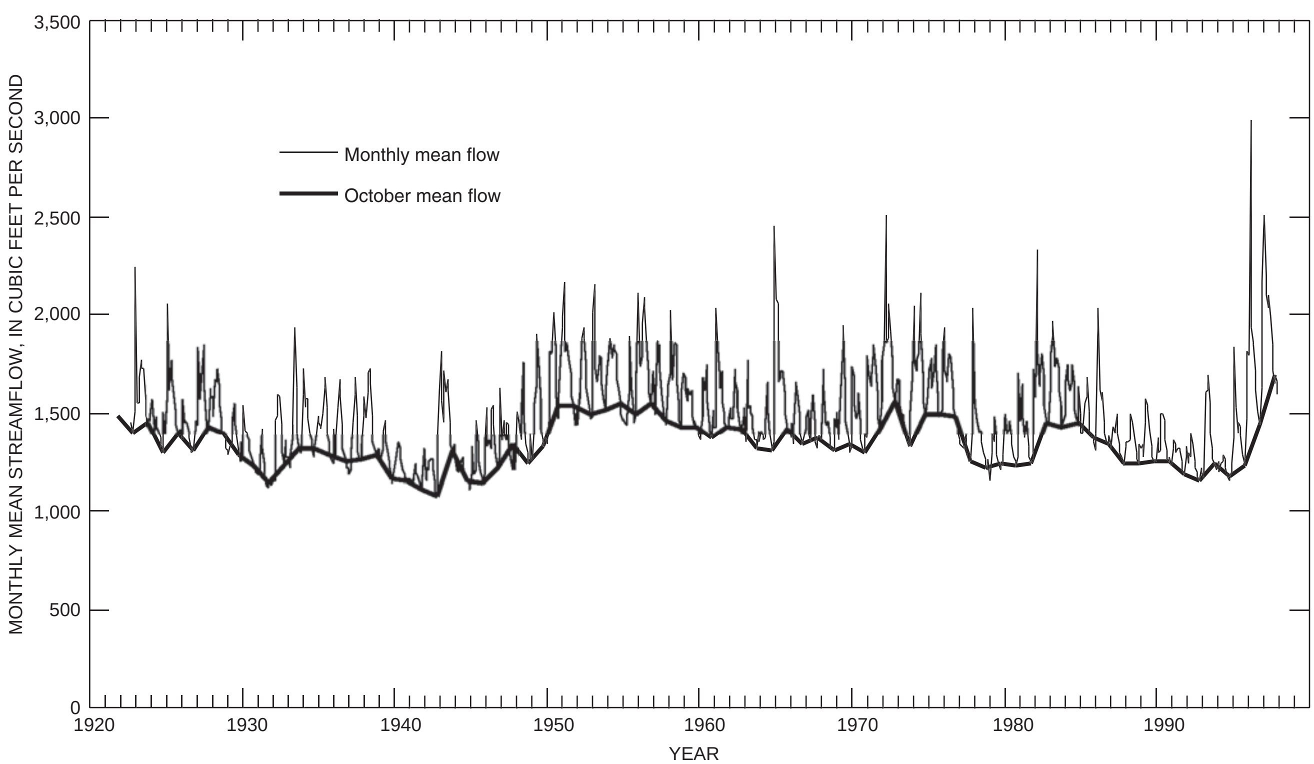


Figure 15. Monthly mean flow of the Metolius River near Grandview. (The line connecting the October mean flows approximates ground-water discharge.)

the period 1984–98 and that of Whitewater River was 53 ft^3/s during the period 1983–98. Depending on the amount of the mean October flow of these streams that is glacial in origin, the mean October flow of the Metolius River near Grandview that is derived from ground-water discharge is between 1,220 and 1,350 ft^3/s.

A variety of regional geologic factors controls the location of ground-water discharge to streams and springs in and along the margins of the Cascade Range. Many large spring areas and gaining stream reaches, such as Fall and Spring Rivers, coincide with the boundary of the La Pine and Shukash structural basins. The low-permeability basin-filling sediments likely divert ground water toward the surface by acting as an impediment to subsurface flow.

Geologic structure can also influence ground-water discharge in and along the margins of the Cascade Range. The tremendous amount of ground water discharging to the upper Metolius River and its tributaries is undoubtedly due in large part to the major fault system along the base of Green Ridge (fig. 4). Green Ridge is a 20-mile long escarpment that marks the eastern margin of a north-south trending graben into which the Cascade Range in that area has subsided (Allen, 1966; Priest, 1990). Vertical movement along this fault system is estimated to be over 3,000 ft (Conrey, 1985). The fault system may influence ground-water discharge in two ways. First, elevation of the valley on the downthrown side of the fault system is anomalously low when compared to surrounding terrane a similar distance from the Cascade Range. Low-elevation areas commonly are regions of ground-water discharge. Second, the fault itself likely impedes eastward movement of ground water flowing from the Cascade Range, forcing ground water to discharge to the river. The impediment to eastward ground-water movement could be due to low-permeability crushed or sheared rock along the fault planes or the juxtaposition of permeable strata on the west side of the fault system against low-permeability strata on the east. Analysis of carbon isotope data (James and others, 1999) suggests that the water discharged from the Metolius River springs includes a component of deep regional ground water, implying that there is vertical permeability locally along the escarpment.

Local geology also affects the location of ground-water discharge. Many springs occur along the edges or ends of Quaternary lava flows. Ground water emerges at these locations because saturated permeable zones in or at the base of the lava flows intersect land surface. Some springs, such as those at the upper end of Davis Creek, emerge in buried stream channels at the ends of intracanyon lava flows.

The total average amount of ground water discharging to streams in and along the margin of the Cascade Range in the study area is estimated to be approximately 2,600 ft^3/s. This includes discharge to streams in the southern part of the study area, in the Tumalo and Squaw Creek drainages, and in the Metolius River drainage (table 7). Approximately one-half of this amount discharges in the Metolius River drainage.

The third major setting in which ground water discharges to streams is the region around the confluence of the Deschutes, Crooked, and Metolius Rivers and extending downstream to the vicinity of Pelton Dam. Russell (1905, p. 88) provides an early description of ground-water inflow in this region:

> Crooked River at Trail Crossing, at the time of my visit in early August [1903], had shrunk to a brook of tepid, muddy, and unwholesome water, across which one could step dry-shod from stone to stone. Its volume, by estimate, was not more than 2 cubic feet per second.... On descending the canyon about 12 miles lower down its course I was surprised to find a swift-flowing, clear stream of cool, delicious water, by estimate 100 feet wide and 3 feet deep, with a volume of not less than 300 cubic feet per second. This remarkable renewal or resuscitation of a stream in an arid land is due to the inflow of Opal and other similar springs.

Stearns (1931) also recognized the large amount of ground water discharging to streams in the area while investigating the geology and hydrology of the middle Deschutes Basin for potential dam sites. Stearns used stream-gage data to conservatively estimate ground-water inflow to the lower Crooked River between Trail Crossing and the gaging station near Culver (now under Lake Billy Chinook) to be 950 ft^3/s. He also used gage data to estimated ground-water inflow to the Deschutes River between Bend and Madras at about 600 ft^3/s. These numbers are generally consistent with modern estimates when the effects of irrigation development and of Round Butte Dam are considered.

Ground-water discharge to the lower Crooked River and middle Deschutes River was estimated from OWRD seepage runs (fig. 12, table 5). Ground-water discharge to the lower Crooked River between Terrebonne and the gage below Opal Springs was approximately 1,100 ft^3/s in June 1994 (fig. 16, table 5).

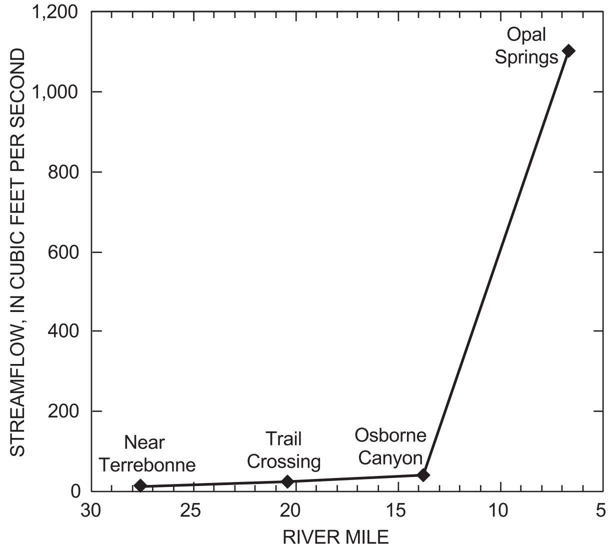


Figure 16. Gain in flow of the lower Crooked River, Oregon, due to ground-water discharge between river miles 27 and 7, July 1994.

Most of this inflow entered the Crooked River below Osborne Canyon, about 7 miles upstream from the gaging station below Opal Springs. The Deschutes River gained approximately 400 ft^3/s along the 10-mile reach above the gaging station near Culver, just above Lake Billy Chinook, during seepage runs in May 1992 and May 1994 (fig. 17, table 5). About 300 ft^3/s of this gain was from ground-water discharge directly to the Deschutes River, and the remaining 100 ft^3/s was mostly from ground-water discharge to lower Squaw Creek near its confluence with the Deschutes River. A seepage run made along Squaw Creek in April 1994, combined with data from the seepage run along the Deschutes River a month later, showed Squaw Creek gaining approximately 94 ft^3/s from springflow in the lower 1.7 miles from Alder Springs to the confluence (table 7).

The ground-water discharge estimates from seepage runs on the lower Crooked River, Deschutes River, and Squaw Creek are probably conservative estimates of long-term mean annual ground-water discharge. The seepage runs were conducted after a period of several relatively dry years. The monthly mean streamflows for the months during which the seepage runs were conducted were low compared to the long-term mean monthly flows (Hubbard and others, 1993, 1995). Temporal variations in ground-water discharge are discussed in a later section.

Ground-water inflow to Lake Billy Chinook, estimated from stream-gaging-station data, is roughly 420 ft^3/s (the middle of the range in table 7). From Round Butte Dam downstream to Dry Creek at river mile 91.8 (about 2.5 miles below Shitike Creek), the Deschutes River gains about 400 ft^3/s from ground-water inflow (table 7). There is no significant ground-water inflow directly to the Deschutes River downstream from this point. The total amount of ground water discharging to the Deschutes and Crooked

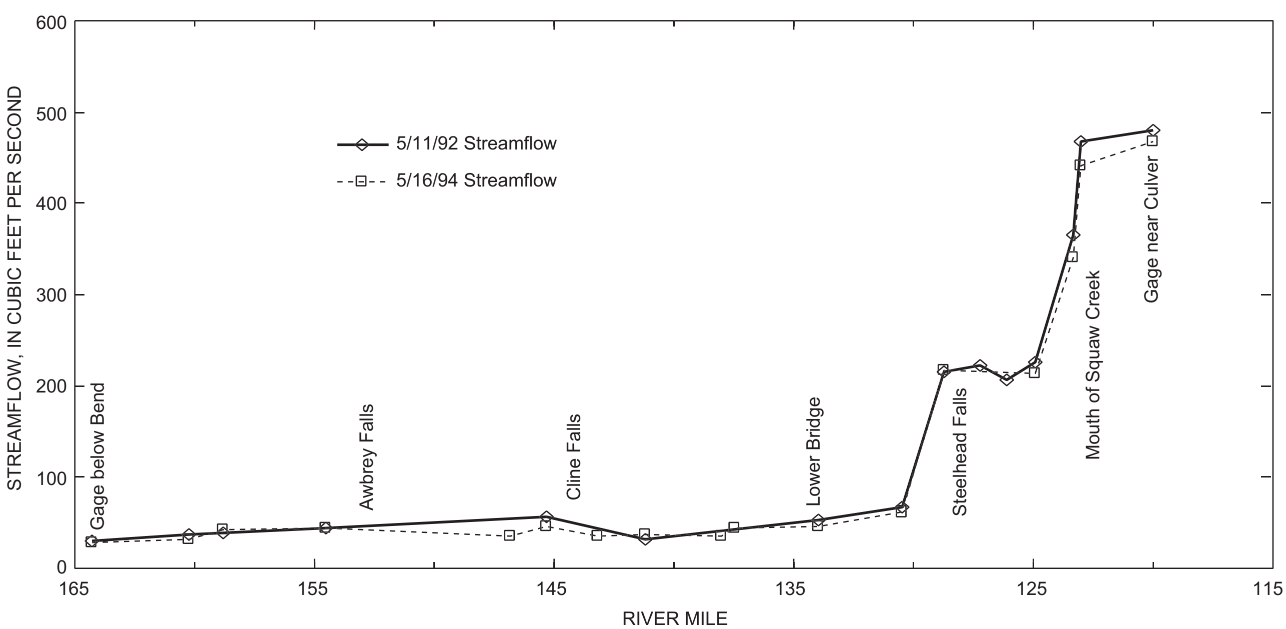


Figure 17. Gain in flow of the Deschutes River, Oregon, due to ground-water discharge between river miles 165 and 120, May 1992 and May 1994. (Some of the gain is due to ground-water discharge along the lower 2 miles of Squaw Creek.)

Rivers in the area extending from about 10 miles above Lake Billy Chinook to Dry Creek is approximately 2,300 ft^3/s. This is probably a conservative estimate for the reasons previously discussed.

The ground-water discharge estimate in the confluence area (2,300 ft^3/s) cannot be simply added to the discharge estimate for streams emanating from the Cascade Range (2,600 ft^3/s) to estimate average net ground-water discharge to streams in the basin. The resulting value exceeds the total estimated ground-water recharge for the entire upper Deschutes Basin. This is because the streams to which ground water discharges in the upper basin lose some of that water (as much as 600 ft^3/s) back to the ground-water system through stream and canal leakage. This water discharges once again in the confluence area. Therefore, a fraction of the ground water discharged in the confluence area has entered and been discharged from the ground-water system twice.

Ground-water discharge in the confluence area is controlled primarily by geology. Sceva (1960), in a report prepared for the Oregon Water Resources Board, was the first to describe the influence of the geology on regional ground-water flow and discharge. His basic conceptual model was largely corroborated by subsequent data collection and analysis. In a later report he states: "A barrier of rocks having a low permeability transects the Deschutes River Basin near Madras. This barrier forces all of the ground water to be discharged into the river system... (Sceva, 1968, p. 5)."

The Deschutes Basin is transected by a broad ridge composed of the John Day Formation, a rock unit of very low permeability that extends, with varying degrees of exposure, from the Gray Butte area north to the Mutton Mountains (outside and to the northwest of the study area) and east into the John Day Basin (fig. 4). This broad ridge is part of a regional uplift extending from central to northeastern Oregon known as the Blue Mountain anticline (Orr and others, 1992). The John Day Formation in this area consists of tuffaceous claystone, air-fall and ash-flow tuffs, and lava flows (Robinson and others, 1984). The ridge of the John Day Formation represents an ancient upland that formed the northern and eastern boundary of the basin into which the permeable Deschutes Formation was deposited. North of Madras, the Deschutes Formation, through which most regional ground water in the upper basin moves, becomes increasingly thin and eventually ends. Because the John Day Formation has such low permeability, ground water cannot move farther north in the subsurface and is forced to discharge to the Crooked and Deschutes Rivers, which have fully incised the Deschutes Formation (fig. 18). Analysis of stream-gaging data shows that there is no significant ground-water discharge to the Deschutes River downstream from the area where the John Day Formation forms the walls of the river canyon.

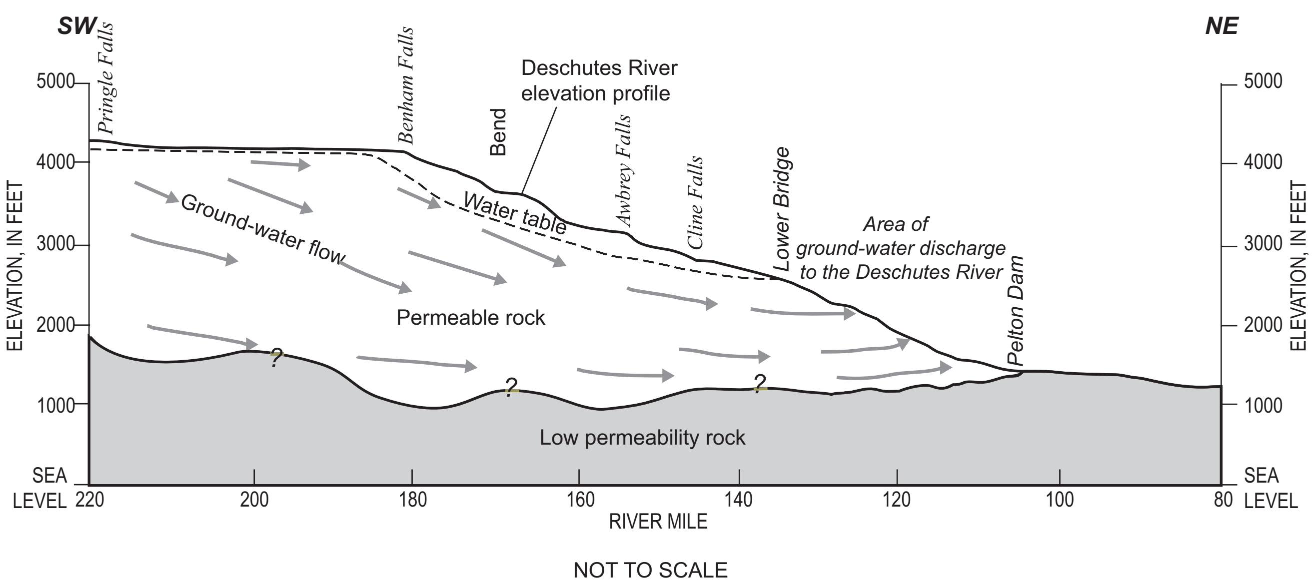


Figure 18. Diagrammatic section showing the effect of geology on ground-water discharge along the Deschutes River upstream of Pelton Dam.

Temporal Variations in Ground-Water Discharge to Streams

Ground-water discharge to streams not only varies from place to place, but varies with time as well. The rate of ground-water discharge varies on many time scales, but for this study, annual and decadal time scales are examined. Annual discharge variations are driven by the seasonal variations in precipitation and ground-water recharge. Decadal variations in ground-water discharge in the Deschutes Basin are driven by variations in precipitation and recharge due to climate cycles. Longer-term variations in discharge, occurring over many decades, can be caused by long-term climate trends. Ground-water discharge variations at all of these time scales can be influenced by human activity. Temporal variations in ground-water discharge in the basin are discussed in the following paragraphs.

Virtually all the data on temporal variations in ground-water discharge were derived from stream gages, where continuous records of stream discharge were recorded (fig. 11). Data from stream gages are useful for estimating ground-water discharge only in certain circumstances. Regulation of streamflow at upstream dams or other control structures precludes the use of some gages for estimating ground-water discharge. If the gage is at a location where it is known that the streamflow is virtually entirely from ground-water discharge, such as with spring-fed streams like Fall River, then the gage provides a continuous direct measurement of ground-water discharge. In such cases, the gage can provide information on variations in ground-water discharge at many time scales ranging from daily to long term. In other circumstances, such as along the lower Crooked River at Opal Springs, streamflow can only be assumed to represent ground-water discharge during the driest months of the year when surface runoff from upstream is negligible compared to known inflow from springs. In cases such as this, the gage cannot be used to evaluate seasonal variations in ground-water discharge, but can provide information on year-to-year variations. In some circumstances, a set of gages operated concurrently on a stream can be used to estimate ground-water inflow to the stream between the gages as long as there is no unmeasured tributary inflow or diversion along the intervening reach.

Stream-gage data suitable for estimating temporal variations in ground-water discharge are available for only a few locations in the upper Deschutes Basin because stream gages are typically located and operated for other reasons. However, the main ground-water discharge settings are represented in the available data.

Stream-gage data are available for a number of small spring-fed streams along the margin of the Cascade Range in the southern part of the basin, including Cultus, Quinn, and Fall Rivers, and Browns Creek. The flow in these streams is almost entirely ground-water discharge, as indicated by constant flow throughout the year (fig. 14). The gages on these streams provide an approximate continuous measure of ground-water discharge. The flow in these streams does vary seasonally, and they do exhibit annual peaks in flow. The magnitude of the peak flow is attenuated and the timing of the peak flow is delayed when compared with runoff-dominated streams such as Cultus, Deer, and Big Marsh Creeks (fig. 14). The differences between ground-water- and surface-water-dominated streams is apparent in the statistics of their mean monthly flows (table 8). The range in mean monthly flows for surface-water-dominated streams is over 200 percent of their mean annual flow. The months with the highest mean flows for surface-water-dominated streams are May and June. The range in mean monthly flows for ground-water-dominated streams, in contrast, is only 11 to 58 percent of their mean annual flows, and the high flow may occur any month from May through September. The peaks in flow seen in ground-water-dominated streams are caused by the same snowmelt events that provide peak discharge to runoff-dominated streams. Because the water must percolate through the soil and move through the subsurface before discharging to spring-fed streams, the peaks in flow are attenuated and delayed.

The time lag between the annual peak snowmelt and the peak in the flow of these spring-fed streams is proportional to the degree of attenuation of annual flow peak; in other words, the more subdued the peak flow, the longer the time lag (Manga, 1996). A mathematical model for ground-water-dominated streams in the Cascade Range developed by Manga (1997) relates the degree of attenuation and the time lag of the peak streamflow to the generalized geometry and hydraulic properties of the aquifers feeding the stream. In Manga's model, the annual recharge pulse caused by snowmelt is essentially diffused along the length of the aquifer causing the attenuation and delay in the peak flow. This suggests that streams fed by aquifers with large areas are likely to have more uniform flow and a longer delay between recharge events and peak flows when compared to streams fed by aquifers with small capture areas.

Table 8. Statistical summaries of selected nonregulated streams in the upper Deschutes Basin, Oregon
[Source: Moffatt and others, 1990; ft^3/s, cubic feet per second]

Station name	Station number	Period of record	Mean annual flow (ft^3/s)	Highest mean monthly flow (ft^3/s)	Month of highest mean monthly flow	Lowest mean monthly flow (ft^3/s)	Month of lowest mean monthly flow	Variation as percentage of mean annual flow
Deschutes River below Snow Creek	14050000	1937–87	151	227	August	99	March	85
Cultus River above Cultus Creek	14050500	1923–87	63	75	July	50	February–March	40
Cultus Creek above Crane Prairie Reservoir	14051000	1924–62	22	73	June	1.2	October	326
Deer Creek above Crane Prairie Reservoir	14052000	1924–87	7.5	28	May	0.2	September	371
Quinn River near La Pine	14052500	1938–87	24	33	July	19	November–January	58
Browns Creek near La Pine	14054500	1923–87	38	43	September	34	February–March	24
Fall River near La Pine	14057500	1938–87	150	159	May	142	February	11
Big Marsh Creek at Hoey Ranch	14061000	1912–58	72	182	May	21	September	224
Squaw Creek near Sisters	14075000	1906–87	105	224	June	62	March	154
Metolius River near Grandview	14091500	1912–87	1,500	1,640	June	1,350	October	19

The spring-fed streams in the southern Deschutes Basin exhibit decadal flow variations in addition to annual variations. Individual peak periods on Fall River, for example, are roughly 5 to 14 years apart. Decadal variations in annual mean discharge can be substantial. Stream-gage data show that between 1939 and 1991 the annual mean flow of Fall River varied from 81 to 202 ft^3/s and the annual mean flow of Cultus River ranged from 36 to 96 ft^3/s. These decadal variations in ground-water discharge are driven by climate cycles. Comparing the ground-water discharge variations with precipitation at Crater Lake in the Cascade Range (both as cumulative departures from normal) shows that periods of high ground-water discharge generally correspond with periods of high precipitation (fig. 19).

Stream-gage data also provide information on temporal variations in ground-water discharge in the Metolius River drainage. As mentioned in the preceding section, the only long-term gage on the Metolius River is in the lower part of the drainage near Grandview, which measures discharge from a relatively large area. Because the drainage area represented by this gage includes runoff-dominated streams, the data cannot be used to evaluate seasonal variations in ground-water discharge. Evaluating the late summer and early fall flows, when most streamflow is ground-water discharge, however, can provide information on the long-term variations in ground-water discharge in the basin.

Before evaluating base flow to the Metolius River, the effects of tributary streams potentially carrying glacial meltwater during the late summer must be considered. In figure 20, a graph of October mean discharge values for the Metolius River is shown with similar graphs of Jefferson Creek and Whitewater River. Subtracting the flow of Jefferson Creek and Whitewater River shifts the graph of the Metolius River downward, but does not affect the overall shape of the graph or magnitude of variation (fig. 20). This suggests that the variations in October mean flows in the Metolius River are not greatly affected by these glacial streams and probably reflect variations in ground-water discharge.

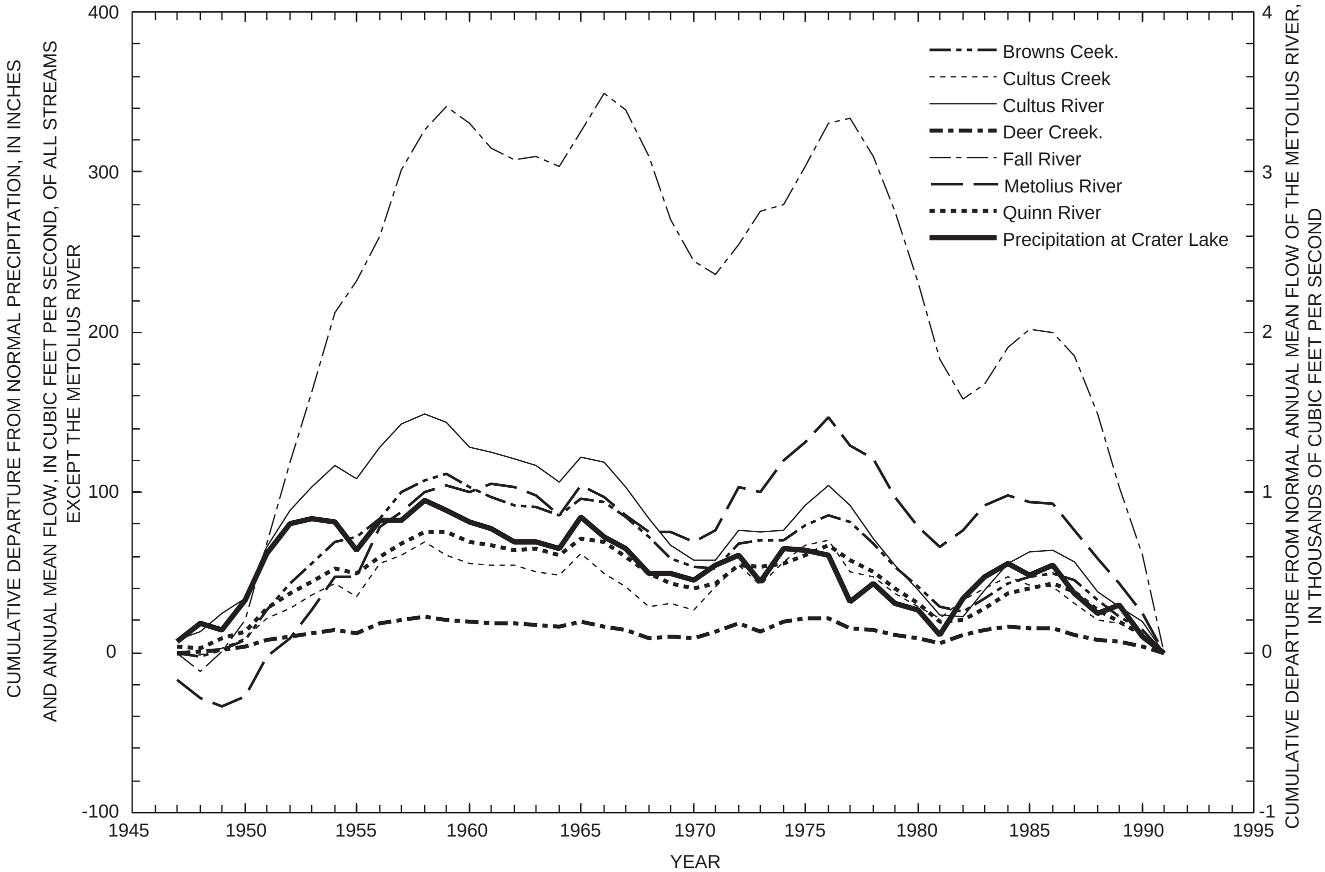


Figure 19. Cumulative departure from normal annual mean flows of selected streams in the upper Deschutes Basin, and cumulative departure from normal annual precipitation at Crater Lake, Oregon, 1947–91.

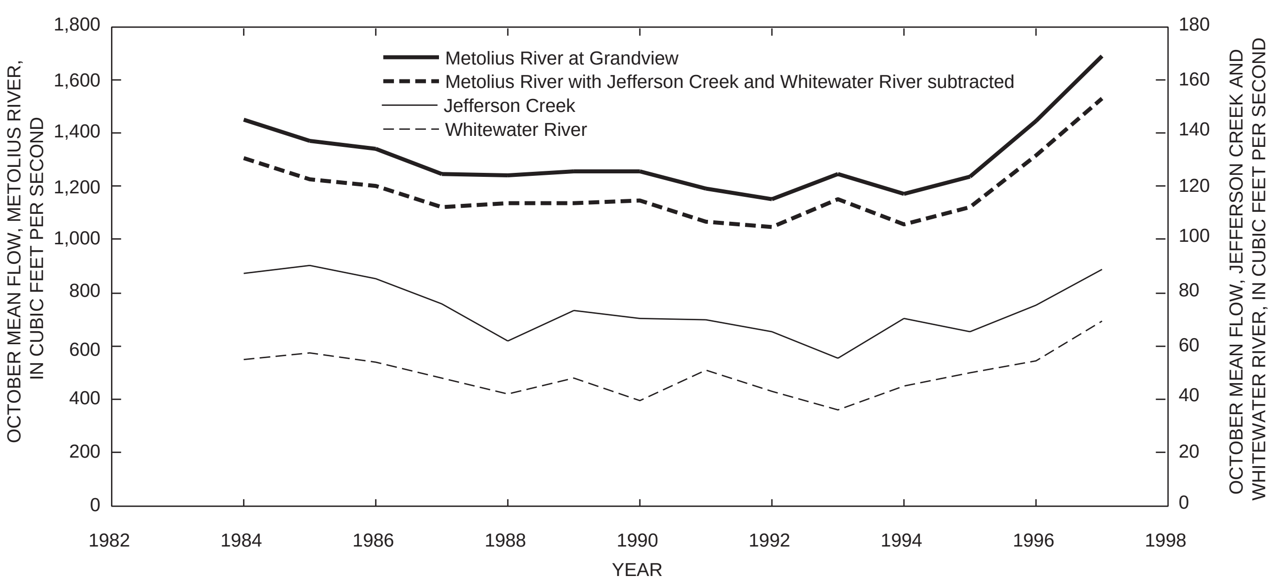


Figure 20. October mean flows of the Metolius River (near Grandview), Jefferson Creek, and Whitewater River, upper Deschutes Basin, Oregon, 1984–97.

Variations in long-term discharge of the Metolius River at Grandview exhibit a pattern similar to that seen in other Cascade Range streams. Comparison of the annual mean discharge of the Metolius River with precipitation at Crater Lake (both as cumulative departures from normal) shows that variations in base flow of the Metolius River follow variations in Cascade Range precipitation to a large degree, as is the case with other Cascade streams (fig. 19). Because of the size of the drainage basin, the magnitude of the decadal variation in ground-water discharge to the Metolius River is less than that in the smaller ground-water-dominated streams in the upper basin. For example, the 407 ft^3/s variation in October mean discharge of the Metolius River from 1962 to 1997 is about 30 percent of the mean October discharge for the period. The variation in October mean discharge for Fall River, by comparison, is about 74 percent of the mean October discharge flow for the same period.

Stream-gage data also allow evaluation of temporal variations in ground-water discharge in the area near the confluence of the Deschutes and Crooked Rivers. Data are available for reaches of both the Crooked and Deschutes Rivers above Lake Billy Chinook. In both cases, unmeasured tributary inflow during parts of the year preclude analysis of seasonal variations and allow analysis only of interannual and longer-term variations.

Variations in ground-water discharge to the Deschutes River in the confluence area can be evaluated by comparing discharge records from stream gages below Bend and near Culver just above Lake Billy Chinook. Seepage runs (table 5), discussed in a preceding section, indicate that most of the ground-water discharge to this reach occurs within 10 miles of Lake Billy Chinook.

Two major tributaries, Tumalo and Squaw Creeks, join the Deschutes River between the Bend and Culver gages. Neither of these tributaries have gaging stations near their mouths. During the irrigation season (April to November), most of the flow of these streams is diverted. Tumalo Creek flows only a few cubic feet per second at its confluence with the Deschutes River during this time (table 5). Squaw Creek typically flows about 100 ft^3/s at its confluence with the Deschutes River during the irrigation season (table 5), but nearly all of this flow is from springs (including Alder Springs) within 1.7 miles of the mouth. Flow in Squaw Creek above the springs is typically only a few cubic feet per second. It is reasonable, therefore, to consider the net gain in streamflow along the Deschutes River between the gages below Bend and near Culver during the late summer and early fall to be almost entirely due to ground-water discharge along the lower part of that reach, including the lower 2 miles of Squaw Creek.

A graph of the difference between August mean flows at the Bend and Culver gages from 1953 to 1997 (fig. 21) shows that August mean ground-water discharge varied from 420 to 522 ft^3/s and exhibited a pattern of variation similar to other streams in the basin. The 102 ft^3/s variation in August mean ground-water discharge to this reach of the Deschutes River from 1962 to 1997 is about 22 percent of the mean August value. This is less than the base flow variations of 30 and 76 percent for the Metolius and Fall Rivers, respectively, during this same period. The smaller variation in ground-water discharge to the Deschutes River results from the larger size of the ground-water contributing area and the distance from the source of recharge.

Variations in ground-water discharge to the lower Crooked River can be evaluated using the gage below Opal Springs. This gage is located in the midst of the most prominent ground-water discharge area in the Deschutes Basin. A seepage run made in June 1994 (table 5) showed that ground-water discharge between Terrebonne and the gage at Opal Springs (a distance of about 21 miles) exceeded 1,100 ft^3/s, of which over 1,000 ft^3/s entered the river in the lower 7 miles of this reach. During much of the year, the streamflow at the Opal Springs gage includes a large amount of surface runoff in addition to ground-water discharge (fig. 22). During the irrigation season, however, most of the flow above Terrebonne is diverted, and flow from upstream into the ground-water discharge area is normally minuscule compared with the volume of ground-water inflow. Therefore, the late-summer flow at the Opal Springs gage is presumed to be almost entirely ground-water discharge except during anomalous storm events or reservoir releases.

August mean flows at the Opal Springs gage between 1962 and 1997 (fig. 22), representing ground-water discharge, exhibit climate-driven long-term variations apparent in other streams in the basin. August mean discharge for the period from 1962 to 1997 ranged from 1,133 to 1,593 ft^3/s, a variation of 460 ft^3/s, or 35 percent of the mean August discharge. The variation in July mean flows for the same period was only 28 percent. This variation is larger than one would expect given the volume of discharge, apparent size of the ground-water contributing area, and the observed variations in discharge to the Deschutes River.

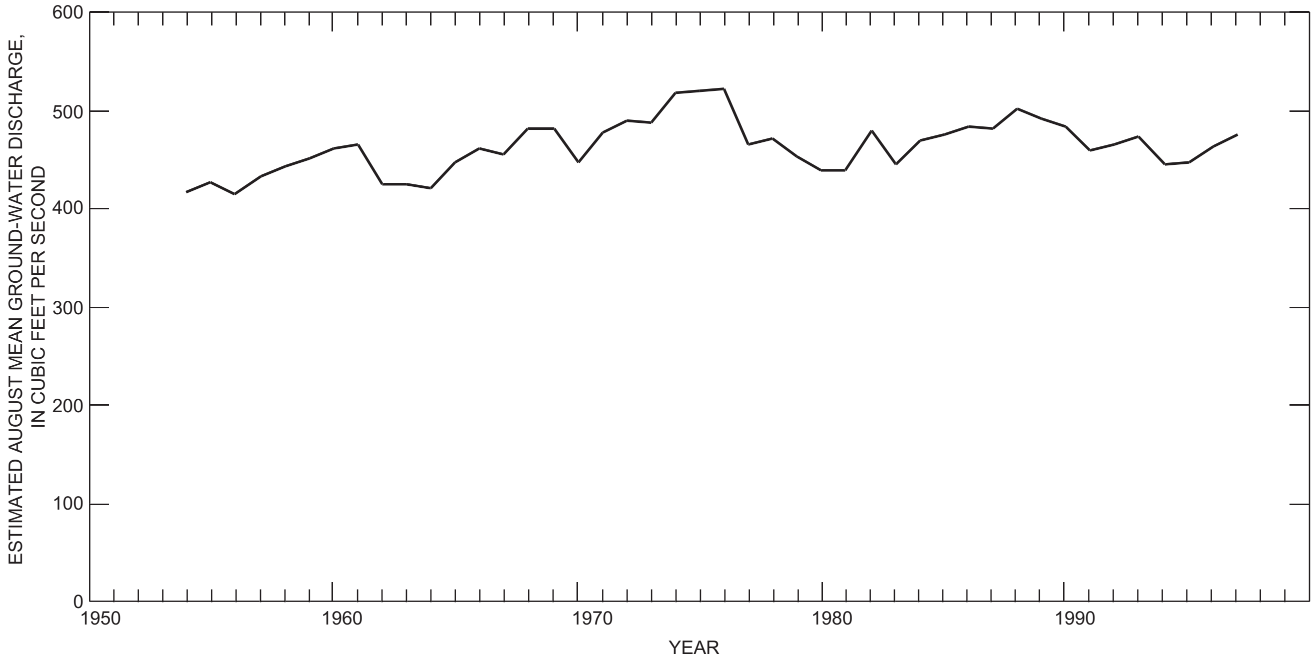


Figure 21. Approximate August mean ground-water discharge to the middle Deschutes River between Bend and Culver, based on the difference between August mean streamflows at gages below Bend and near Culver, 1954–97. (Fluctuations are caused by variations in ground-water discharge.)

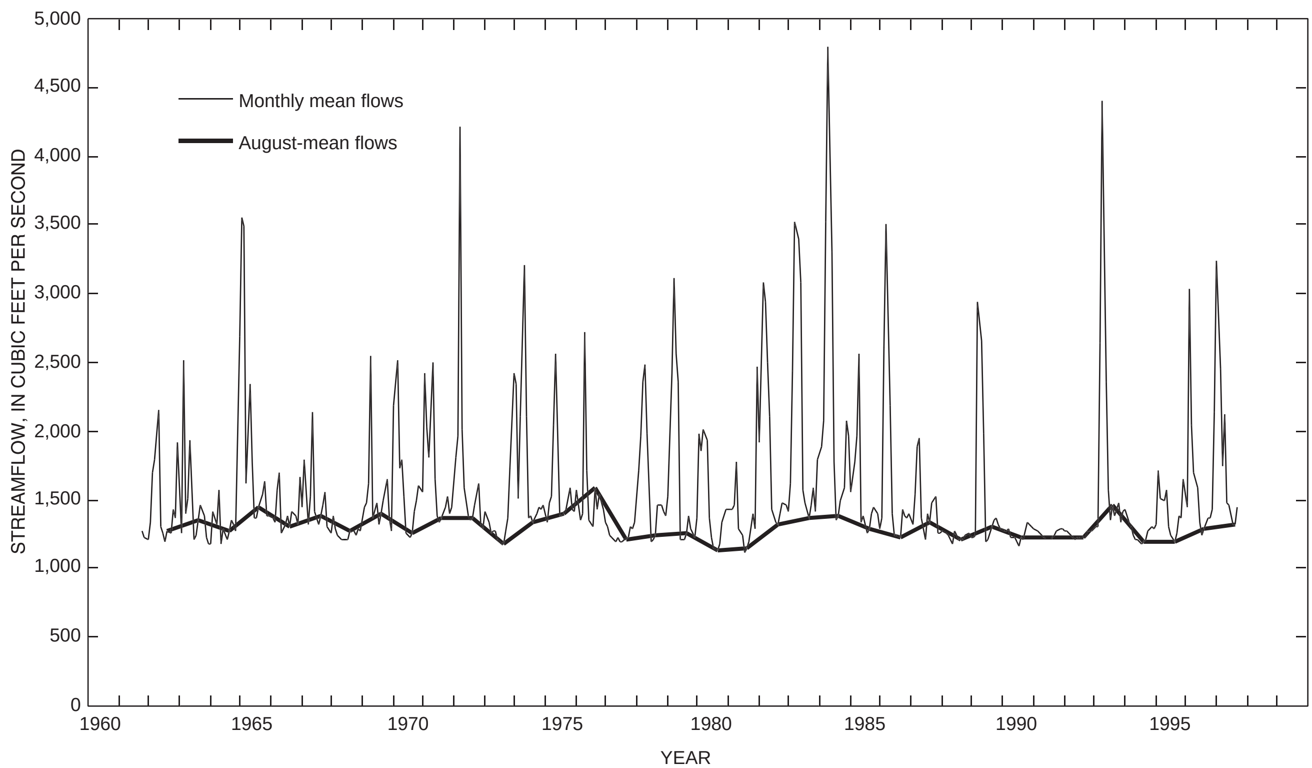


Figure 22. Monthly mean flows of the Crooked River at the gage below Opal Springs, 1962–97. (The line connecting August mean flows approximates late-season ground-water discharge.)

This variation may be due to streamflow from above the ground-water discharge area. The Crooked River above the gage includes a very large area of runoff-dominated streams and two major reservoirs. The larger-than-expected variation may also be due to variations in canal leakage, which contributes ground-water inflow to the lower Crooked River.

Variations in ground-water discharge to the Metolius, Deschutes, and Crooked Rivers are driven by the same climatic trends and parallel each other. The variations, therefore, are additive and can combine to account for variations in late season monthly mean discharge on the order of 1,000 ft^3/s below the confluence area at the gage near Madras. Late-season (July to September) mean monthly flows at the gage near Madras, which are primarily ground-water discharge, average about 4,000 ft^3/s. Therefore, climate-driven variations in ground-water discharge can account for late-season streamflow variations of 25 percent at Madras.

Analysis of stream-gage data from the lower Crooked River from the early 1900s through the 1960s shows an increase in ground-water discharge that is attributed to irrigation canal leakage. The graph of August mean discharge of the lower Crooked River (fig. 23) includes data from two different gage sites. Prior to the construction of Round Butte Dam and filling of Lake Billy Chinook, the gage was operated on the Crooked River at a now-inundated location near Culver, about 5.6 miles downstream from the present gage location. The flow is different at these two sites because the lower (former) site includes flow from springs not measured by the present gage, causing an offset between the two hydrographs.
The hydrograph of August mean discharge of the lower Crooked River shows an overall increase of approximately 400 to 500 ft^3/s between 1918 and the early 1960s (fig. 23). The increase is given as a range because the exact amount is uncertain due to year-to-year variability in the flow. This steady, long-term trend of increasing discharge is not observed in other streams, such as the Metolius River, and does not appear to be caused by climate. It is also different from later long-term variations in August mean flows. This increase in base flow to the lower Crooked River is, however, similar in volume to estimated annual mean irrigation canal losses. Moreover, the growth of the increase is similar to that of estimated canal leakage (fig. 23). The return of water lost through canal leakage back to the surface as base flow to the Crooked River is consistent with ground-water flow directions in the area.

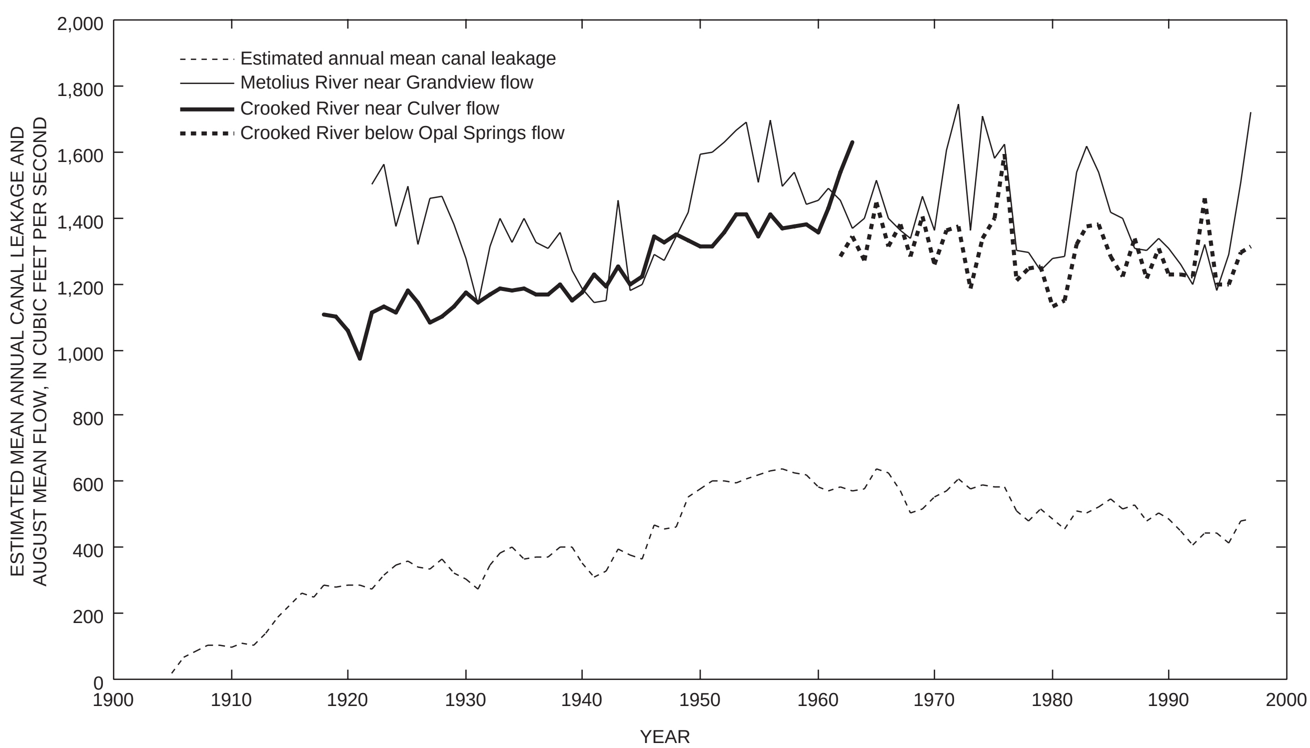


Figure 23. August mean flows of the Crooked River below Opal Springs, the Metolius River near Grandview, and estimated annual mean leakage from irrigation canals, 1905–97.

Ground-Water Discharge to Wells

Ground water is pumped from wells for a variety of uses in the upper Deschutes Basin, including irrigation, public supply, and private domestic use. Irrigation is primarily agricultural, but can include watering of golf courses and parks. Public-supply systems include publicly and privately owned water utilities, which are typically located in urban and suburban areas. Public-supply use includes not only drinking water, but also commercial, industrial, and municipal uses. Private domestic use generally refers to pumpage by individual wells that typically supply a single residence. Pumpage for each of these uses is discussed in this section.

Irrigation Wells

Pumpage of ground water for irrigation was estimated using water-rights information from the State of Oregon and crop-water-requirement estimates (fig. 24). Crop-water requirements were estimated, as previously described, for each irrigated 40-acre tract in the study area. The proportion of each tract irrigated with ground water was identified using water-rights information from the State of Oregon. A well serving as the primary source of water was identified for each tract irrigated using ground water. Where multiple wells supply water to the same 40-acre tract, the amount of water was proportioned between the wells based on the instantaneous rate information in the water-right files. For example, if it was determined that the crop-water requirements plus irrigation-efficiency requirements totaled 100 acre-ft/yr in a particular 40-acre tract, and that there were two wells with water rights listing instantaneous rates of 1 and 3 ft^3/s, then the two wells would be assigned annual pumpage rates of 25 and 75 acre-ft/yr respectively.

The crop-water requirements for all tracts, or parts thereof, were summed for each well. These sums were then divided by the irrigation efficiency (0.75) to derive an estimate of the total pumpage from each well. Water not lost through irrigation inefficiency or transpiration by plants is assumed to return to the ground-water system through deep percolation below the root zone and not be consumptively used.

Pumpage of ground water for irrigation was estimated to be about 14,800 acre-ft/yr (an average annual rate of 20.4 ft^3/s) during 1994, the year in which the crop-water requirements were estimated. Ground-water pumpage was estimated for each year from 1978 through 1997 by adjusting the 1994 pumpage up or down using an index reflecting the potential evapotranspiration and accounting for the change in the number of water rights with time. Potential evapotranspiration values were derived from the DPM (described in a previous section of this report) and adjusted to more accurately reflect rates measured by the BOR at the AgriMet site near Madras. Estimated ground-water pumpage for irrigation from 1978 to 1997 is shown in figure 24. The geographic distribution of average annual ground-water pumpage for irrigation from 1993 to 1995 is shown in figure 25.

Public-Supply Wells

Public water-supply systems use a large proportion of the ground water pumped in the upper Deschutes Basin. Pumping for public water supplies has increased steadily in recent years in response to population growth (fig. 26). Total ground-water pumpage for public-supply use as of 1996 was estimated to be about 15,100 acre-ft/yr, an average rate of about 20.8 ft^3/s. Public-supply pumpage is concentrated primarily in urban and major resort areas, with scattered pumpage by smaller, rural systems (fig. 27).

Public-supply pumpage was estimated using data provided by operators of the 19 major municipal water systems and private water utilities in the upper basin. The quality and completeness of data from these systems varied widely. Some systems have totalizing flow meters on their wells, while others estimate pumpage using hour meters and known or calculated pumping rates. Complete records were not available for all systems for all years of interest. A variety of techniques was employed to estimate pumpage where records were incomplete or missing. Where data from early years were not available, pumpage was estimated by using estimates of the number of individuals served or the number of connections to the system. In cases where data were missing for certain time intervals, pumpage was estimated by interpolating between prior and later months or years. In some cases, total pumpage for a system was available, but pumping rates for individual wells within the system were only available for a few years or not at all. In such cases, the total pumpage each year was divided between the wells based on available data, and the proportions held constant from year to year.

Part of the ground water pumped for public supply returns to the ground-water system through

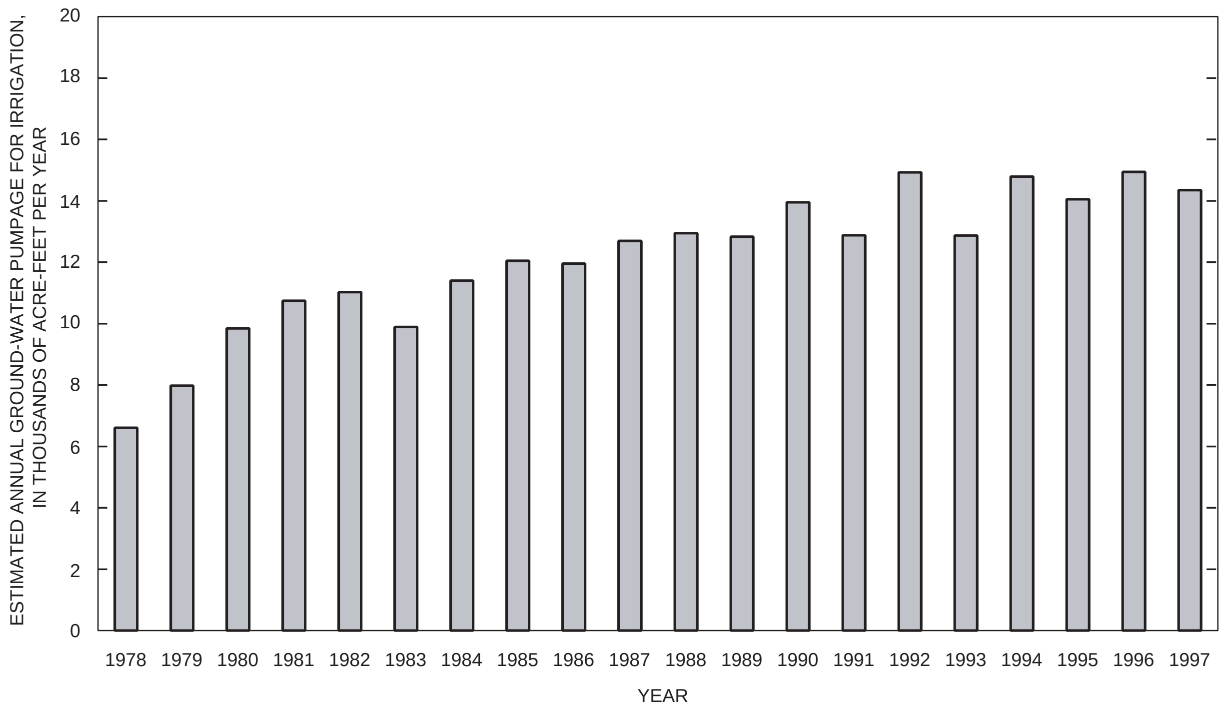


Figure 24. Estimated annual ground-water pumpage for irrigation in the upper Deschutes Basin, Oregon, 1978–97.

a variety of processes, such as seepage from sewage infiltration ponds, leakage from transmission lines, infiltration from on-site septic systems (drainfields), and deep percolation during irrigation. The fraction of public-supply pumpage not returned to the ground-water system through these processes is considered to be consumptively used. The proportion of the gross public-supply pumpage that is consumptively used is not precisely known. Because most of the water returned to sewage treatment plants is returned to the ground-water system, subtracting the volume of water delivered to these plants from the gross amount pumped from wells can provide an estimate of the amount of ground water that is consumptively used.

Measurements of ground-water pumpage and wastewater flow for the cities of Redmond and Bend provide information on the percentage of ground-water pumpage consumptively used. Monthly measurements for Redmond from 1988 to 1997 show that, depending on the month, 22 to 92 percent of the ground water pumped is returned to the sewage treatment plant as wastewater (Pat Dorning, City of Redmond, written commun., 1999). Return flows for the city of Bend are comparable to those of Redmond (Roger Prowell, City of Bend, oral commun., 1999). During winter, when water use is relatively low, 80 to 90 percent of the ground water pumped is returned as wastewater, and only 10 to 20 percent is unaccounted for. During summer, when water production is about four times the winter rate, only about 20 to 40 percent of the ground water pumped is returned as waste-water, leaving 60 to 80 percent unaccounted for. The water not returned as wastewater is not, however, all consumptively used. Part of the water not returned as wastewater returns to the ground-water system through leakage from supply and sewer lines. This type of leakage may account for as much as 8 percent of the total pumpage (Jan Wick, Avion Water Company, oral commun., 1999). A large amount of the increased water production during the summer is used for irrigation of lawns, gardens, and parks. Much of this water is used consumptively, lost through evaporation and transpiration by plants, but some percolates below the root zone and returns to the ground-water system. Because municipalities and urban home owners generally employ relatively efficient irrigation techniques such as sprinklers, as opposed to inefficient techniques such as flood irrigation, it is probably reasonable to assume that a large proportion of the increased summer production is used consumptively, but the exact amount in unknown.

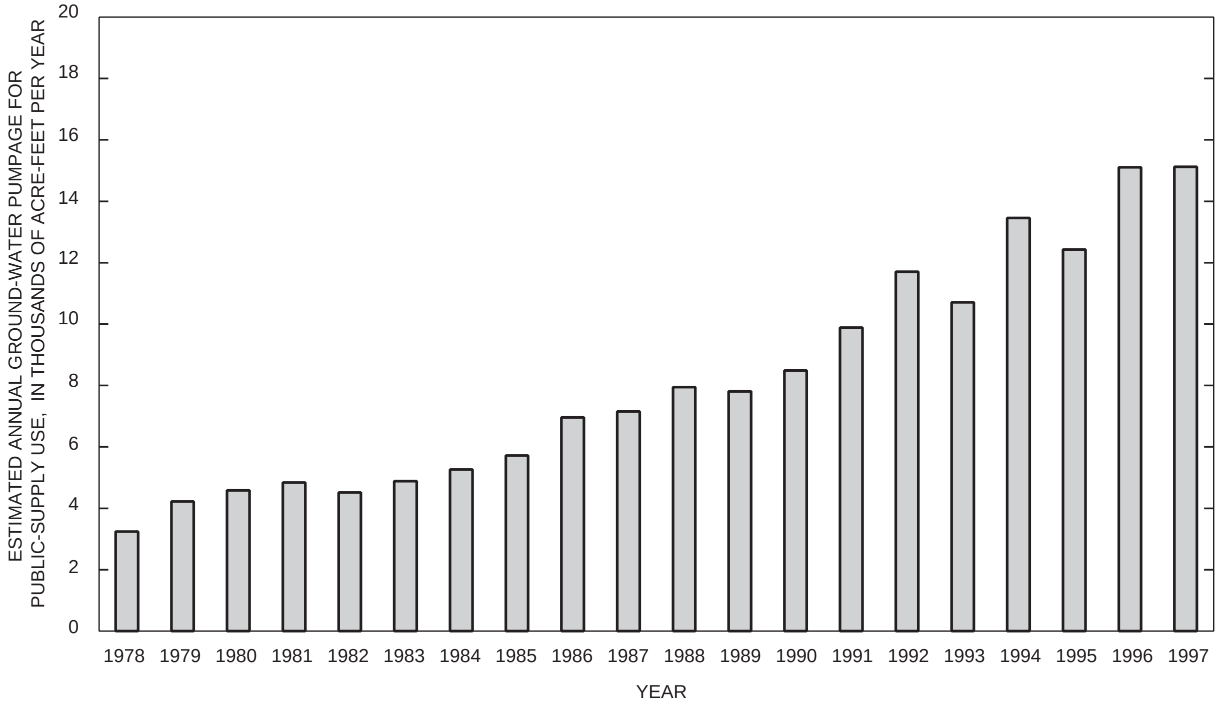


Figure 26. Estimated annual ground-water pumpage for public-supply use in the upper Deschutes Basin, Oregon, 1978–97. (Gross pumping figures do not represent actual consumptive use; a significant proportion of the pumped water returns to the ground-water system.)

Additional sources of error may be present in consumptive-use estimates based on wastewater return flow. In urban areas, some of the wastewater returned to sewage treatment plants is lost through evaporation from sewage lagoons or infiltration ponds. If sewage effluent is used to irrigate fields, a considerable amount may be lost through evapotranspiration. Consumptive-use estimates may be low if it is assumed that all the wastewater returned to sewage treatment plants is returned to the ground-water system.

Estimates of the proportion of ground-water pumpage that is actually consumed and not returned to the ground-water system are clearly influenced by many sources of error and must be considered approximate. Available data suggests that consumptive use ranges from approximately 10 percent of the total pumpage during winter, to approximately 50 to 70 percent during the high-water-use summer. On an annual basis, about 43 percent of the ground water pumped by the city of Redmond, for example, is returned as wastewater, leaving 57 percent of the water unaccounted for. Return-flow figures and transmission-loss estimates suggest that consumptive use of ground water in urban areas is probably somewhat less that 50 percent of the gross annual pumpage.

Private Domestic Wells

Not all residents of the upper Deschutes Basin are connected to public water supplies; many rely on private domestic wells. Private domestic well use was estimated using OWRD water-well-report files, data from the Oregon Health Division, Drinking Water Section (Dennis Nelson, written commun., 1999), population data from the State of Oregon (1999), and 1990 census data (U.S. Department of Commerce, 1993). As of 1995, an estimated 34,000 individuals, about 27 percent of the population of the study area, obtained water from private domestic wells or small water systems. The percentage of residents on private wells varies between counties. As of 1995, about 22,000 people, or 24 percent of the population, obtained water from private wells in Deschutes County. In Jefferson County, about 1,900 people, 12 percent of the population, relied on private wells. In Crook County, about 8,000 people, 52 percent of the population, obtained water from private wells. An estimated 1,900 people relied on private wells in Klamath County in the study area.

The amount of ground-water pumpage by private domestic wells can be roughly estimated based on number of individuals served by such wells.

Per capita water use in the upper Deschutes Basin, estimated by using data from public water-supply systems, varies considerably between systems. Records from public water suppliers indicate that average daily per capita water use for the largest public-supply systems in the study area ranges from 100 to 300 gal/d. Some of these systems supply commercial and municipal uses, and the per capita figures from them are not representative of rural dwellings. Many of the private wells in the study area are in rural residential areas served by irrigation districts, so well water is not used for irrigation of lawns and gardens. Because water from private domestic wells is used primarily for indoor use and not irrigation, per capita pumpage from rural residential domestic wells is considered for estimation purposes to be at the lower end of the calculated range, 100 gal/d.

If an average per capita pumpage of 100 gal/d is used, ground-water pumpage by private domestic wells (assuming 34,000 individuals are served) is approximately 3.4 million gal/d, which equals an average annual rate of 5.3 ft^3/s. As is discussed in the previous section, all of this water is not used consumptively. Virtually all of the homes on private domestic wells also use on-site septic systems, so most of the water pumped is returned to the ground-water system through drainfields. Actual consumptive use of ground water by private domestic wells in the upper Deschutes Basin is, therefore, likely less than 1 to 2 ft^3/s.

Ground-Water Discharge to Evapotranspiration

Most consumption of water by evapotranspiration occurs in the unsaturated zone. This water is intercepted as it percolates downward through the unsaturated zone prior to becoming ground water. Evapotranspiration from the unsaturated zone is accounted for by the DPM and occurs outside of the ground-water budget. Thus, the evapotranspiration of water from the unsaturated zone is not considered ground-water discharge. There are, however, circumstances in which evapotranspiration does consume ground water from the saturated zone. This occurs when the water table is sufficiently shallow to be within the rooting depth of plants, on the order of 5 to 10 ft deep. Evapotranspiration of water in this manner is considered ground-water discharge.

Broad areas with shallow ground-water conditions as described above are rare in the upper Deschutes Basin. The La Pine subbasin is the only significant large region in the study area with shallow ground-water conditions necessary for evapotranspiration from the water table. Areas of shallow ground water occur in the drainages of the upper Metolius River and Indian Ford Creek as well, but these are small in comparison to the La Pine subbasin. The potential amount of evapotranspiration from the water table in the La Pine subbasin was estimated to evaluate the significance of this process to the overall ground-water budget.

The DPM described earlier in this report calculated the amount of potential evapotranspiration throughout the study area. It also calculated the proportion of the potential evapotranspiration satisfied by evapotranspiration from the unsaturated zone. The proportion of the potential evapotranspiration not satisfied in this manner is the remaining amount that could be satisfied by evapotranspiration from the water table, and is termed the residual evapotranspiration. The DPM estimated that the residual evapotranspiration in the La Pine area equals an average annual instantaneous rate of about 5.7×10^{-8} ft/s (feet per second) (22 in./yr), which is equivalent to about 1.6 $ft^3/s/mi^2$. The probable area over which the water table is within 10 ft of land surface in the La Pine subbasin is estimated to be about 50 mi^2, based on water-level measurements in the La Pine subbasin taken in June 1999. During that time of year, the rate of evapotranspiration would be greatest. If the maximum residual evapotranspiration is lost to evapotranspiration over the entire 50 mi^2, it would represent an average annual rate of about 80 ft^3/s. To transpire at the full residual evapotranspiration rate, however, the water table would have to be virtually at land surface. In reality, the water table is probably near the margin of the rooting depth of plants, so the actual amount of evapotranspirative loss from the water table is probably much less than 80 ft^3/s. The values for evapotranspiration presented in this section are rough estimates, but serve to illustrate the magnitude of the probable ground-water discharge through evapotranspiration for comparison with other parts of the ground-water flow budget.

GROUND-WATER ELEVATIONS AND FLOW DIRECTIONS

Hydrologists describe the force driving ground-water movement as *hydraulic head*, or simply, *head.* Ground water flows from areas of high head to areas of low head. In an unconfined aquifer, such as a gravel deposit along a stream or a fractured lava flow near land surface, the elevation of the water table represents the head at the upper surface of the aquifer. Ground water flows in the direction the water table slopes, from high-elevation (high-head) areas toward low-elevation (low-head) areas. The change in head with distance, or *head gradient*, is simply the slope of the water table. Some aquifers, however, are confined by overlying strata with low permeability called *confining units.* A confined aquifer, for example, may be several hundreds of feet below land surface. The water in such an aquifer is often under pressure. When a well penetrates the aquifer, the water will rise in the well to some elevation above the top of the aquifer. The elevation to which the water rises is the head at that place in the aquifer. Water moves in confined aquifers from areas of high head to areas of low head just as in unconfined aquifers. Multiple confined aquifers can occur one on top of another separated by confining units. The heads in multiple confined aquifers may differ with depth resulting in vertical head gradients. If a well connects multiple aquifers with different heads, water can flow up or down the well from the aquifer with high head to the aquifer with low head. The distribution of head in an unconfined aquifer is represented by the elevation and slope of the water table. The distribution of head in a confined aquifer is represented by an imaginary surface known as a *potentiometric surface*. A potentiometric surface can be delineated by evaluating the static water-level elevations in wells that penetrate a confined aquifer.

In this report, the distinction between confined and unconfined aquifers is not critical to most of the discussion and is generally not made. The term *ground-water elevation* is used instead of head in the following discussion because it is more intuitively understandable. Furthermore, the term *water table* is used loosely to describe the general distribution of ground-water elevation in an area whether the aquifers are confined or unconfined. The important concept is that ground water moves from areas of high ground-water elevation (high head) to areas of low ground-water elevation (low head). In the upper Deschutes Basin, ground-water elevations are highest in the Cascade Range, the locus of ground-water recharge in the basin, and lowest in the vicinity of the confluence of the Deschutes, Crooked, and Metolius Rivers, the principal discharge area.

The geographic distribution of ground-water elevations in the upper Deschutes Basin was determined in this study using a variety of types of data. In the developed parts of the study area, primarily the areas of privately owned land, water-level elevations were determined by measuring water levels in wells. In some instances, conditions precluded measurements and water levels reported by drillers were used. Data from geothermal exploration wells provided a small amount of water-level information in the Cascade Range and at Newberry Volcano. Very few water wells exist in the vast tracts of public land that compose much of the upper Deschutes Basin. In those areas, the sparse water-well data was augmented with elevation data from large volume springs and gaining stream reaches. Major discharge features such as these represent points at which the water-table elevation and land-surface elevation coincide.

Horizontal Ground-Water Flow

In the upper Deschutes Basin, ground water moves along a variety of paths from the high-elevation recharge areas in the Cascade Range toward the low-elevation discharge areas near the margins of the Cascade Range and near the confluence of the Deschutes, Crooked, and Metolius Rivers. The generalized ground-water elevation map (fig. 28), based on hydraulic-head measurements in deep wells and on the mapped elevations of major springs and gaining stream reaches, shows the general direction of regional ground-water flow in different parts of the upper basin. The map is generalized and does not reflect local areas of shallow ground water caused by irrigation and canal and stream leakage.

In the southern part of the upper Deschutes Basin, ground water flows from the Cascade Range (including the Mt. Bachelor area) towards the high lakes area and the Deschutes and Little Deschutes Rivers in the La Pine subbasin. Ground water flows from Newberry Volcano toward the La Pine subbasin and toward the north. The water table in the La Pine subbasin is relatively flat, with an elevation of about 4,200 ft and a slight gradient generally toward the north-northeast. In this area the water table is shallow, often within several feet of land surface. North of Benham Falls, the gradient increases dramatically and the water table slopes steeply to the northeast. As a result, the regional water table, which is very close to land surface in the La Pine subbasin, is several hundred feet below land surface near Bend.

Ground-water elevations are relatively high in the southeast part of the Deschutes Basin near Millican, indicating that ground water flows from that area toward the northwest into the lower parts of the basin. As described previously, some water likely enters the southeastern part of the Deschutes Basin from the Fort Rock Basin (Miller, 1986). In the northern part of the study area, ground water flows from the Cascade Range to the northeast into the lower part of the basin toward ground-water discharge areas near the confluence of the Deschutes, Crooked, and Metolius Rivers.

In the central part of the study area, around Bend, Redmond, and Sisters, the water table is relatively flat between an elevation of 2,600 and 2,800 ft, although there is a gradual gradient to the north toward the confluence area (fig. 28). The water table in the Bend area is generally hundreds of feet below land surface. The northward slope of the water table is less than the northward slope of the land, however, so the water table is closer to land surface in the Redmond area. North of Redmond, the deep canyons of the Deschutes and Crooked Rivers are incised to the elevation of the regional water table, so ground water flows toward, and discharges to, streams that act as drains to the ground-water flow system. Water-level contours are generally parallel to the canyons in the confluence area, indicating flow directly toward the rivers.

A striking feature of the generalized water-table map (fig. 28) is the linear zone of closely spaced contours (indicating a high horizontal head gradient) that trends northwest-southeast across the upper basin. There are at least four possible explanations for this feature. First, the feature generally follows the topography. It also is likely related to the distribution of precipitation, which shows a similarly oriented high gradient region, particularly in the northern part of the mapped area. The flattening of the water-table surface to the northeast, which partly defines the high-gradient zone, is likely due to permeability contrasts related to the stratigraphy. The low-gradient area in the northeastern part of the map corresponds to that part of the Deschutes Formation where permeable fluvial deposits are an important component. Lastly, the linear zone could be, in part, an artifact of the geographic and vertical distribution of head data, particularly southeast of Bend where data are sparse. The northwest-trending high-head-gradient zone does not generally correspond with mapped faults.

Vertical Ground-Water Flow

Ground-water elevation (or head) can vary vertically as well as horizontally. At many locations, wells with different depths have different water levels. In recharge areas, where water enters the ground-water system, ground water generally moves downward and there is a downward head gradient (fig. 29). In recharge areas, water-level elevations are lower in deep wells and higher in shallow wells. If a well penetrates multiple aquifers in a recharge area, water can flow downward in the well from one aquifer to another. In areas where ground-water flow is primarily horizontal and there is little vertical movement of water, vertical gradients are small. In discharge areas, water from deep aquifers under pressure moves upward from depth and there is an upward head gradient. In discharge areas, deep wells have higher water-level elevations than shallow wells, and, if upward head gradients are sufficiently large, water levels in deep wells can be above land surface, causing water from the wells to flow at land surface.

Downward head gradients are common throughout much of the upper Deschutes Basin, including the Cascade Range and lower parts of the basin around Bend and Redmond. In the Cascade Range, the large amount of recharge causes downward movement of ground water and strong downward head gradients. Evidence of this downward flow in the Cascade Range is commonly seen in temperature-depth logs of geothermal wells (Blackwell, 1992; Ingebritsen and others, 1992). Temperature data show downward flow to a depth of at least 1,640 ft below land surface in an exploration well drilled near Santiam Pass (Blackwell, 1992). Similar large downward head gradients were observed in the Mt. Hood area in the Cascade Range north of the study area by Robison and others (1981).

Downward head gradients in the lower parts of the basin result primarily from artificial recharge from leaking irrigation canals. Ground-water elevations are artificially high in areas around networks of leaking irrigation canals. In some places, artificially high ground-water levels are observed only in scattered wells close to major canals. In other places, such as north and northwest of Bend, high ground-water elevations are maintained over a broad region by canal leakage. There are also isolated areas of shallow ground water that may be related to natural recharge from stream leakage.

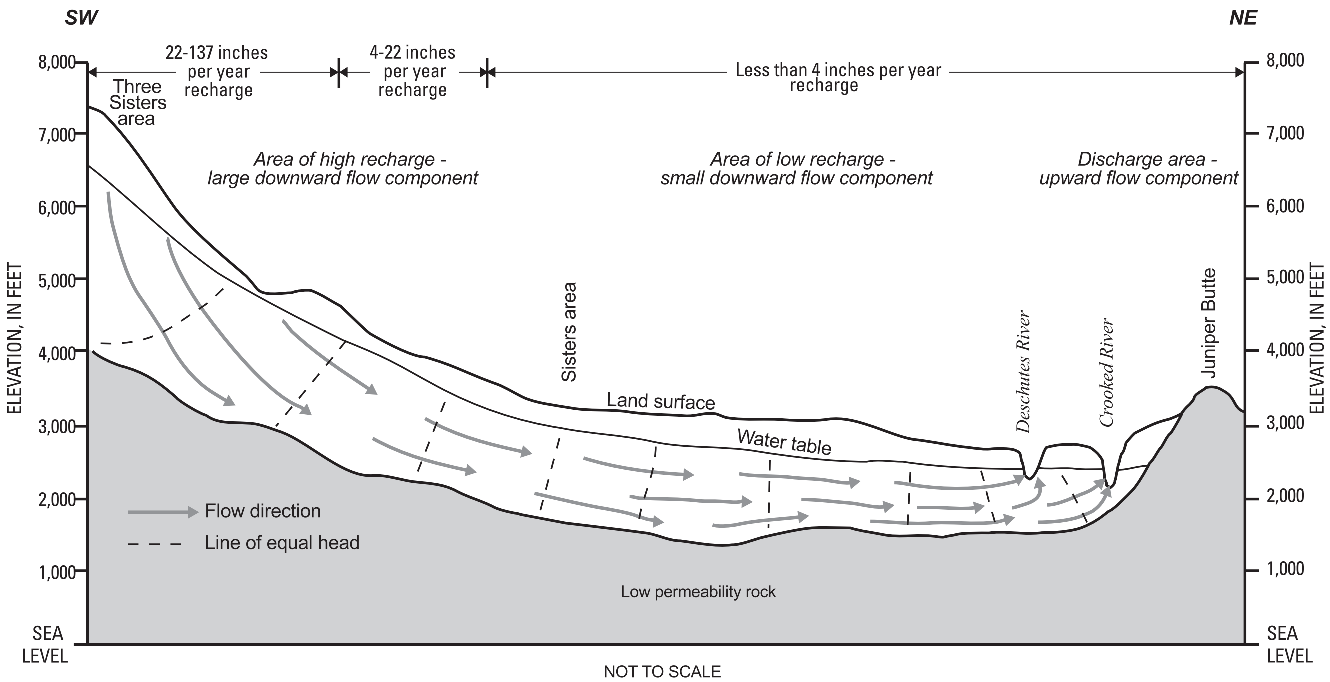


Figure 29. Diagrammatic section southwest-northeast across the upper Deschutes Basin, Oregon, showing flow directions and lines of equal hydraulic head.

Separate sets of water-level elevation contours for shallow wells (generally 100 to 300 ft deep) and deep wells (generally 500 to 900 ft deep) were drafted for the area around Bend, Redmond, and Sisters (fig. 30). In the area north and northwest of Bend, water-level elevations in shallow wells are 200 to 400 ft higher than water-level elevations in deep wells. At some locations, water levels in shallow and deep wells differ by over 500 ft. The shape and location of this area of high water levels suggests that it is caused by canal losses; for the most part it does not coincide with potential natural sources of recharge. Caldwell (1998) showed that shallow ground water is isotopically very similar to canal and stream water, which also suggests that canal and stream leakage are a principal source of recharge for shallow ground water.

There are isolated areas in the upper Deschutes Basin where anomalously high ground-water elevations likely result from natural causes. Such areas are present along the Deschutes River about halfway between Bend and Redmond (near Awbrey Falls) and west of Redmond. Elevated shallow water levels in these areas are likely caused by natural leakage from the Deschutes River. The relatively high shallow ground water in the Sisters area is also probably natural, as no significant source of artificial recharge is present.

Local recharge from leaking irrigation canals throughout the populated areas in the lower basin, and the resulting vertical head gradients, cause water-level elevations to vary from well to well in an area depending on the depth. In addition, water-level elevations can vary as the canals are turned on and off. Consequently, it can be difficult to accurately predict the depth to water at many locations, particularly where data from wells are sparse.

Upward head gradients are not commonly encountered in the upper Deschutes Basin. There are a number of possible causes for this. There is widespread artificial recharge from canal leakage and deep percolation of irrigation water throughout much of the populated area resulting in widespread downward gradients over most of the area where there are data. In addition, the streams to which most ground water discharges in the lower basin have cut deep into the aquifer system, allowing much of the water to discharge laterally without upward vertical movement. Finally, there are few wells that penetrate to depths below the elevation of streams in the major discharge area, where upward gradients would be expected.

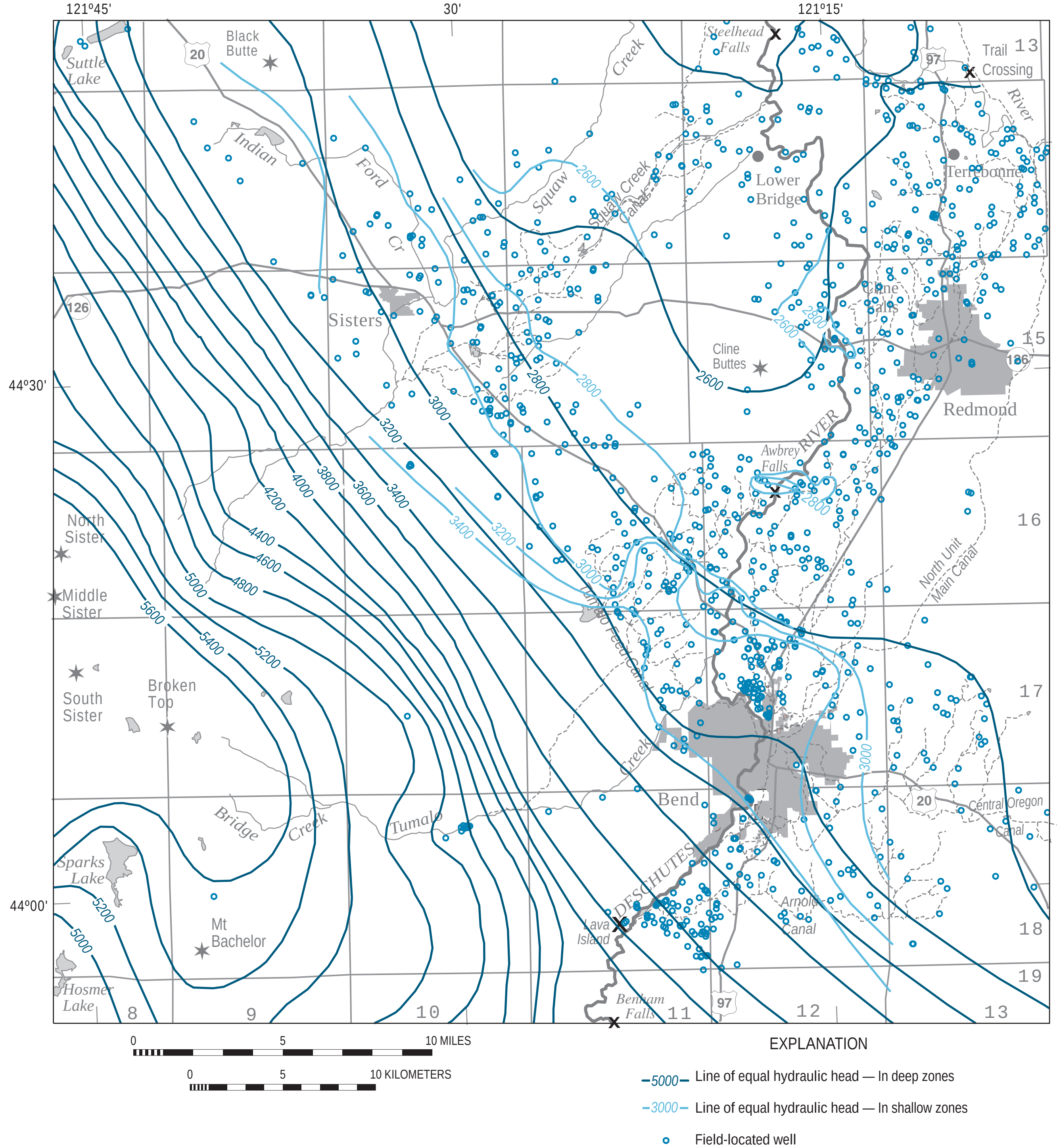


Figure 30. Generalized lines of equal hydraulic head for shallow and deep water-bearing zones in the central part of the upper Deschutes Basin, Oregon. (Elevated heads in shallow zones are due to infiltration of water from leaking irrigation canals, on-farm losses, and stream leakage.)

A substantial upward head gradient exists in the area of the lower Crooked River at depths below river level. A 740-ft well drilled near river level at Opal Springs had an artesian flow of 4,500 gal/min and a shut-in pressure of 50 pounds per square inch, indicating that the aquifer tapped by the well has a hydraulic head (water-level elevation) over 115 ft above the elevation of the river. This large upward gradient indicates upward ground-water flow toward the river.

FLUCTUATIONS IN GROUND-WATER LEVELS

The elevation of the water table is not static; it fluctuates with time in response to a number of factors, the most important of which are variations in recharge, canal operation, and pumping. In this section, ground-water-level fluctuations in the upper Deschutes Basin are described, the controlling factors identified, and the implications with regard to the regional hydrology are discussed.

Ground-water-level fluctuation data are collected by taking multiple water-level measurements in the same well over a period of time. Multiple water-level measurements are available for 103 wells in the upper Deschutes Basin. These wells were monitored for periods ranging from less than 1 year to more than 50 years; measurements were taken at intervals ranging from once every 2 hours (using automated recording devices) to once or twice a year. Fourteen wells in the basin have been monitored by OWRD for periods ranging from 9 to more than 50 years. Generally, measurements have been taken in these wells one to four times a year. Seventy-three wells were measured quarterly during this study for periods ranging from 1 to 4 years. Nineteen of these wells also were measured quarterly for 1 to 2 years during the late 1970s. Sixteen wells were instrumented with continuous recorders, devices that measured and recorded the water-level elevation every 2 hours. These short-interval measurements effectively create a continuous record of water-level elevation changes. Graphs of water-level fluctuations in all of these wells are published in the data report for this study (Caldwell and Truini, 1997).

Large-Scale Water-Table Fluctuations

The most substantial ground-water-level fluctuations in the upper Deschutes Basin, in terms of both magnitude and geographic extent, occur in and adjacent to the Cascade Range, including parts of the La Pine subbasin. These fluctuations are exemplified by the hydrographs of wells 21S/11E-19CCC, near La Pine, and 15S/10E-08ACD, near Sisters (fig. 31). The water level in both these wells fluctuates up to 20 ft with a cycle averaging roughly 11 years. A comparison of these water-level fluctuations with precipitation at Crater Lake in the Cascade Range (fig. 31) indicates that periods of high ground-water-level elevations generally correspond to periods of high precipitation, and low water-level elevations correspond to periods of low precipitation. This relation, of course, is to be expected. During periods of high precipitation, the rate of ground-water recharge exceeds, at least temporarily, the rate of discharge. When ground-water recharge exceeds discharge, the amount of ground water in storage must increase, causing the water table to rise. During dry periods, in contrast, the rate of discharge may exceed the rate of recharge, and ground-water levels drop as a result.

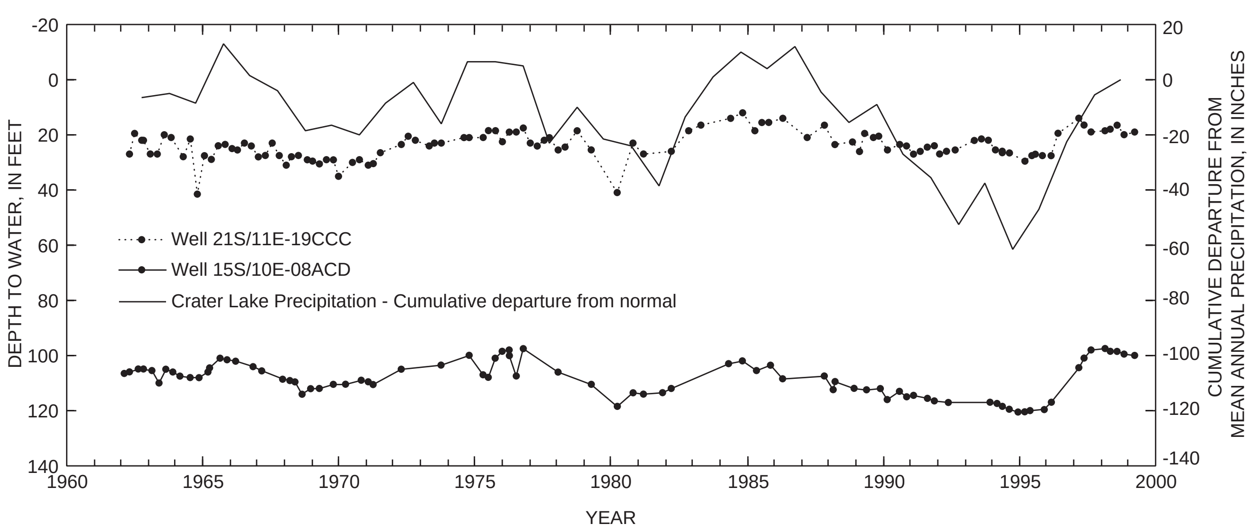


Figure 31. Static water levels in two long-term observation wells in the upper Deschutes Basin, Oregon, and cumulative departure from normal annual precipitation at Crater Lake, Oregon, 1962–98.

Fluctuations in the water-table elevation in response to variations in recharge are most prominent in the Cascade Range, the primary recharge area. A comparison of hydrographs of wells at varying distances from the Cascade Range (fig. 32) shows that as distance from the recharge area increases, the magnitude of fluctuations decreases, and the timing of the response is delayed.

During the period from 1993 through early 1999, ground-water levels in and near the Cascade Range, such as in wells 14S/9E-08ABA and 15S/10E-08ACD, rose over 20 ft in response to an abrupt change from drought conditions to wetter-than-normal conditions. Wells 15S/10E-36AAD2 and 15S/10E-02CDA, a few miles to the east of Sisters, farther away from the Cascade Range, showed a smaller rise in water level (less than 20 ft), and a slight delay in response. Well 14S/12E-09ACB several miles farther east near Lower Bridge, exhibited only a slight rise in water level, less than 2 ft, in response to the end of the drought, and an apparent delay in response. Long-term trends in wells with seasonal fluctuations, such as well 14S/12E-09ACB, are evaluated by comparing annual high and low water levels from year to year. Farther east near Redmond, water levels in wells 15S/13E-04CAB and 15S/13E-18ADD had barely stopped declining even 2 years after the end of the drought. Water levels in these wells had not started to rise as of early 1999.

Long-term records show that the water level in well 15S/13E-18ADD has fluctuated about 10 ft since 1971 compared to 23 ft in well 15S/10E-08ACD to the west closer to the recharge area (Caldwell and Truini, 1997, fig. 8). In addition, the decadal-scale peaks and troughs in the hydrograph of well 15S/13E-18ADD are broad and lag those of the well 15S/10E-08ACD by roughly 2 years.

The eastward-increasing delay in the water-level response to changes in recharge in the Cascade Range is depicted by a series of maps in figure 33. These maps show the annual direction of water-level change from March 1994 to March 1998 for observation wells throughout the upper basin. From March 1994 to March 1995, during the drought, water levels dropped in nearly all wells. Between March 1995 and March 1996, water levels in wells along the Cascade Range margin rose while water levels in wells to the east continued to decline. Over the next 2 years, the trend of rising water levels migrated eastward.

The attenuation and delay of water-level fluctuations with distance from the recharge source is analogous to the attenuation and delay in ground-water discharge peaks with increasing basin size, as discussed in the previous section. The effects of recharge variations are diffused with distance in the aquifer system.

Water-level fluctuations are attenuated with increasing depth as well as with increasing horizontal distance from the recharge area. This can be seen by comparing the hydrographs of wells 21S/11E-19CCC and 22S/10E-14CCA, which are about 5 miles apart in geographically similar settings in the La Pine subbasin (fig. 34). Well 21S/11E-19CCC is 100 ft deep and produces water from a sand and gravel deposit between a depth of 95 and 100 ft. Well 22S/10E-14CCA is 555 ft deep and taps water-bearing zones between 485 and 545 ft below land surface within a thick sequence of fine-grained sediment. The water level in the well 21S/11E-19CCC was declining until early 1995 when it started to rise in response to the end of drought conditions. The water level rose over 15 ft by early 1997 in a manner similar to wells close to the Cascade Range. The water level in well 22S/10E-14CCA, in contrast, declined until early 1996, and by 1999 had risen only about 7 ft in response to the end of drought conditions.

Local-Scale Water-Table Fluctuations

In addition to basinwide ground-water-elevation fluctuations, smaller-scale, localized water-table fluctuations occur. These more isolated water-table fluctuations are caused by varying rates of recharge from local sources, such as leaking streams and canals, and by ground-water pumping.

Water-level fluctuations due to irrigation canal leakage occur in many wells throughout the irrigated areas in the central part of the study area, with water levels rising during the irrigation season when canals are flowing and dropping when canals are dry. The magnitude of these annual fluctuations varies with the proximity of the well to the canal, the depth of the well, and the local geology. Annual fluctuations due to canal leakage of nearly 100 ft have been documented (see well 17S/12E-08ABD in Caldwell and Truini (1997), p. 20), although fluctuations in the range of 1 to 10 ft are more common.

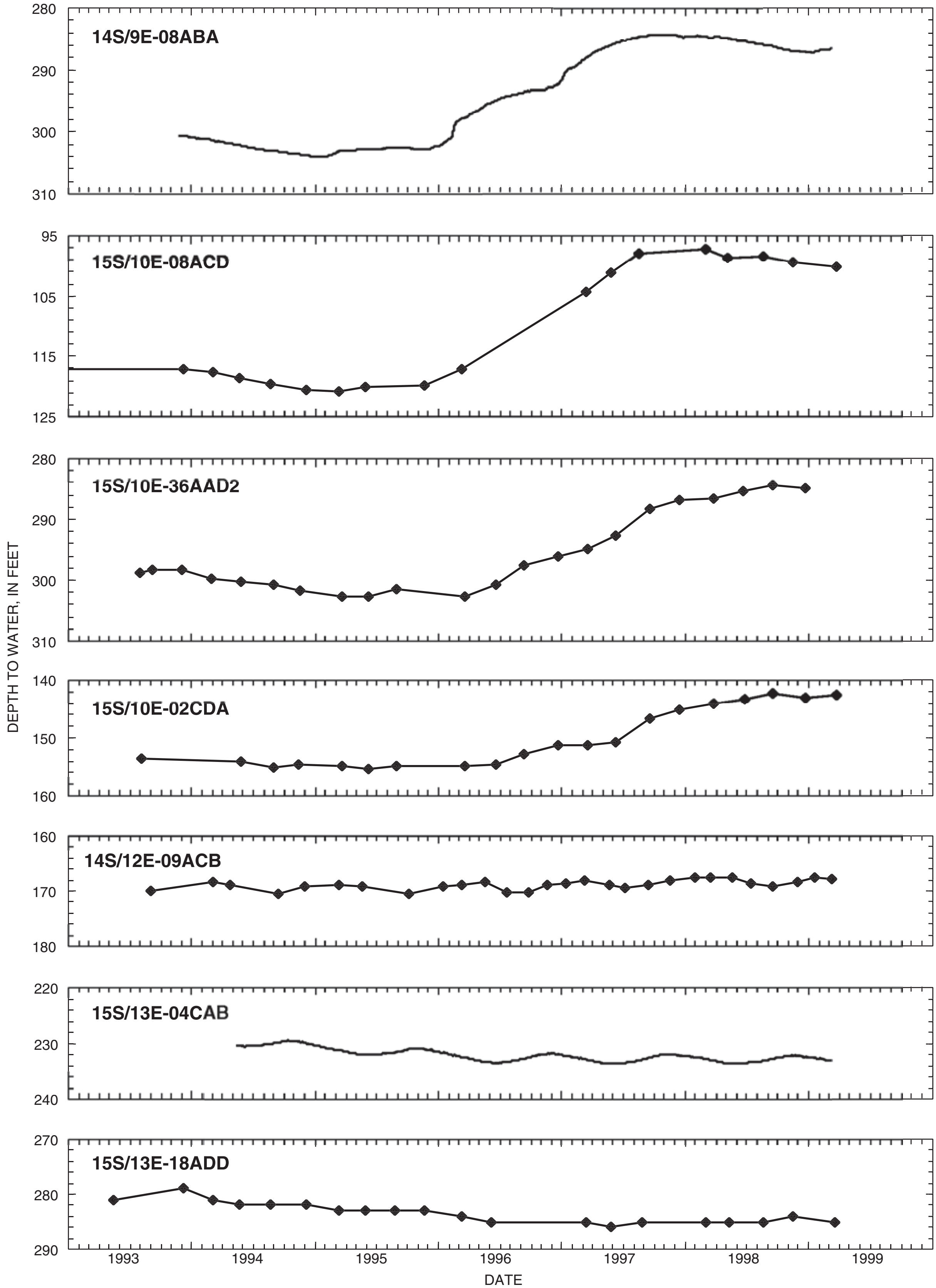


Figure 32. Variations in static water levels of selected wells at various distances from the Cascade Range, 1994–98. (The hydrographs show that the abrupt rise in water level in response to the change from drought conditions to wetter-than-normal conditions observed in the Cascade Range [uppermost hydrograph] is attenuated and delayed eastward out into the basin.)

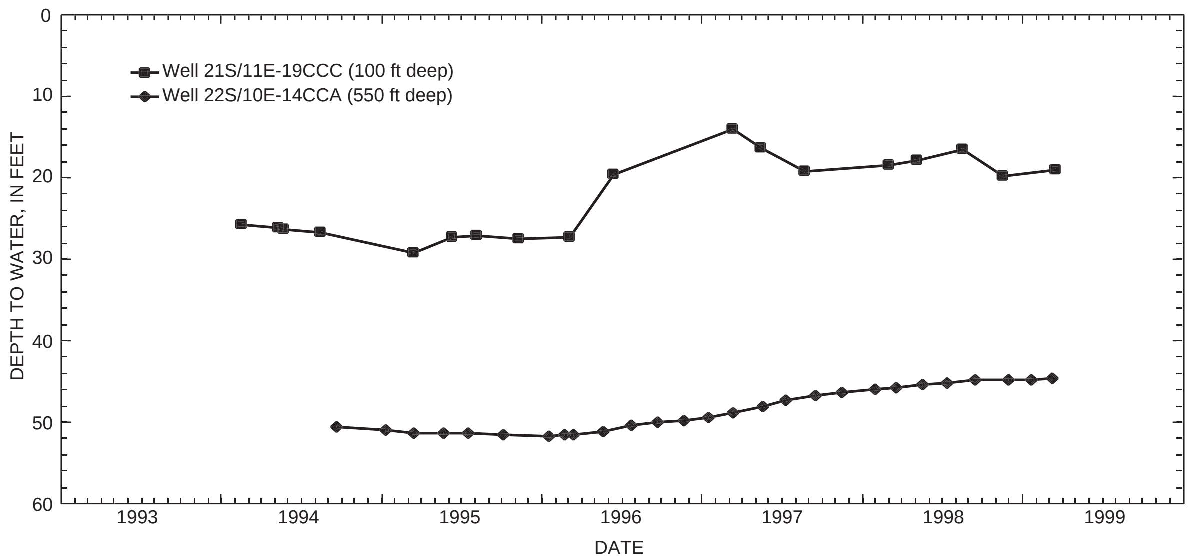


Figure 34. Static water-level variations in a shallow well and a deep well in the La Pine subbasin, Oregon.

Ground-water levels can respond rapidly to canal leakage, even at considerable depths, particularly in areas where fractured lava dominates in the subsurface. The water level in well 18S/12E-03DDC responds in a matter of days to the operation of main irrigation-diversion canals, which are about one-half mile away (fig. 35). The water level in this well starts to rise shortly after the canals start flowing and starts to drop soon after they are shut off for the season, peaking late in the irrigation season. In addition, the water table responds to periods of short-term operation of the canal, typically for several days during the winter for stock watering. The static water level in well 18S/12E-03DDC is over 600 ft below land surface, and the shallowest wells in the area have water levels of 300 to 400 ft below land surface. The rapid response of the water table to canal leakage at such depth is likely due to rapid downward movement of water through interconnected vertical fractures in the lava flows.

Water-table fluctuations can be more subdued and delayed in areas underlain by sedimentary materials where there are no vertical fractures and there is more resistance to downward movement of water. Well 15S/13E-04CAB (fig. 36) shows an annual water-level fluctuation that differs substantially from that of well 18S/12E-03DDC (fig. 35). The amount of fluctuation is somewhat less and the hydrograph is smooth, nearly sinusoidal, reflecting no short-term effects due to winter stock runs. In addition, the annual peak water level in well 15S/13E-04CAB, which occurs in October or November, is much later than that of well 18S/12E-03DDC, which occurs in August or September. The hydrograph of well 15S/13E-04CAB in figure 36 also shows a year-to-year decline in water levels due to drought effects superimposed on the annual fluctuations.

Water levels are affected by variations in streamflow as well as canal operation. In areas where stream elevations are above the adjacent ground-water elevations, streams typically lose water to the ground-water system due to leakage through the streambed. In some areas, the rate of stream leakage is not constant, but varies with streamflow. As streamflow increases and the elevation of the stream rises, a larger area of the stream bed is wetted providing a larger area through which water can leak.

The most substantial stream losses measured in the basin occur along the Deschutes River between Sunriver and Bend, where the river loses, on average, about 113 ft^3/s (fig. 12). The amount of loss is known to be stage-dependent and to vary with streamflow (fig. 13). This means that the ground-water recharge in the vicinity of the Deschutes River between Benham Falls and Bend varies with streamflow as well.

The variations in local recharge caused by changes in streamflow cause water-level fluctuations in some wells between Benham Falls and Bend (fig. 37). The stage and discharge in the Deschutes River in this reach is controlled by reservoir operations upstream. Streamflow is highest from April to October as water is released from the reservoirs to canal diversions near Bend.

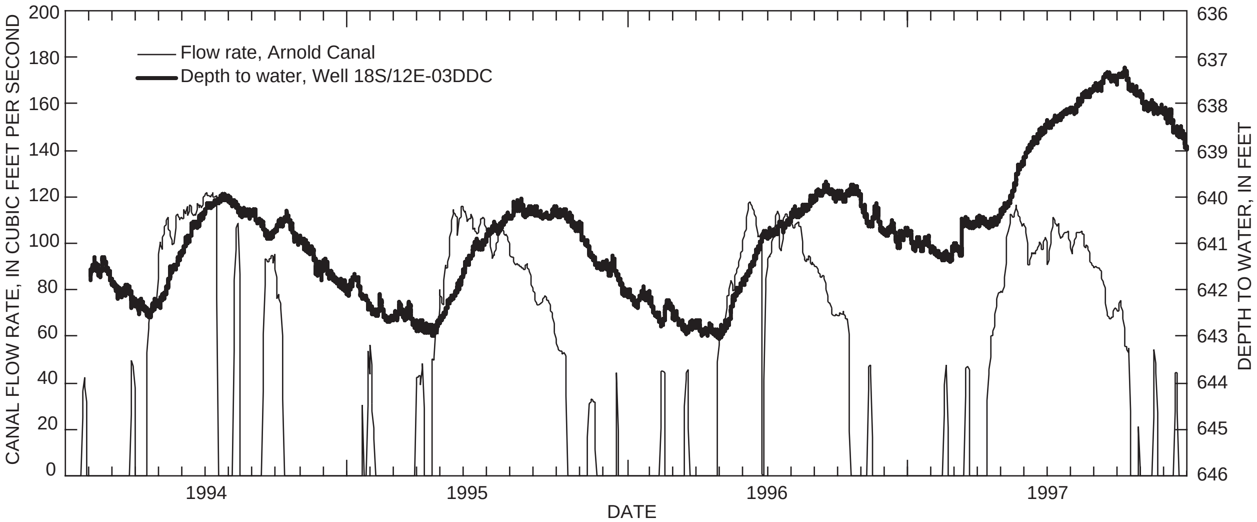


Figure 35. Relation between static water-level variations in a deep well near Bend, Oregon, and flow rate in a nearby irrigation canal.

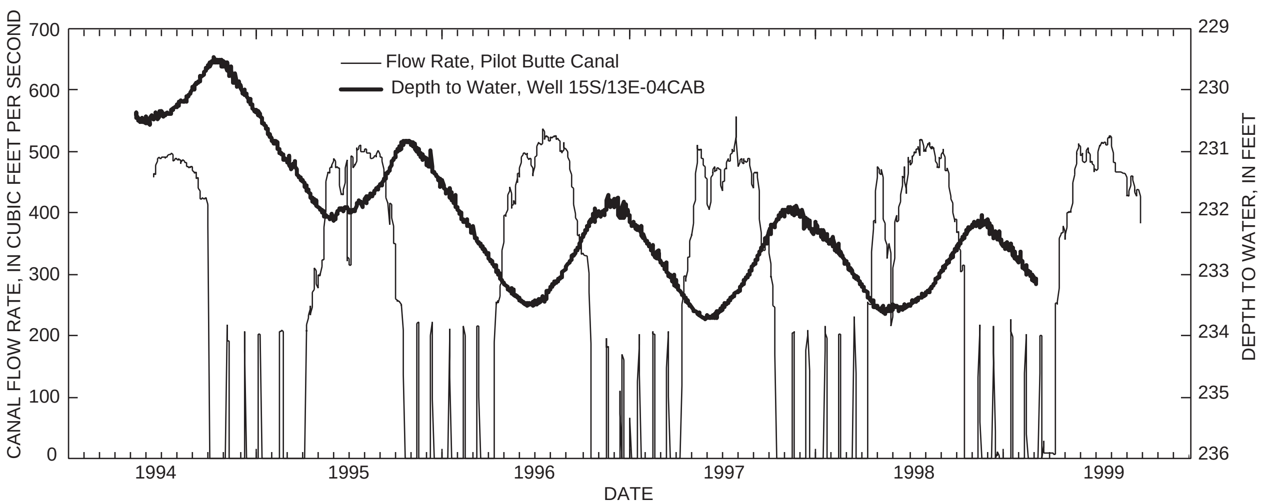


Figure 36. Relation between static water-level variations in a well near Redmond, Oregon, and flow rate in a nearby irrigation canal.

As a result, changes in streamflow (and stage) can be relatively abrupt. The water level in well 19S/11E-16ACC, about 500 ft from the river near the Benham Falls gage, rises and falls in response to river stage (fig. 37). Abrupt changes in streamflow usually manifest in the well within a few to several days. These effects are much less pronounced, however, in wells farther from the river. The water level in well 18S/11E-21CDD, about 1 mile from the river, also fluctuates in response to river stage, but the fluctuations are subdued and the hydrograph is nearly sinusoidal, showing only the slightest inflections in response to abrupt changes in streamflow. In addition, the peaks and troughs in the hydrograph of well 18S/11E-21CDD lag those of well 19S/11E-16ACC and river stage by 1 to 2 months.

The relation between ground-water levels and streamflow is apparent in ground-water discharge areas as well as in recharge areas; however, the process is reversed. In areas of losing streams (recharge areas), streamflow variations can cause water-table fluctuations as described in the previous paragraph. In ground-water discharge areas, however, water-table fluctuations cause variations in streamflow. This is illustrated by comparing a graph of the discharge of Fall River, a spring-fed stream, with a graph of typical long-term water-table fluctuations at the Cascade Range margin as seen in well 15S/10E-08ACD (fig. 38).

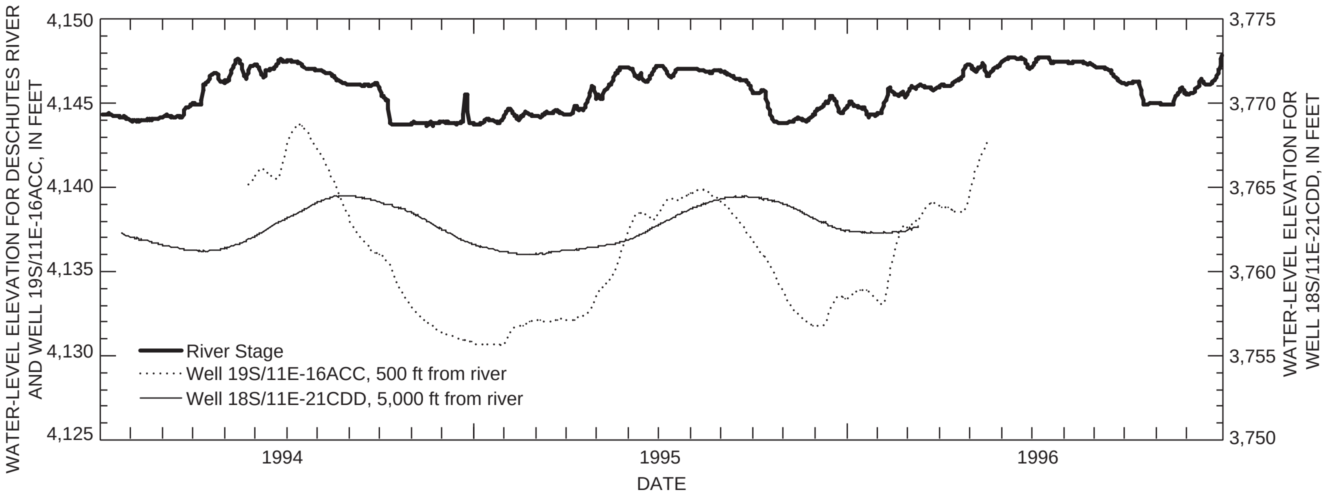


Figure 37. Relation between static water-level variations in two wells at different distances from the Deschutes River and stage of the river at Benham Falls.

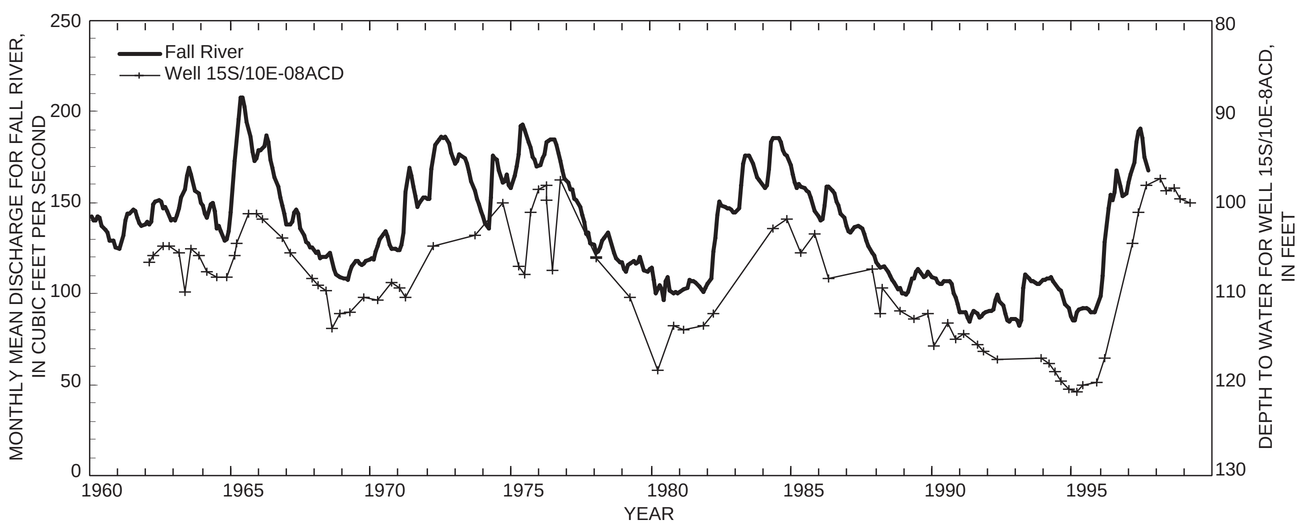


Figure 38. Relation between monthly mean discharge of Fall River and static water-level variation in a well near Sisters, Oregon, 1962–97.

It can be seen that spring flow increases during periods when the water table is high, and decreases when the water table is low. This process works on a larger scale to cause the temporal variations in ground-water discharge to major streams described previously.

Water-table fluctuations can be caused by ground-water pumping as well as by variations in recharge. When a well is pumped, the water table in the vicinity of the well is lowered due to the removal of ground water from storage. A conical depression centered around the well develops on the water table (or potentiometric surface in the case of a confined aquifer) and expands until it captures sufficient discharge and/or induces enough new recharge to equal the pumping rate. After pumping ceases, the water table recovers as the aquifer returns to pre-pumping conditions. Key factors that determine the magnitude of water-table fluctuations caused by pumping are the aquifer characteristics, the rate and duration of pumping, the presence of aquifer boundaries, and the number of wells. In aquifers that have low permeability, pumping-induced water-table fluctuations can be large and even interfere with the operation of other wells. If the long-term average pumping rate exceeds the rate at which the aquifer can supply water, water levels will not recover fully and long-term water-level declines will occur.

Water-table fluctuations caused by ground-water pumping are apparent in only a few of the wells monitored in the upper Deschutes Basin. Pumping effects appear to be small (less than a few feet of drawdown), seasonal in nature, and of limited geographic extent. No long-term water-level declines caused by pumping are apparent in any of the data.

Nearly all of the wells that were measured quarterly and that show annual fluctuations have high water levels during or shortly after the irrigation season, indicating that the water-table fluctuation is caused by canal leakage. A few of the wells that were measured quarterly show low water levels during the summer, suggesting a possible influence from irrigation pumping, but the small number of water-level measurements prevents any definite conclusions. These occurrences are not widespread.

Of the 16 wells that had continuous water-level recorders, pumping effects are apparent only in well 14S/12E-09ACB in the Lower Bridge area (fig. 39). This unused well shows an annual cycle in which the water level drops during the irrigation season, from about April to about September, and then rises during the off season. The annual variation is approximately 2 to 3 ft. The shape of the hydrograph of this well indicates drawdown and recovery most likely due to pumping of an irrigation well about a mile away. Although irrigation pumping causes a seasonal water-level decline in this well, there is no evidence of any long-term water-level decline. The only obvious long-term water-level trend seen in the well is the basinwide trend related to climate cycles. The lack of any apparent long-term pumping effects in this well is significant, because the Lower Bridge area contains the highest concentration of irrigation wells in the basin.

Water levels in the two other centers of ground-water pumping in the basin, the Bend and Redmond areas, show no apparent influence from ground-water pumping. Large amounts of ground water are pumped in both of these areas for public water-supply use, yet no pumping-related seasonal or long-term trends are apparent in observation well data. Any pumping influence is likely small due to the high aquifer permeability, and is undetectable due to the masking effects of canal leakage and climate-driven water-level fluctuations.

Ground-water levels in part of Jefferson County rose dramatically in response to the filling of Lake Billy Chinook behind Round Butte Dam in 1964. Water levels in two wells (11S/12E-21ABB and 11S/12E-26AAC) monitored by Portland General Electric, on opposite sides of the dam and about a mile away, rose approximately 120 and 100 ft, respectively, within about 10 years of filling of the reservoir (fig. 40). Because these are the only two wells monitored in the area with records extending back to the time prior to the filling of the reservoir, the full extent and magnitude of the effects of the reservoir are not clearly known. A comparison of water-level elevations mapped by Stearns (1931) with those mapped during this study (fig. 28) suggests that water levels have risen as much as 100 ft over a fairly large region from Round Butte, south to Juniper Butte, and extending east as far as Highway 97. Increases in water-level elevation were likely even greater close to the reservoir. No data are available to evaluate the probable water-level rise west and north of the reservoir, but water levels were almost certainly similarly affected. Water levels appear to have risen north of Round Butte in the vicinity of Lake Simtustus as well, but data are sparse and the magnitude and extent of any water-level rise are unknown. Although data are scarce, water levels appear not to have been affected as far north and east as Madras. A 1953 water-level measurement in one of the city of Madras water-supply wells is comparable to measurements made recently, long after the effects of Lake Billy Chinook should have been apparent.

Some of the wells in Jefferson County show an anomalous rising water-level trend that appears to have started in the mid-1980s. The hydrograph of well 11S/12E-26AAC (fig. 40) shows that the water level appeared to have largely stabilized in response to the filling of Lake Billy Chinook by the mid 1970s, but then started an upward trend beginning about 1985, rising over 20 ft since that time. Of the four other wells in the vicinity with sufficient record, two do not show this recent rising trend (fig. 40, well 11S/12E-21ABB), and two show water level rises of approximately 2 and 6 ft. This local water-table rise is an enigma in that it occurs during a period when water levels were dropping throughout much of the upper basin as a result of drought. There are no apparent changes in irrigation practices or canal operations

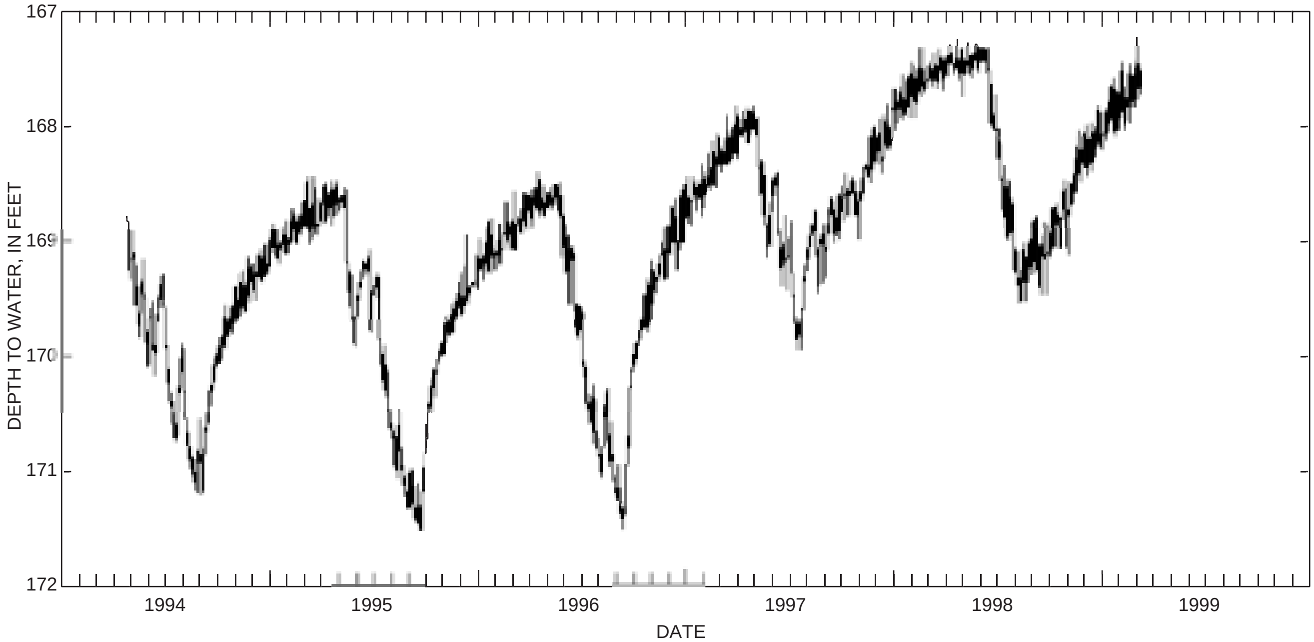


Figure 39. Static water level in an unused irrigation well near Lower Bridge (14S/12E-09ACB), showing seasonal pumping effects from nearby irrigation wells and long-term climatic effects.

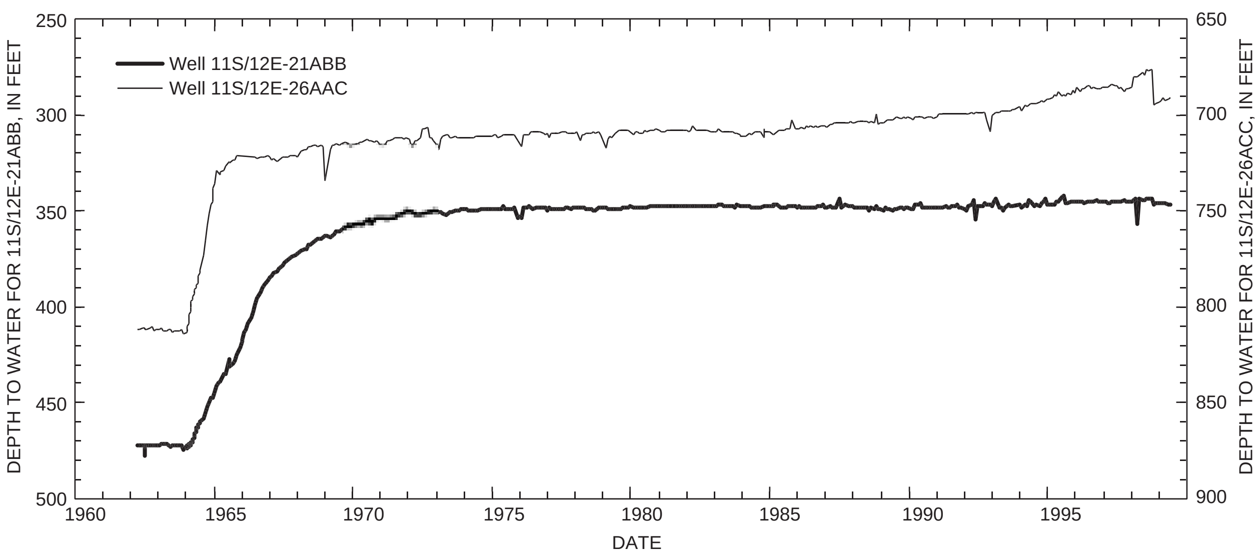


Figure 40. Water levels in two wells near Round Butte Dam, showing the rise in ground-water elevations caused by the filling of Lake Billy Chinook.

that could account for the observed upward trend. Water levels in wells in the Madras area rose after the city changed their primary source of water from wells to Opal Springs and greatly reduced their ground-water pumping, but this occurred in 1987, 2 years after the water level appears to have started to rise in well 11S/12E-26AAC (fig. 40). Although not entirely coincident, this reduction in pumping may have contributed to the observed water-level rise. It is also possible that the rise is a boundary effect related to the filling of Lake Billy Chinook, implying that the ground-water system is not yet in equilibrium with the reservoir even though water levels appeared to have stabilized in the late 1970s.

SUMMARY AND CONCLUSIONS

Regional ground-water flow in the upper Deschutes Basin is primarily controlled by the distribution of recharge, the geology, and the location and elevation of streams. Ground water flows from the principal recharge areas in the Cascade Range and Newberry Volcano, toward discharge areas along the margin of the Cascade Range and near the confluence of the Deschutes, Crooked, and Metolius Rivers.

At the regional scale, distribution of recharge mimics that of precipitation. The annual precipitation rate shows considerable geographic variation throughout the upper Deschutes Basin. The Cascade Range, which constitutes the western boundary of the basin, locally receives in excess of 200 inches per year, mostly as snow. The central part of the study area, in contrast, typically receives less than 10 inches per year. The young Quaternary volcanic deposits and thin soils in the Cascade Range allow rapid infiltration of much of the rain and snowmelt, making the Cascade Range the locus of ground-water recharge for the basin. The average annual rate of recharge from precipitation basinwide (1962–97) is about 3,800 ft^3/s (cubic feet per second). Precipitation provides relatively little ground-water recharge in the low-elevation areas in the central part of the basin; however, leaking irrigation canals are locally a significant source of recharge. It is estimated that 46 percent of the water diverted for irrigation is lost through canal leakage. The average annual rate of leakage from irrigation canals during 1994 was estimated to be 490 ft^3/s. Part of the ground water recharged in the Cascade Range discharges to spring-fed streams at lower elevations in the range and along margins of adjacent lowlands. The remainder of the ground water continues in the subsurface toward the central part of the basin, where most of it discharges to the Deschutes, Crooked, and Metolius Rivers in the vicinity of their confluence.

Most ground water in the upper Deschutes Basin flows through Neogene and younger deposits of the Cascade Range and Deschutes Formation. The underlying late Eocene to early Miocene deposits of the John Day Formation and the hydrothermally altered rocks at depth beneath the Cascade Range generally have very low permeability and are neither a significant source of ground water nor a medium through which it can easily flow. These older rocks crop out along the northern and eastern margins of the study area and underlie much of the upper basin at depth. Low-permeability rock units constitute the lower, northern, and eastern boundaries to the regional flow system.

The interaction between ground water and streams is controlled largely by the relative elevations of the water table and adjacent streams. In the La Pine subbasin, south of Benham Falls, the water-table elevation is near land surface. Stream gains and losses along most of the Deschutes and Little Deschutes Rivers in this area are small, indicating relatively little net exchange between ground water and surface water. North of Benham Falls, the northward slope of the water table is larger than the slope of the land surface, so depths to ground water increase northward toward Bend. In the central and eastern parts of the study area, ground-water elevations are typically hundreds of feet below the elevations of streams. Although ground-water levels are considerably below stream elevations in this area, streams do not lose appreciable amounts of water, because streambeds have been largely sealed by infiltration of fine sediment. One notable exception is the Deschutes River, which loses on average approximately 113 ft^3/s between Sunriver and Bend, likely into the youthful Holocene basalt erupted from Lava Butte.

The Deschutes and Crooked Rivers have incised canyons in the northern part of the study area. The canyons become increasingly deep northward toward Lake Billy Chinook, reaching depths of several hundred feet below the surrounding terrain. About 10 to 15 miles above their confluence, the canyons of the Deschutes and Crooked Rivers are of sufficient depth to intersect the regional water table, and both streams gain flow from ground-water discharge. Seepage runs show that the Deschutes River and lower Squaw Creek combined gain about 400 ft^3/s from ground-water discharge in this area prior to entering Lake Billy Chinook, and the lower Crooked River gains about 1,100 ft^3/s before entering the lake. Ground-water discharge to Lake Billy Chinook is roughly 420 ft^3/s. The total ground-water discharge in the confluence area is approximately 2,300 ft^3/s. This ground-water discharge, along with the flow of the Metolius River (which is predominantly ground-water discharge during the dry seasons), makes up virtually all the flow of the Deschutes River at Madras during the summer and early fall.

Geologic factors are the primary cause of the large ground-water discharge in the confluence area. The permeable Neogene deposits, through which virtually all regional ground water flows, become increasingly thin northward as the low-permeability John Day Formation nears the surface. The John Day Formation is exposed in the canyon of the Deschutes River about 10 miles north of Lake Billy Chinook near Pelton Dam, marking the northern extent of the permeable regional aquifer system. Most of the regional ground water in the upper basin discharges to the Deschutes and Crooked Rivers south of this location. There is no appreciable ground-water discharge directly to the Deschutes River downstream of this point, and the small gains in streamflow that do occur result primarily from tributary inflow.

Geological evidence and hydrologic budget calculations indicate that virtually all ground water not consumptively used in the upper Deschutes Basin discharges to the stream system upstream of the vicinity of Pelton Dam. Moreover, virtually the entire flow of the Deschutes River at Madras is supported by ground-water discharge during the summer and early fall. Ground water and surface water are, therefore, directly linked, and removal of ground water will ultimately diminish streamflow.

Analysis of the fluctuations of water-table elevations and ground-water discharge rates in response to stresses on the ground-water system, such as canal operation, stream-stage variation, and climate cycles, indicates that the effects of such stresses are delayed and attenuated with distance. The effects of ground-water pumping can be expected to be attenuated and delayed in a similar manner and spread out over time and space. Depending on the location of a well, several years may pass between the time pumping starts and the time the effects of the pumping are reflected in diminished discharge. It is important to note that the same physical processes that delay the onset of the effects of pumping on the streams also cause those effects to linger after pumping ends. So several years may also pass between the time pumping stops and the time the effects on streamflow end.

Presently, the effects of pumping cannot be measured below the confluence of the Deschutes, Crooked, and Metolius Rivers. The total consumptive use of ground water in the upper Deschutes Basin as of the mid-1990s is estimated to be about 30 ft^3/s: 20 ft^3/s for irrigations and 10 ft^3/s for public water supplies (assuming 50 percent of public-supply pumpage is consumptively used). Streamflow at the Madras gage, which is largely ground-water discharge during the summer, is about 4,000 ft^3/s. Streamflow measurement techniques used at the gage have an accuracy of +/– 5 percent, resulting in a range of error of about +/– 200 ft^3/s. Because total estimated consumptive ground-water use is less than 1 percent of the ground-water discharge at Madras, it is well within the expected range of measurement error. The amount of ground-water use also is small compared to the observed natural fluctuations in ground-water discharge.

Streamflow in the Deschutes Basin fluctuates dramatically at a variety of time scales due to many factors, including runoff variations, reservoir and canal operation, and climate cycles. The ground-water component of streamflow also fluctuates widely. For example, August mean ground-water discharge to the Deschutes River between Bend and Culver varied over 100 ft^3/s between 1962 and 1997 due to climate cycles. The August mean flow of the Crooked River below Opal Springs, which is mostly ground-water discharge, varied 460 ft^3/s during the same period. Ground-water discharge to the Metolius River, based on October mean flows, varied over 400 ft^3/s from 1962 to 1997. Combined, these climate-driven ground-water discharge fluctuations could account for variations in late-season monthly mean flows of the Deschutes River at Madras on the order of 1,000 ft^3/s. Natural fluctuations of ground-water discharge of this magnitude in the confluence area totally mask the effects of ground-water withdrawal at present levels of development.

Although the effects of historic ground-water pumping cannot be measured below the confluence area, the effects of canal leakage are easily discernible in the streamflow records. The August mean flows of the lower Crooked River increased between the early 1900s and the early 1960s by roughly 400 to 500 ft^3/s in a manner that paralleled the increase in estimated canal leakage north of Bend during the same period. The correlation indicates that a large proportion of the water lost from leaking irrigation canals north of Bend is discharging to the lower Crooked River upstream of the Opal Springs gage. This is consistent with the hydraulic-head distribution and ground-water flow directions in the area.

Although the effects of historic ground-water pumping on streamflow cannot be discerned in the streamflow record below the confluence area, it is possible that such effects could be measurable on smaller streams in the upper Deschutes Basin. Most tributary streams emanating from the Cascade Range, such as Fall River, Squaw Creek, and Indian Ford Creek, are either spring fed or otherwise hydraulically connected to the ground-water system. The ground-water discharge to these streams, and consequently streamflow, could be diminished to a measurable degree depending on the amount of ground-water pumping and the proximity of pumping to the stream. Long-term streamflow records, however, are not available to assess possible effects of historic ground-water development on smaller streams. Streamflow records are available for only a small number of tributary streams in the upper Deschutes Basin, and the gages that are operated are generally not in locations where the impacts of ground-water pumping are likely to be detected given the present geographic pattern of development.

Some stream reaches, for example the Deschutes River between Bend and Lower Bridge, are perched above the ground-water system. Although leakage from such streams can provide recharge to the ground-water system, the rate of leakage is independent of ground-water elevation changes. Therefore, ground-water pumping will have little or no affect on the rate of leakage along such reaches.

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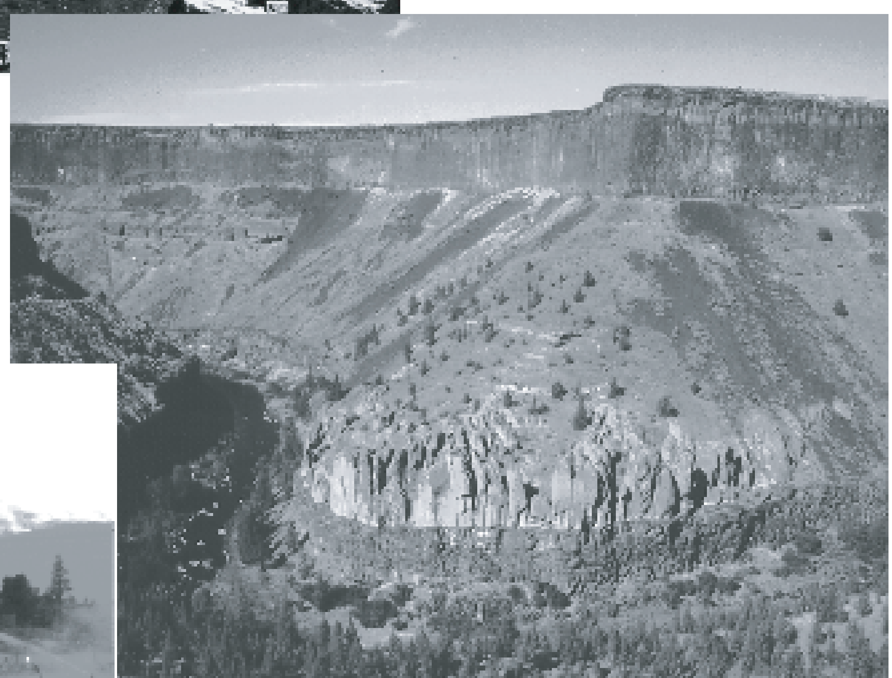
Ground-Water Hydrology of the Upper Deschutes Basin, Oregon



U.S. DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY
Water-Resources Investigations
Report 00-4162

Prepared in cooperation with
OREGON WATER RESOURCES DEPARTMENT;
CITIES OF BEND, REDMOND, AND SISTERS;
DESCHUTES AND JEFFERSON COUNTIES;
THE CONFEDERATED TRIBES OF THE
WARM SPRINGS RESERVATION OF OREGON;
and U.S. ENVIRONMENTAL PROTECTION AGENCY

Cover photographs:

Top: Steelhead Falls on the Deschutes River near Crooked River Ranch, Oregon.

Middle: Crooked River Canyon at Crooked River Ranch, Oregon.

Bottom: North and Middle Sister with a wheel-line irrigation system in the foreground near Sisters, Oregon. (Photographs by Rodney R. Caldwell, U.S. Geological Survey.)