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Hydrogeologic Framework of the Yakima River Basin Aquifer System, Washington

Scientific Investigations Report 2009–5152

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

Cover: Photograph of Clark Well No. 1, located on the north side of the Moxee Valley in North Yakima, Washington. The well is located in township 12 north, range 20 east, section 6. The wall was drilled to a depth of 940 feet into an artesian zone of the Ellensburg Formation, and completed in 1897 at a cost of $2,000. The original flow from the well was estimated at about 600 gallons per minute, and was used to irrigate 250 acres in 1900 and supplied water to 8 small ranches with an additional 47 acres of irrigation. (Photograph was taken by E.E. James in 1897, and was printed in 1901 in the U.S. Geological Survey Water-Supply and Irrigation Paper 55.)

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By J.J. Vaccaro, M.A. Jones, D.M. Ely, M.E. Keys, T.D. Olsen, W.B. Welch, and S.E. Cox

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KEN SALAZAR, Secretary

U.S. Geological Survey
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U.S. Geological Survey, Reston, Virginia: 2009

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Contents

Abstract 1
Introduction 2
Purpose and Scope 2
Description of Study Area 4
    Location and Setting 4
    Development of Water Resources 4
    Overview of the Geology 7
Well Information 10
    References to Wells 10
    Well Data and Their Use 10
Hydrogeologic Units 10
Hydraulic Characteristics of Units 20
    Lateral Hydraulic Conductivity 20
        Basin-Fill Units 20
        Columbia River Basalt Group Units 20
        Bedrock Units 25
        Estimates from Specific-Capacity Data 25
        Estimates from Aquifer Tests 28
    Vertical Hydraulic Conductivity 28
        Basin-Fill Units 28
        Columbia River Basalt Group Units 29
        Bedrock Units 29
    Storage Coefficient 29
Hydrochemistry 30
    Variation in Major-Ion Concentration 31
    Deuterium and Oxygen-18 31
    Methane 33
    Carbon and Tritium Isotopes 33
    Noble Gases 37
Groundwater in the Yakima River Basin 38
    Occurrence 38
    Conditions of Occurrence 43
    Flow System 53
Groundwater Use 59
Water-Level Trends 64
Water Budget for the Yakima River Basin 79
Summary and Conclusions 80
Acknowledgments 81
Selected References 82

Contents—Continued

Appendix A. Methods and Assumptions Used to Estimate Lateral Hydraulic Conductivity from Specific-Capacity Data 91

Appendix B. Hydrochemical Data from Groundwater Samples, Yakima River Basin Aquifer System, Washington 92

Appendix C. Methods Used to Estimate Groundwater Recharge for the Area Outside the Yakima River Basin in the Extended Study Area 102

Appendix D. Estimation of Mean Annual Values and Development of Time-Series of Monthly Values for Surface-Water Diversions, Yakima River Basin, Washington 105

Plates

[in pocket]

Plate 1. Maps showing location of six structural basins and extents of basin-fill hydrogeologic units, Yakima River basin aquifer system, Washington.

Plate 2. Maps showing surficial extent of bedrock hydrogeologic units, and extent and altitude of the top of the Columbia River Basalt Group hydrogeologic units, Yakima River basin aquifer system, Washington.

Plate 3. Map showing generalized altitude of the water-table in six structural basins, spring 2001, Yakima River basin aquifer system, Washington.

Plate 4. Maps showing generalized altitude of groundwater levels for the Saddle Mountains, Wanapum, and Grande Ronde units, approximating spring 2001 conditions, Yakima River basin aquifer system, Washington.

Figures

Figure 1. Map showing Yakima River basin, Washington 3

Figure 2. Map showing land use and land cover, Yakima River basin, Washington, 1999 ... 5

Figure 3. Selected tributaries, diversion canals, return flows, and streamflow-gaging stations, Yakima River basin, Washington 6

Figure 4. Map showing distribution of depths of water wells, Yakima River basin, Washington 8

Figure 5. Map showing simplified surficial geology, Yakima River basin, Washington.... 9

Figure 6. Map showing location of wells with water-level measurements, June 2000 through June 2002, Yakima River basin, Washington 11

Figure 7. Map showing location of six structural basins and geologic structure, Yakima River basin aquifer system, Washington 16

Figure 8. Map showing extent of surficial hydrogeologic units, Yakima River basin aquifer system, Washington 19

Figure 9. Map showing reported yields of wells, Yakima River basin aquifer system, Washington 26

Figures—Continued

Figure 10. Graph showing frequency distribution of lateral hydraulic conductivity values estimated from specific-capacity data, Yakima River basin aquifer system, Washington 27
Figure 11. Box plots showing distribution of lateral hydraulic conductivity estimated from aquifer test data, Yakima River basin aquifer system, Washington 28
Figure 12. Graph showing relation between stable isotopic composition (deuterium and oxygen-18) samples in all water samples, precipitation in the Toppenish basin, surface water in the Yakima River basin, and groundwater in the Toppenish basin, Washington 32
Figure 13. Map showing the spatial distribution of dissolved methane concentrations and carbon-13 ratios of dissolved methane in groundwater, Yakima River basin aquifer system, Washington 34
Figure 14. Graphs showing relation between tritium and δ13C and carbon-14 (14C) concentrations in dissolved inorganic carbon of groundwater samples from the Toppenish and Yakima basins, Washington and δ13C and fluoride and chloride concentrations 35
Figure 15. Graph showing concentrations of noble gases in groundwater samples from Yakima River basin aquifer system, Washington, and calculated concentrations of noble gases in air-saturated water 37
Figure 16. Map showing spatial distribution of mean annual recharge for current conditions, 1960–2001, Yakima River basin aquifer system, Washington 40
Figure 17. Map showing surface-water irrigation districts, Yakima River basin, Washington 41
Figure 18. Map showing generalized depth to water table for parts of the Yakima River basin aquifer system, Washington 44
Figure 19. Hydrographs showing depth-to-water in four wells, Yakima River basin, Washington 46
Figure 20. Maps showing location and depth of historical and present-day flowing wells, and location of historical and present-day springs, Yakima River basin aquifer system, Washington 49
Figure 21. Hydrographs showing depth-to-water in wells completed in the Wanapum unit, and wells completed in the Wanapum 12N/21E-16L01) and Saddle Mountains (12N21E-16N02) units, Yakima River basin aquifer system, Washington 51
Figure 22. Map showing spatial distribution of pumpage from the Yakima River basin aquifer system, Washington, 2001 62
Figure 23. Map showing spatial distribution of cumulative pumpage from the Yakima River basin aquifer system, Washington, 1960–2001 63
Figure 24. Map showing location and estimated mean annual quantity of surface-water diversions in the Yakima River basin, Washington 64
Figure 25. Graphs showing examples of groundwater level hydrographs with stable or small declines, Yakima River basin, Washington 66
Figure 26. Graphs showing examples of groundwater level hydrographs with moderate declines, Yakima River basin, Washington 67
Figure 27. Graphs showing examples of groundwater level hydrographs with large declines, Yakima River basin, Washington 68

Figures—Continued

Figure 28. Graphs showing examples of groundwater level hydrographs with very large declines, Yakima River basin, Washington ... 70
Figure 29. Map showing location of irrigation wells completed in the basin-fill deposits, Yakima River basin aquifer system, Washington ... 72
Figure 30. Map showing generalized trends in groundwater levels, Saddle Mountains unit, Yakima River basin aquifer system, Washington ... 74
Figure 31. Map showing generalized trends in groundwater levels, Wanapum unit, Yakima River basin aquifer system, Washington ... 75
Figure 32. Map showing generalized trends in groundwater levels, Grande Ronde unit, Yakima River basin aquifer system, Washington ... 76
Figure 33. Map showing spatial distribution of cumulative groundwater recharge for 1960–2001, Yakima River basin aquifer system, Washington ... 77
Figure 34. Map showing spatial distribution of the difference between groundwater recharge and pumpage acummulated for 1960–2001 for areas with pumpage, Yakima River basin aquifer system, Washington ... 78

Tables

Table 1. Information for the hydrogeologic units of the Yakima River basin aquifer system, Washington ... 12
Table 2. Correlation chart showing the regional relation between generalized geologic and hydrogeologic units in the basin-fill and Columbia River Basalt Group units, and bedrock units for the Yakima River basin aquifer system, Washington ... 14
Table 3. Summary of previous estimates of lateral hydraulic conductivity, Yakima River basin aquifer system, Washington ... 21
Table 4. Summary of lateral hydraulic conductivity values, estimated from specific-capacity data, Yakima River basin aquifer system, Washington ... 27
Table 5. Summary of estimated annual pumpage, by pumpage category in 5-year increments, Yakima River basin, Washington, 1960–2000 ... 60
Table 6. Generalized mean annual water budget for the Yakima River basin, Washington ... 79

Conversion Factors, Datums, and Abbreviations and Acronyms

Conversion Factors

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
acre	4,047	square meter (m^2)
acre	0.004047	square kilometer (km^2)
section (640 acres or 1 square mile)	259.0	square hectometer (hm^2)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
cubic inch (in^3)	16.39	cubic centimeter (cm^3)
acre-foot (acre-ft)	1,233	cubic meter (m^3)
	Flow rate	
foot per day (ft/d)	0.3048	meter per day (m/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per square mile [(ft^3/s)/mi^2]	0.01093	cubic meter per second per square kilometer [(m^3/s)/km^2]
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Pressure	
pound per square foot (lb/ft^2)	0.04788	kilopascal (kPa)
	Specific capacity	
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter [(L/s)/m]
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Hydraulic gradient	
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

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

$$°F=(1.8\times°C)+32.$$

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft^3/d)/ft^2]ft. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Specific conductance is given in microsiemens per centimeter at 25 degrees Celsius (µS/cm at 25°C).

Concentrations of chemical constituents in water are given either in milligrams per liter (mg/L) or micrograms per liter (µg/L).

Conversion Factors, Datums, and Abbreviations and Acronyms—Continued

Datums

Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88).

Horizontal coordinate information is referenced, unless otherwise noted, to the North American Datum of 1983 (NAD 83).

Altitude, as used in this report, refers to distance above the vertical datum.

Abbreviations and Acronyms

CID	Cascade Irrigation District
CO_2	carbon dioxide
CRBG	Columbia River Basalt Group
DIC	dissolved inorganic carbon
DPM	Deep Percolation Model
DTW	depth to water table
GMWL	Global Meteoric Water Line
He	helium
K_h	lateral hydraulic conductivity
K_v	vertical hydraulic conductivity
MAP	mean annual precipitation
NWIS	National Water Inventory System
PMC	percent modern carbon
Reclamation	Bureau of Reclamation
STP	standard temperature and pressure
TWSA	total water supply available
USGS	U.S. Geological Survey
WaDOE	Washington State Department of Ecology
WIP	Wapato Irrigation Project
YN	Yakama Nation

Well Numbering System

The USGS assigns numbers to wells and springs in Washington that identify their locations in a township, range, and section. Well number 20N/15E-26N01 indicates, successively, the township (T.20 N.) and the range (R.15 E.) north and east of the Willamette baseline and meridian. The first number following the hyphen indicates the section (26) within the township and the letter following the section number (N) gives the 40-acre subdivision within the section. The number (01) following the letter is the sequence number of the well within the 40-acre subdivision. An 'S' following the sequence number indicates that the site is a spring, a 'D1' after the sequence number indicates that the original reported depth of the well has been changed once and successive numbers indicate the number of changes in the well depth. An 'R' following the sequence number indicates the well has been reconditioned. A 'P1' or an 'A' after the sequence number indicates a group of nested piezometers, with successive numbers or letters assigned to each piezometer in the group.

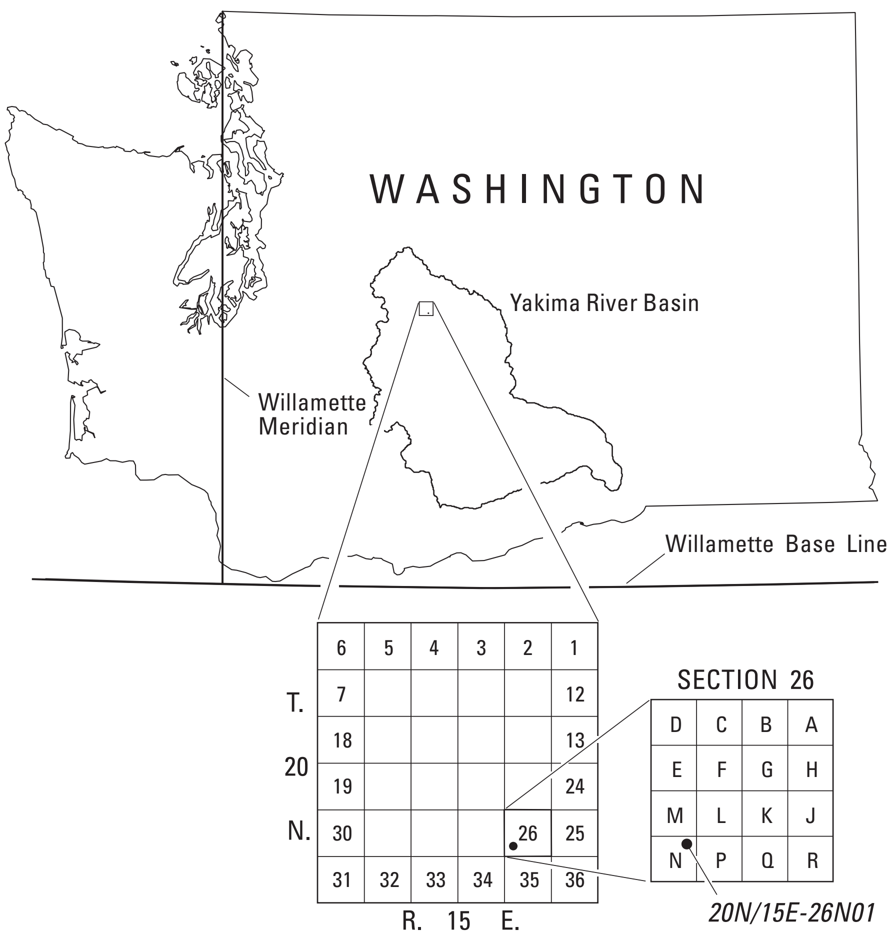

Diagram of well numbering system in the State of Washington.

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By J.J. Vaccaro, M.A. Jones, D.M. Ely, M.E. Keys, T.D. Olsen, W.B. Welch, and S.E. Cox

Abstract

The Yakima River basin aquifer system underlies about 6,200 square miles in south-central Washington. The aquifer system consists of basin-fill deposits occurring in six structural-sedimentary basins, the Columbia River Basalt Group (CRBG), and generally older bedrock. The basin-fill deposits were divided into 19 hydrogeologic units, the CRBG was divided into three units separated by two interbed units, and the bedrock was divided into four units (the Paleozoic, the Mesozoic, the Tertiary, and the Quaternary bedrock units). The thickness of the basin-fill units and the depth to the top of each unit and interbed of the CRBG were mapped. Only the surficial extent of the bedrock units was mapped due to insufficient data. Average mapped thickness of the different units ranged from 10 to 600 feet.

Lateral hydraulic conductivity (K_h) of the units varies widely indicating the heterogeneity of the aquifer system. Average or effective K_h values of the water-producing zones of the basin-fill units are on the order of 1 to 800 ft/d and are about 1 to 10 ft/d for the CRBG units as a whole. Effective or average K_h values for the different rock types of the Paleozoic, Mesozoic, and Tertiary units appear to be about 0.0001 to 3 ft/d. The more permeable Quaternary bedrock unit may have K_h values that range from 1 to 7,000 ft/d. Vertical hydraulic conductivity (K_v) of the units is largely unknown. K_v values have been estimated to range from about 0.009 to 2 ft/d for the basin-fill units and K_v values for the clay-to-shale parts of the units may be as small as 10^{-10} to 10^{-7} ft/d. Reported K_v values for the CRBG units ranged from 4×10^{-7} to 4 ft/d.

Variations in the concentrations of geochemical solutes and the concentrations and ratios of the isotopes of hydrogen, oxygen, and carbon in groundwater provided information on the hydrogeologic framework and groundwater movement. Stable isotope ratios of water (deuterium and oxygen-18) indicated dispersed sources of groundwater recharge to the CRBG and basin-fill units and that the source of surface and groundwater is derived from atmospheric precipitation. The concentrations of dissolved methane were larger than could be attributable to atmospheric sources in more than 80 percent of wells with measured methane concentrations. The concentrations of the stable isotope of carbon-13 of methane were indicative of a thermogenic source of methane. Most of the occurrences of methane were at locations several miles distant from mapped structural fault features, suggesting the upward vertical movement of thermogenic methane from the underlying bedrock may be more widespread than previously assumed or there may be a more general occurrence of unmapped (buried) fault structures. Carbon and tritium isotope data and the concentrations of dissolved constituents indicate a complex groundwater flow system with multiple contributing zones to groundwater wells and relative groundwater residence time on the order of a few tens to many thousands of years.

Potential mean annual recharge for water years 1950–2003 was estimated to be about 15.6 in. or 7,149 ft^3/s (5.2 million acre-ft) and includes affects of human activities such as irrigation of croplands. If there had been no human activities (predevelopment conditions) during that time period, estimated recharge would have been about 11.9 in. or 5,450 ft^3/s (3.9 million acre-ft). Estimated mean annual recharge ranges from virtually zero in the dry parts of the lower basin to more than 100 in. in the humid uplands, where annual precipitation is more than 120 in.

Groundwater in the different hydrogeologic units occurs under perched, unconfined, semiconfined, and confined conditions. Groundwater moves from topographic highs in the uplands to topographic low areas along the streams. The flow system in the basin-fill units is compartmentalized due to topography and geologic structure. The flow system also is compartmentalized for the CRBG units but not to as large an extent as in the basin-fill units. Regional groundwater flow discharges to surface-water drainage features in the lowlands in the structural basins. The gradient of the water table ranges from about 7 ft/mi to more than 400 ft/mi, with the smaller gradients in the topographically smooth parts of the structural basins. Typically, hydraulic gradients are similar to the topographic gradients. The lateral hydraulic gradient in the CRBG units also is highly variable and ranges from about 3 to 14 ft/mi in the flatter parts of the structural basins to as much as 800 ft/mi in areas of steep terrain. The hydraulic gradient of the CRBG units generally is within 5 degrees of the topographic gradient.

About 312,000 acre-ft (about 430 ft^3/s) of groundwater was pumped in 2000 for multiple uses, about 60 percent of which was for irrigation. Allowable acreage of groundwater irrigation rights is about 130,000 acres. Mean annual surface-water diversions are about 5,800 ft^3/s, of which about 66 percent was delivered for irrigation and 25 percent was for power production.

Excluding the initial water-level rises due to surface-water irrigation that started in the late 1800s, long-term measurements of the levels indicate they have been stable or have declined less than 20 ft in most of the aquifer system. Declines from 21 to 300 ft have occurred in some areas due to pumpage. The largest declines have been in the CRBG units. The pumpage from the basin-fill units, from which about 50 percent of the pumpage occurs, appears to have caused declines in only two small areas, but the withdrawals may be reducing groundwater discharge.

Introduction

Surface water in the Yakima River basin, in south-central Washington (fig. 1) is under adjudication. The amount of surface water available for appropriation is unknown, but there are increasing demands for water for municipal, fisheries, agricultural, industrial, and recreational uses. These demands must be met by groundwater withdrawals and (or) by changes in the way water resources are allocated and used. On-going activities in the basin for enhancement of fisheries and obtaining additional water for agriculture may be affected by groundwater withdrawals and by rules implemented under the Endangered Species Act for salmonids that have been either listed or were proposed for listing in the late 1990s. An integrated understanding of the groundwater flow system and its relation to the surface-water resources is needed in order to implement most water-resources management strategies in the basin. In order to gain this understanding, a study of the Yakima River basin aquifer system began in June 2000. The study is a cooperative effort of the U.S. Geological Survey (USGS), Bureau of Reclamation (Reclamation), the Washington State Department of Ecology (WaDOE), and the Yakama Nation (YN).

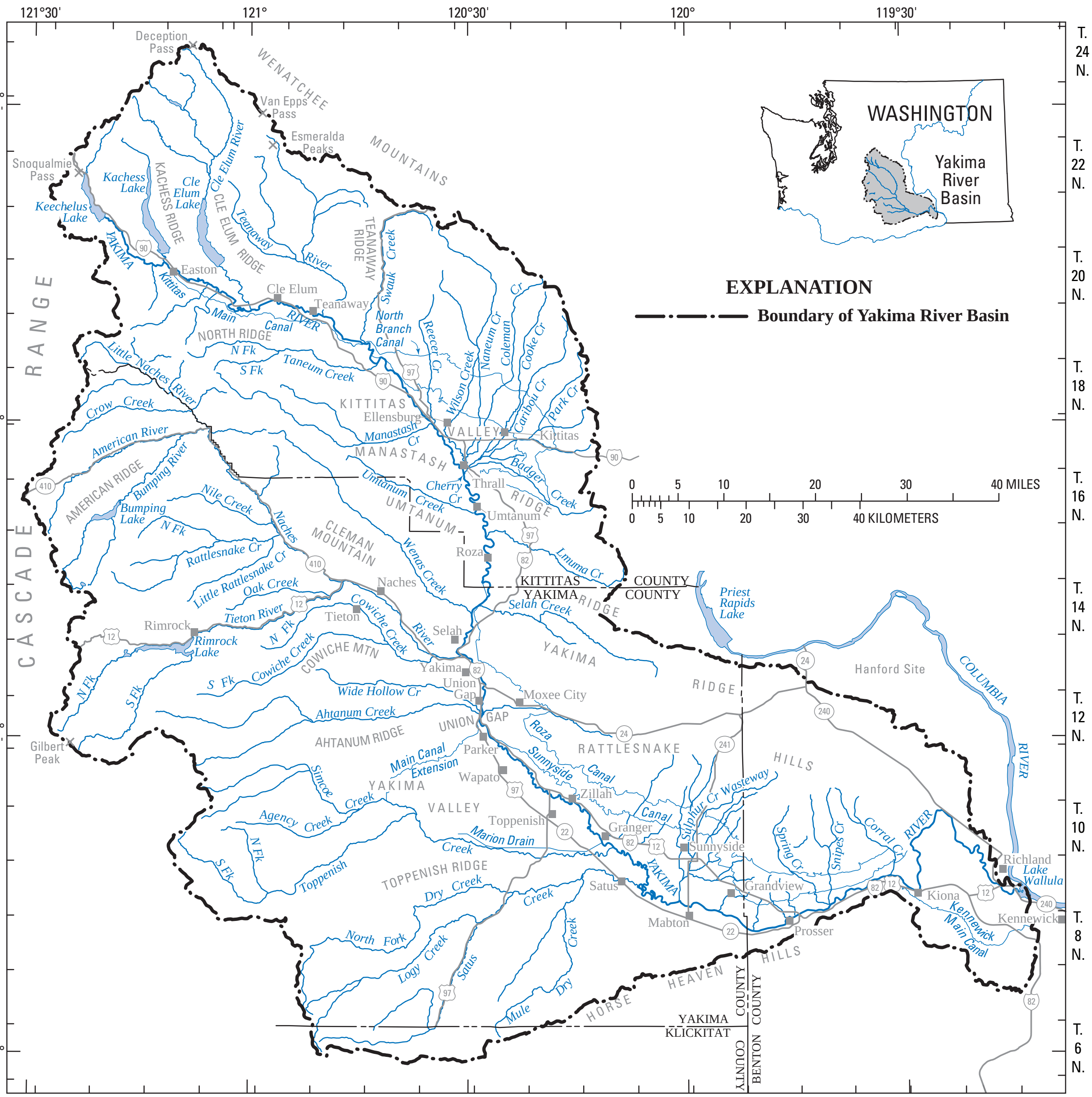
The overall objectives of the study are to fully describe the groundwater flow system and its interaction with and relation to surface water, and to provide baseline information for a management tool—a numerical model of the system. The conceptual model of the flow system and the results of the study can be used to guide and support actions taken by management agencies with respect to groundwater availability and to provide information to other stakeholders and interested parties. The numerical model will be developed as an integrated tool to assess short-term to long-term management activities, including the testing of the potential effects of alternative management strategies for water development and use.

The study includes three phases. The first phase includes (1) project planning and coordination, (2) compiling, documenting, and assessing available data, and (3) initial data collection. The second phase consists of data collection to support the following Phase 2 work elements: (1) mapping of hydrogeologic units, (2) estimating groundwater pumpage, (3) developing estimates of groundwater recharge, (4) assessing groundwater-surface water interchanges, and (5) constructing maps of groundwater levels. Together, these five elements provide the information needed to describe the groundwater flow system, develop the conceptual model, and provide the building blocks for the hydrogeologic framework. In the third phase, a regional-scale numerical model of the groundwater flow system will be constructed in order to integrate the available information. The numerical model will be used to enhance understanding of the flow system (including a water budget for the aquifer system) and its relation to surface water, and to test alternative management strategies.

The results of selected work elements of this study have been described in a series of reports (Jones and others, 2006; Vaccaro and Maloy, 2006; Vaccaro and Sumioka, 2006; Vaccaro, 2007; Vaccaro and Olsen, 2007a, b; Jones and Vaccaro, 2008; Keys and others, 2008; Vaccaro and others, 2008).

Purpose and Scope

This report presents the hydrogeologic framework of the Yakima River basin aquifer system, including a brief description of the hydrogeologic units composing the aquifer system and their characteristics, hydrochemistry, groundwater occurrence and conditions, movement of groundwater in the flow system, groundwater use, and the status and trends of groundwater levels. The information is described in terms of the characteristics of the hydrogeologic units that compose the aquifer system. This information was used to develop a generalized water budget for the Yakima River basin.



Base modified from U.S. Geological Survey data and other digital sources
Universal Transverse Mercator projection, Zone 10
Horizontal Datum: North American Datum of 1927 (NAD 27)

Figure 1. Yakima River basin, Washington.

Description of Study Area

The location and setting of the Yakima River basin, a summary of the development of water resources in the basin, and an overview of the geology are presented to provide a general background for understanding the study area.

Location and Setting

The Yakima River basin encompasses about 6,200 mi^2 in south-central Washington (fig. 1). The Yakima River basin produces a mean annual unregulated streamflow (adjusted for regulation and without diversions or returns) of about 5,600 ft^3/s (4.1 million acre-ft or about 0.9 $(ft^3/s)/mi^2$) and a regulated streamflow of about 3,600 ft^3/s (2.6 million acre-ft or about 0.6 $(ft^3/s)/mi^2$). The basin includes three Washington State Water Resource Inventory Areas (WRIA—numbers 37, 38, and 39), part of the Yakama Nation lands, and three ecoregions (Cascades, Eastern Cascades, and Columbia Basin—Omernik, 1987; Cuffney and others, 1997). Almost all of Yakima County, more than 80 percent of Kittitas County, and about 50 percent of Benton County are in the basin. Less than 1 percent of the basin, principally in an unpopulated upland area, lies in Klickitat County.

The headwaters of the basin are on the upper, humid eastern slope of the Cascade Range, where mean annual precipitation is more than 120 in. The basin terminates at the confluence of the Yakima and Columbia Rivers in the low-lying, arid part of the basin that receives about 6 in. of precipitation per year. Altitudes in the basin range from 400 ft to nearly 8,000 ft. Eight major rivers and numerous smaller streams are tributary to the Yakima River (fig. 1), the largest of which is the Naches River. Most of the precipitation in the basin falls during the winter months as snow in the mountains. The mean annual precipitation over the entire basin is about 27 in. (about 12,300 ft^3/s or 8.9 million acre-ft). The spatial pattern of mean annual precipitation resembles the pattern of the basin's highly variable topography. The difference between the mean annual precipitation and mean annual unregulated streamflow is 6,700 ft^3/s (about 4.8 million acre-ft). On the basis of this difference and the assumption of only small groundwater inflow to or outflow from the aquifer system, about 55 percent of the precipitation was lost to evapotranspiration under natural conditions.

The basin is separated into several broad valleys by east-west trending anticlinal ridges. The valley floors slope gently toward the Yakima River. Few perennial tributary streams traverse these valleys. Most of the population and economic activity occurs in these valleys.

Irrigated agriculture is the principal economic activity in the basin. The average annual surface-water demand met by Reclamation's Yakima Project is about 2.5 million acre-ft; an additional 336,000 acre-ft of demand in the lower part of the basin is separate from the demand met by the Project. Additional surface-water demand that is not met by Reclamation occurs in smaller tributaries and on the large rivers; this demand is based on State appropriated water. More than 95 percent of the surface-water demand is for irrigation of about 500,000 acres in the low-lying semiarid to arid parts of the basin (fig. 2). The demand is partly met by storage of nearly 1.1 million acre-ft of water in five Reclamation reservoirs. The major management point for Reclamation is the streamflow-gaging station at the Yakima River near Parker at river mile 103.7 (USGS station number 12505000, fig. 3); this site is just downstream of the Sunnyside and Wapato (main) canal diversions. Just upstream of this site at about river mile 106.9 is the location that is considered the dividing line between the upper (mean annual precipitation of 7–145 in.) and lower (mean annual precipitation of 6–45 in.) parts of the Yakima River basin. About 45 percent of the water diverted for irrigation is eventually returned to the river system as surface-water inflows and groundwater discharge, but at varying time-lags (Bureau of Reclamation, 1999). During the low-flow period, these return flows, on average, account for about 75 percent of the streamflow downstream of the streamflow-gaging station near Parker. Most of the surface-water demand in the basin downstream of Parker is met by these return flows and not by the release of water from the reservoirs. As a result of water use in the basin, the difference between mean annual unregulated (5,600 ft^3/s) and regulated (3,600 ft^3/s) streamflow in the basin is about 2,000 ft^3/s, suggesting that some 1.4 million acre-ft of water, or about 17 percent of the precipitation in the basin, is consumptively used—principally by irrigated crops through evapotranspiration.

Development of Water Resources

Missionaries arrived in the basin in 1848 and established a mission in 1852 on Atanum (now Ahtanum) Creek. They were some of the first non-Indian settlers to use irrigation on a small scale. Miners and cattlemen immigrated to the basin in the 1850s and 1860s, which resulted in a new demand for water. With increased settlement in the mid-1860s, irrigation of the fertile valley bottoms began and the outlying areas were extensively used for raising stock. One of the first known non-Indian irrigation ditches was constructed in 1867 and diverted water from the Naches River (Parker and Storey, 1916; Flaherty, 1975). Private companies later delivered water through canal systems built between 1880 and 1904 for the irrigation of large areas. The development of irrigated agriculture was made more attractive by the construction of the Northern Pacific Railway that reached Yakima in December 1884 and provided a means to transport agricultural goods to markets; two years later, the completion of the railway to the Pacific coast provided new and easily accessible

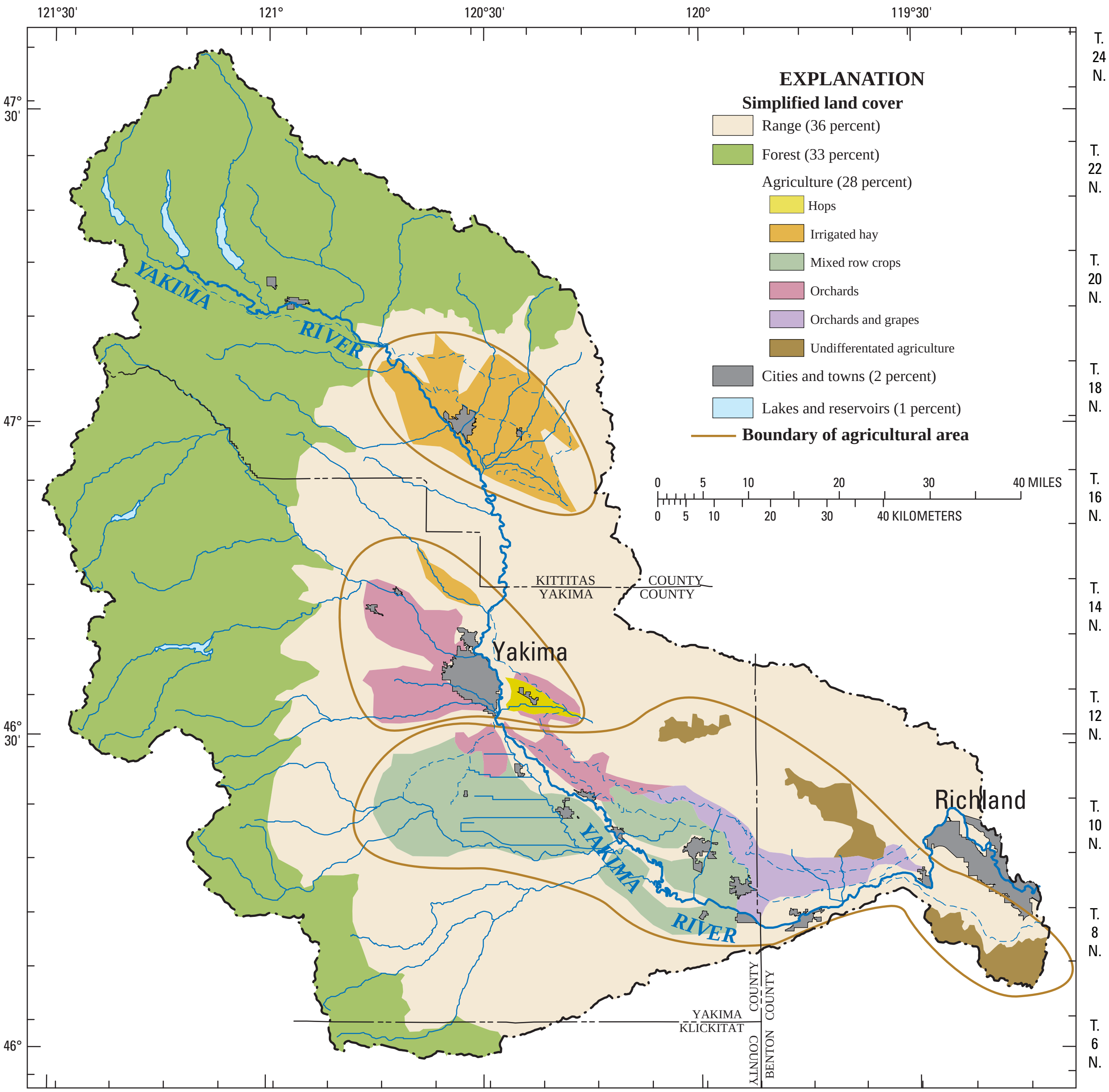


Figure 2. Land use and land cover, Yakima River basin, Washington, 1999.

markets for agricultural products. The State of Washington was created in 1889, spurring further growth in the basin, especially because the cities of Ellensburg and Yakima were in contention for being the state capital. By 1902, about 120,000 acres were under irrigation, mostly by surface water (Parker and Storey, 1916; Bureau of Reclamation, 1999).

The Federal Reclamation Act of 1902 enabled the construction of Federal water projects in the western United States in order to expand the development of the West. In 1905, the Washington State Legislature passed the Reclamation Enabling Act, and the Yakima Federal Reclamation Project was authorized to construct facilities to

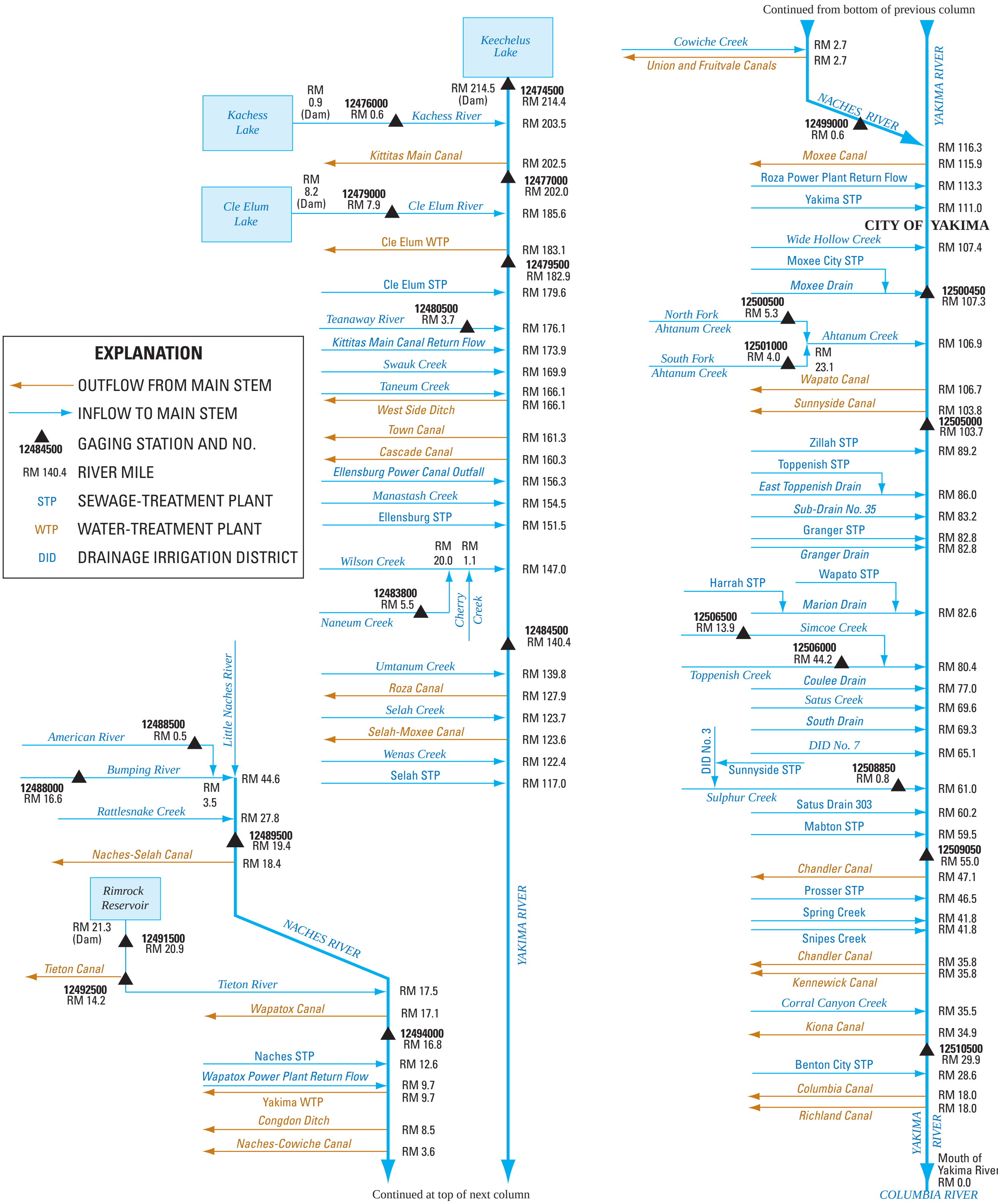


Figure 3. Selected tributaries, diversion canals, return flows, and streamflow-gaging stations, Yakima River basin, Washington.

irrigate about 500,000 acres. As part of the 1905 authorization and extensions, all forms of further appropriation of unappropriated water in the basin were withdrawn (Parker and Storey, 1916). Six dams were constructed as part of the Yakima Project: Bumping Dam in 1910, Kachess Dam in 1912, Clear Creek Dam in 1914, Keechelus Dam in 1917, Tieton Dam (Rimrock Lake) in 1925, and Cle Elum Dam in 1933. The six reservoirs have a total capacity of about 1.07 million acre-ft, of which about 78 percent is stored in the upper arm of the Yakima River and 22 percent is stored in the Naches River arm. The construction of the dams and other irrigation facilities resulted in an extremely complicated surface-water system (fig. 3). These Federal reservoirs provide water storage to meet irrigation requirements of the major irrigation districts at the time of year when the natural streamflow from unregulated streams can no longer meet demands; this time is referred to as the 'storage control' date. Several of the reservoirs also provide instream flows during the winter for the incubation of salmon eggs in redds (gravel spawning nests).

Legal challenges to water rights resulted in the 1945 Consent Decree (U.S. District Court, 1945) that established the framework of how Reclamation operates the Yakima Project to meet water demands. The Decree determined two classes of rights—nonproratable and proratable. When the total water supply available (TWSA, defined as current available storage in the reservoirs, forecasted estimates of unregulated flow, and other sources that are principally return flows) is not sufficient to meet both classes of rights, the proratable (junior) rights are decreased according to the quantity of water available defined by the TWSA. As of 2008, the years when proration levels were defined were 1973, 1977, 1979, 1987–88, 1992–94, 2001, and 2005. This legally mandated method, which was upheld in a 1990 court ruling, generally performs well in most years, but is dependent on the accuracy of the TWSA estimate. In some years, for example 1977, problems have arisen because of errors in the TWSA estimate (Kratz, 1978; Glantz, 1982). In addition, numerous proratable users have obtained groundwater-water rights to pump supplemental water in the years that they receive prorated quantities of surface water. System management also accounts for defined instream flows at selected target points on the river, and for suggested changes in storage releases recommended by the Systems Operations Advisory Committee (SOAC)—the advisory board of fishery biologists representing the different stakeholders (Systems Operations Advisory Committee, 1999). The operations for meeting instream flows are most affected by a 1980 Federal district court decision and by Title XII of a Public Law that instituted (beginning in 1995) new instream flows at two diversions dams (Sunnyside and Prosser).

The drilling of numerous wells for irrigation was spurred by new (post 1945) well-drilling technologies, legal rulings, and the onset of a multi-year dry period in 1977 (Vaccaro, 1995, 2002). Population growth in the basin was, and still is, the driving force behind the increased drilling of shallow domestic wells and deeper public water supply wells. Currently, there are more than 20,000 wells (fig. 4) in the basin. More than 70 percent of these wells are shallow, 10–250 ft deep, domestic wells. On the basis of the digital water-rights database provided by WaDOE (R. Dixon, Washington State Department of Ecology, written commun., 2001) and other information, at least 2,874 active groundwater rights are associated with the wells in the basin that can collectively withdraw an annual quantity of about 529,231 acre-ft during dry years. The irrigation rights are for the irrigation of about 129,570 acres. There are about 16,600 groundwater claims in the basin; these claims are for some 270,000 acre-ft of groundwater (J. Kirk, Washington State Department of Ecology, written commun., 1998). 'A water right claim is a statement of claim to water use that began before the state Water Codes were adopted, and is not covered by a water right permit or certificate. A water right claim does not establish a water right, but only provides documentation of one if it legally exists. Ultimately, the validity of claimed water rights would be determined through general water right adjudications' (Washington State Department of Ecology, 1998). A groundwater claim means a user claims that they were using groundwater continuously for a particular use, prior to 1945, when the State legislature enacted the Ground Water Code.

Overview of the Geology

The Columbia Plateau has been informally divided into three physiographic subprovinces (Meyers and Price, 1979). The western margin of the Columbia Plateau contains the Yakima Fold Belt subprovince and includes the Yakima River basin. The Yakima Fold Belt is a highly folded and faulted region, and within the study area it is underlain by various consolidated rocks ranging in age from Precambrian to Tertiary, and unconsolidated materials and volcanic rocks of Quaternary age. The simplified surficial geology (Fuhrer and others, 1994; more detail shown later) clearly shows the wide variety of rock materials present in the basin (fig. 5). In the Yakima River basin, the headwater areas in the Cascade Range include metamorphic, sedimentary, and intrusive and extrusive igneous rocks. The central, eastern, and southwestern parts of the basin are composed of basalt lava flows of the Columbia River Basalt Group (CRBG) with some intercalated sediments that are discontinuous and weakly consolidated. The lowlands are underlain by unconsolidated and weakly consolidated valley-fill comprising glacial, glacio-fluvial, lacustrine, and alluvium deposits that in places exceed 1,000 ft in thickness (Drost and others, 1990). Wind-blown deposits, called loess, are present locally along the lower valley.

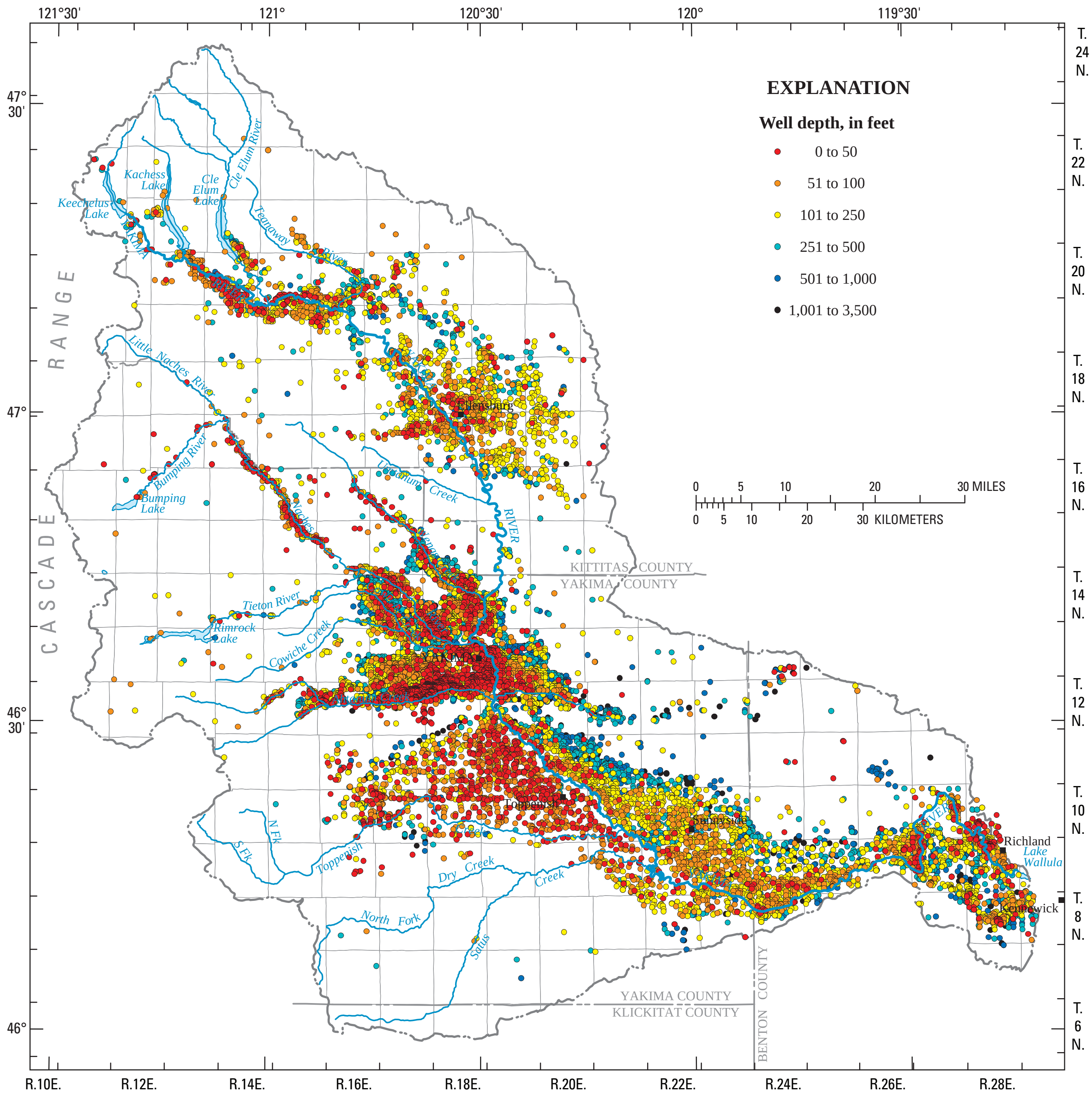


Figure 4. Distribution of depths of water wells, Yakima River basin, Washington.

Valley-fill deposits and basalt lava flows are important for groundwater occurrence in the study area. The basalt consists of a series of flows erupted during various stages of the Miocene Age, from 17 to 6 million years ago. Basalt erupted from fissures located in the eastern part of the Columbia Plateau and individual flows range in thickness from a few feet to more than 100 ft. The total thickness in the central part of the plateau is estimated to be greater than 10,000 ft (Drost and others, 1990) and the maximum thickness in the study area is more than 8,000 ft. Unlike most of the Columbia Plateau, the CRBG in the Yakima Fold Belt is underlain by sedimentary rocks. The valley-fill deposits were eroded from the Cascade Range and from the east-west-trending anticlinal ridges that were formed by the buckling of the basalt sequence during mid- to late-Miocene time. Most of these deposits are part of the Ellensburg Formation. This formation underlies,

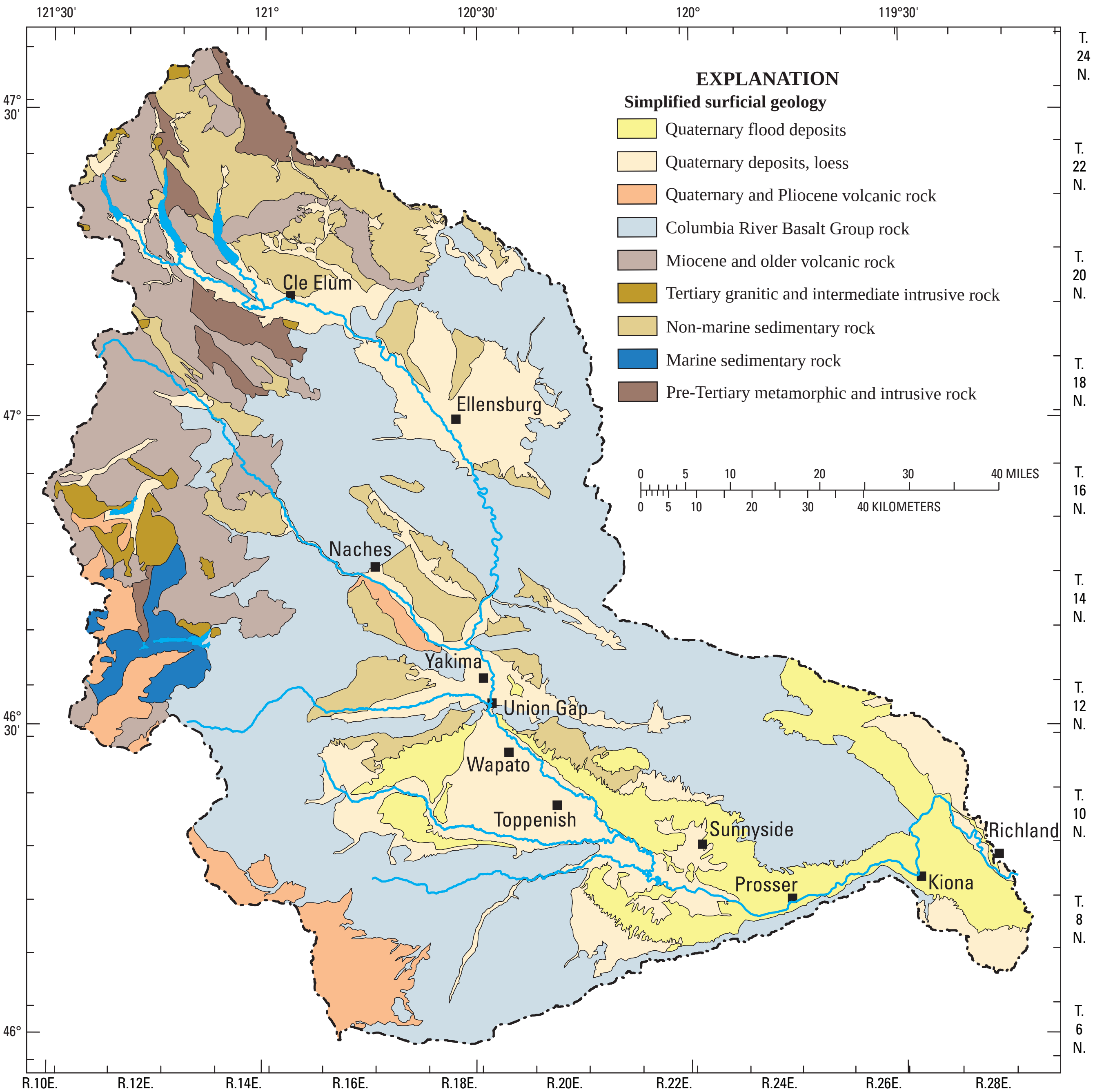


Figure 5. Simplified surficial geology, Yakima River basin, Washington.

intercalates, and overlies the basalts along the western edge, and constitutes most of the thickness of the unconsolidated deposits (informally called the overburden; Drost and others, 1990) in the basins. The basins are narrow to large open synclinal valleys between the numerous anticlinal ridges.

The deposition of a thick, upper sequence of sand, gravel, and some fine-grained material is the result of erosion by glacial ice and transport by meltwater streams. Damming of large lakes by glacial ice during the Pleistocene epoch resulted in the deposition of silt and clay beds in parts of the uplands. When the lakes drained, the fine sediments were exposed, subsequently eroded by wind, and deposited over the lower, eastern parts of the study area. Thus, the unconsolidated materials in the basins that abut and are interbedded with the basalts range in age from Miocene to Holocene.

Well Information

References to Wells

Wells discussed in this report are identified by USGS well number, as described in section "Well-Numbering System." A well number without sequence number indicates that the well is not in the USGS National Water Information System (NWIS) and that the well information is based on a log obtained from WaDOE. Except when discussing hydrochemistry in section "Methane," well numbers are not shown in maps. Instead, townships and ranges of the Public Land Survey System are identified in most maps, which allows the general determination of well locations based on USGS well numbers. This approach provides the option of generally locating a specific well in a variety of maps, such as those showing groundwater recharge or pumpage.

Well Data and Their Use

Four sets of water-level measurements in wells throughout the study area were conducted during the study. The measurements were made in autumn 2000, spring 2001, autumn 2001, and spring 2002. In addition, YN staff made monthly measurements at numerous wells, as well as weekly measurements at nine wells during the period of interest (June 2000–June 2002). A total of about 8,300 water levels were measured during this latter period. The measured wells were selected to obtain a good spatial distribution, both laterally and vertically; the location of wells with water-level measurements during the period June 2000 through June 2002 is shown in figure 6. In addition, historical water-level measurements made by WaDOE and YN that had not been entered into NWIS were compiled. Information and water levels for all these wells were put into NWIS.

Additional historical water levels were available from more than 8,000 wells. A subset of these water levels, which is in NWIS, was measured by the USGS. A second subset of these water levels also is in NWIS, and represents water levels reported on drillers' well logs. A third subset of water levels consists of water levels reported on 'uncoded' well logs. Uncoded well logs were any well logs in USGS paper files that were not in NWIS. Selected information from more than 9,000 of these uncoded wells was put into digital form in order to provide additional information for: (1) mapping the spatial distribution of wells in the basin (fig. 4), (2) evaluating the spatial distribution of well depths, (3) analyzing specific-capacity tests, (4) identifying flowing wells, and (5) mapping water levels. Information on specific-capacity tests and water levels was not available for all of the uncoded wells.

During this study, well information from the described data sets was used for constructing various maps and for analysis of the flow system. For example, lithologic information from more than 6,500 wells was used by Jones and others (2006) and Jones and Vaccaro (2008) to map hydrogeologic units. All of the water-level data also was used in some capacity in the study, for example, examining the relation of water levels to well depth and for data in-filling in areas with sparse water-level measurements in 2000–02. All available information for flowing wells and specific-capacity tests also was utilized.

Hydrogeologic Units

Thirty-four hydrogeologic units were mapped and named in this study by Jones and others (2006) and Jones and Vaccaro (2008). Hydrogeologic unit information for the 28 major units is summarized in table 1 and a correlation chart showing the relation between generalized geologic units and hydrogeologic units is shown in table 2. Information used to help define and map the hydrogeologic units in table 2 is from well logs and published maps of Swanson and Wright (1978), Swanson and others (1979a, b), Tanaka and others (1979), Tabor and others (1982, 1987, 1993), Walsh (1986a, b), Korosec (1987), Phillips and Walsh (1987), Schasse (1987), Gulick and Korosec (1990), Reidel and Fecht (1994a, 1994b), and Schuster (1994a, 1994b, 1994c). Most of the maps in the post-1980 publications were available in digital form from the Washington State Department of Natural Resources (2002). Additional information for the Hanford Site was provided by S.P. Reidel and P.D. Thorne (Battelle Institute, written commun., 2003 and 2005) and for the Toppenish basin by Newell Campbell (unpub. maps, produced for the Yakama Nation, 2001).

Information about the hydrogeologic units defined and mapped in this study is summarized below. A more detailed description of the study methods and hydrogeologic units can be found in Jones and others (2006) and Jones and Vaccaro (2008). The detailed descriptions include such aspects as maps of thickness of selected units, depths to the top of selected units, and hydrogeologic sections.

There are four categories of hydrogeologic units: (1) unconsolidated units composed of Pliocene to Recent sediments that may or may not be subdivided, (2) semi-consolidated to consolidated units (referred to here as consolidated units) composed of Miocene-Pliocene sediments, (3) Miocene CRBG and interbed units, and (4) Paleozoic to Quaternary bedrock units. The first 2 categories include 19 mapped units (pl. 1) and 6 subunits of one mapped

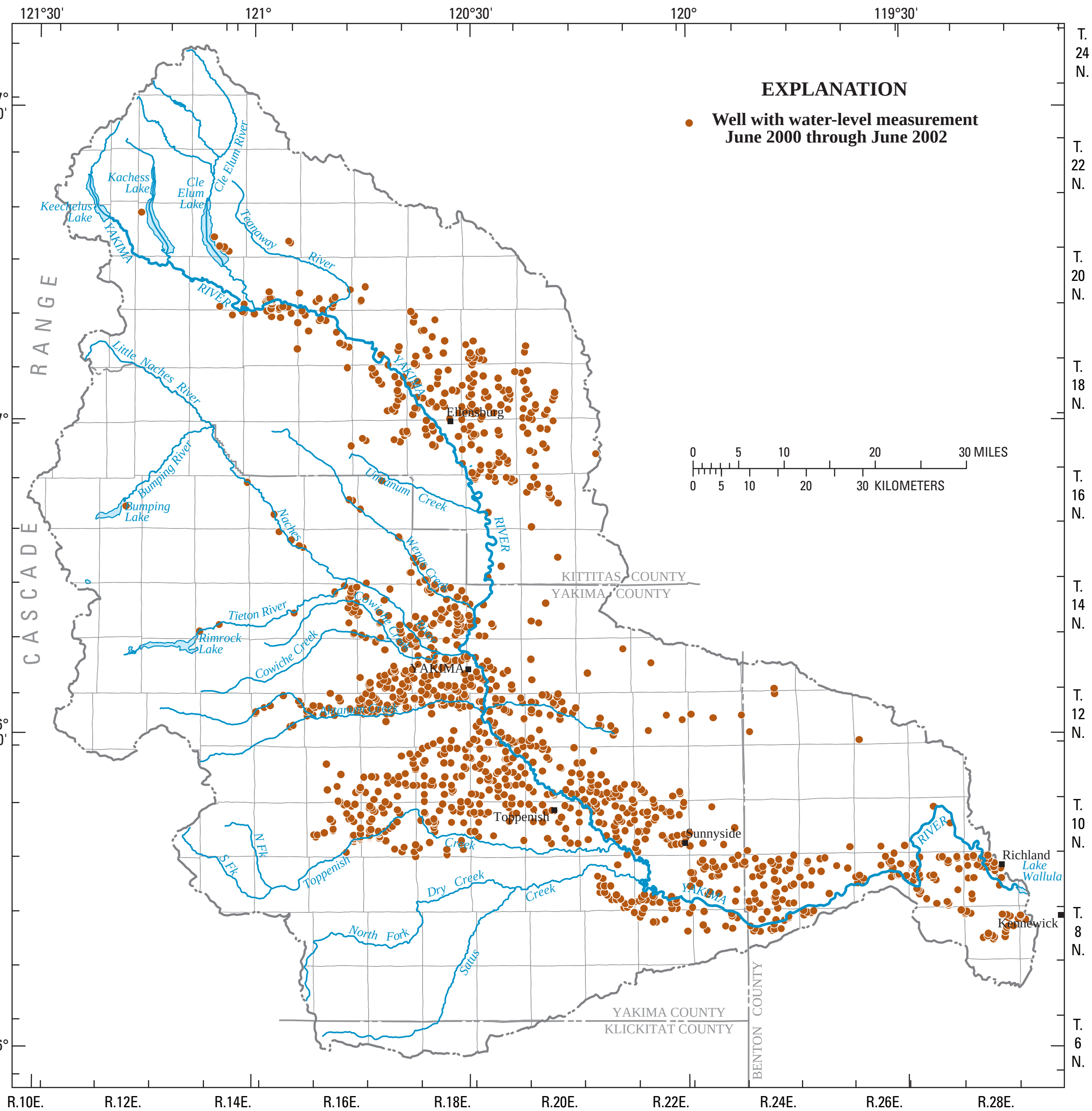


Figure 6. Location of wells with water-level measurements, June 2000 through June 2002, Yakima River basin, Washington.

unconsolidated unit. These two categories consist of basin-fill deposits occurring in six structural-sedimentary basins (fig. 7), herein called structural basins; the geologic structure delineating the basins is clearly defined by the folds and faults shown on figure 7. The structural basins generally are consistent with the groundwater basins of Kinnison and Sceva (1963). Each unit in a structural basin has been named; for example, a unit in the Kittitas basin that is composed of alluvial deposits is named Unit 1 (Jones and others, 2006, table 2). These names, however, may not represent the same type of hydrogeologic unit in different structural basins or represent the same geologic units. For each structural basin, the naming of these units starts at 1 for the uppermost unit and increases with the depth (age) of a unit; thus, if there were three units in a structural basin, they would be numbered 1 through 3.

Table 1. Information for the hydrogeologic units of the Yakima River basin aquifer system, Washington.

[**Abbreviations:** mi^2, square mile; ft, foot; –, not calculated]

Structural basin name	Mapped area (mi^2)	Unit	Lithology	Thickness (ft)		
				Range	Average	Median
Roslyn basin	70	1	Alluvial, lacustrine, and glacial deposits	0 to 360	80	80
		2	Fine-grained lacustrine clay and silt deposits	0 to 530	180	170
		3	Coarse-grained sand and gravel deposits	0 to 240	60	50
		Total basin thickness	All deposits	0 to 700	150	110
Kittitas basin	270	1 (alluvial)	Floodplain alluvial deposits	0 to 100	30	10
		2 (unconsolidated)	Loess, alluvial fan, glacial terrace, and Thorp gravel deposits	0 to 790	180	150
		3 (consolidated)	Ellensburg Formation and undefined continental sedimentary deposits	0 to 2,040	600	350
		Total basin thickness	All deposits	0 to 2,120	500	270
Selah basin	170	1 (alluvial)	Floodplain alluvial deposits	0 to 90	30	30
		2 (unconsolidated)	Loess, alluvial fan, glacial terrace, and Thorp gravel deposits	0 to 290	50	40
		3 (consolidated)	Ellensburg Formation and undefined continental sedimentary deposits	0 to 1,920	320	200
		Total basin thickness	All deposits	0 to 1,920	300	200
Yakima basin	230	1 (alluvial)	Floodplain alluvial deposits	0 to 120	20	20
		2 (unconsolidated)	Loess, alluvial fan, glacial terrace, and Thorp gravel deposits	0 to 350	90	80
		3 (consolidated)	Ellensburg Formation and undefined continental sedimentary deposits	0 to 1,840	510	450
		Total basin thickness	All deposits	0 to 1,840	530	410
Toppenish basin	440	1 (fine-grained unconsolidated)	Touchet Beds, terrace, loess, and some alluvial deposits	0 to 80	10	10
		2 (coarse-grained unconsolidated)	Coarse-grained sand and gravel deposits	0 to 270	90	80
		3 (consolidated)	Consolidated deposits of the upper Ellensburg Formation and undefined continental sedimentary deposits	0 to 970	350	320
		4 (fine-grained deposits)	Top of Rattlesnake Ridge unit of the Ellensburg Formation or ‘Blue Clay unit’	0 to 520	170	140
		5 (coarse-grained deposits)	Base of Rattlesnake Ridge unit of the Ellensburg Formation	0 to 140	20	20
		Total basin thickness	All deposits	0 to 1,210	550	550

Table 1. Information for the hydrogeologic units of the Yakima River basin aquifer system, Washington.—Continued

[**Abbreviations:** mi^2, square mile; ft, foot; –, not calculated]

Unit name	Mapped area (mi^2)	Extent (mi^2)	Lithology	Thickness (ft)		
				Range	Average	Median
Benton basin	1,020	1 (unconsolidated)	Alluvial, alluvial fan, loess, terrace, dune sand, Touchet Beds, Missoula flood, and Ringold Formation deposits.	0 to 870	120	70
		2 (consolidated)	Ellensburg Formation and undefined continental sedimentary deposits	0 to 680	100	60
		Total basin thickness	All deposits	0 to 870	120	60
Columbia River Basalt Group and interbeds						
Saddle Mountains unit	2,289	457 mi^2 surface outcrop, 1,804 mi^2 below surface, 28 mi^2 not present in mapped area	Saddle Mountains Basalt flow members and interbeds	0 to 1,110	550	560
Mabton Interbed	2,206	2,179 mi^2 below surface, 27 mi^2 not present in mapped area		0 to 250	70	70
Wanapum unit	3,444	659 mi^2 surface outcrop, 2,757 mi^2 below surface, 28 mi^2 not present in mapped area	Wanapum Basalt flow members and interbeds	0 to 1,180	600	490
Vantage Interbed	3,087	3,047 mi^2 below surface, 40 mi^2 not present in mapped area		0 to 135	30	20
Grande Ronde unit	5,383	1,547 mi^2 surface outcrop, 3,786 mi^2 below surface, 50 mi^2 not present in mapped area	Grande Ronde Basalt flow members and interbeds	–	–	–
Bedrock units						
Quaternary bedrock		82	Principally volcanics, but with minor amounts of sediments	–	–	–
Tertiary		1,300	Sediments, and volcanic and plutonic rocks	–	–	–
Mesozoic		136	Metamorphic, volcanic, and plutonic rocks	–	–	–
Paleozoic		2	Metamorphic rocks	–	–	–
All bedrock units	1,520			–	–	–

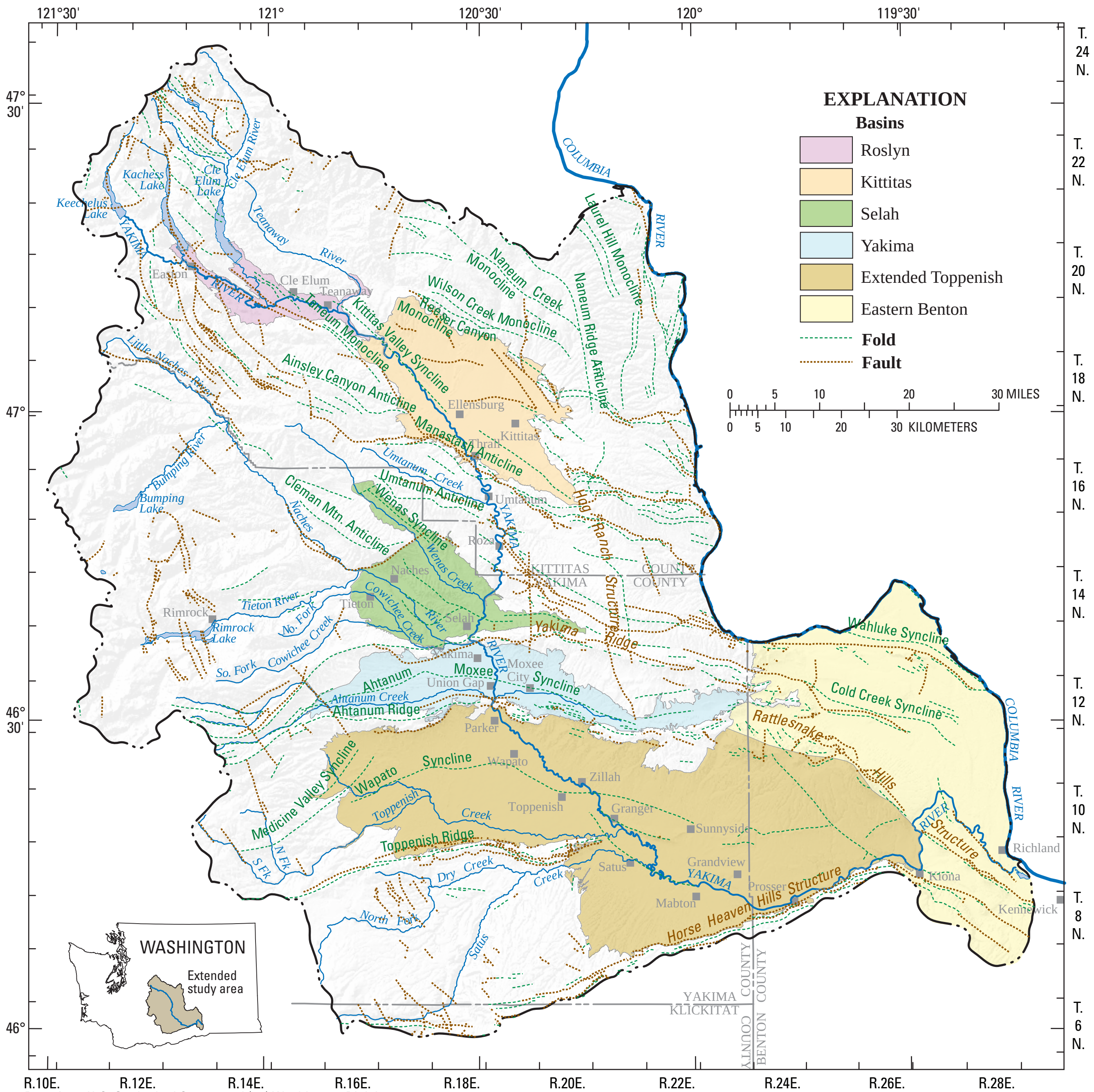
Table 2. Correlation chart showing the regional relation between generalized geologic and hydrogeologic units in the basin-fill and Columbia River Basalt Group units, and bedrock units for the Yakima River basin aquifer system, Washington.

[**Hydrogeologic unit**: B, Benton basin; R, Roslyn basin; K, Kittitas basin; S Selah basin; T, Toppenish basin; Y, Yakima basin. **Abbreviations:** Fm., formation; Mtn, mountain; Cr., creek]

BASIN-FILL AND COLUMBIA RIVER BASALT GROUP UNITS				
ERA	**PERIOD**	**EPOCH**	**SIMPLIFIED GEOLOGIC UNITS**	**HYDROGEOLOGIC UNIT**
CENOZOIC	**Quaternary**	Holocene	Alluvium, alpine glaciation, alluvial fan, dune sand, artificial fill, and peat deposits	Unit 1 (R, K, S. Y, T, B)
		Pleistocene	Alluvium, alpine glacial drift, alluvial fan, Palouse Fm., Lakedale Drift, Lookout Mountain Ranch Drift, Hayden Creek Drift, Kittitas Drift, Evans Creek Drift, unknown continental sedimentary deposits, dune sand, Missoula glacial lake deposits	Unit 1 (R, K, S, Y, T, B), Unit 2 (R, K, S, Y, T)
	Tertiary	Pliocene	Alluvial fan, Ringold Fm., Ellensburg Fm., Dalles Fm., Thorpe Gravel, and unknown continental sedimentary deposits	Unit 1 (B), Unit 2 (R, K, S, Y, T), Unit 3 (R, K, S, Y)
		Miocene	Ellensburg Fm., Ringold Fm., Dalles Fm., Snipes Mountain deposits, and unknown continental sedimentary deposits	Unit 2 (B), Unit 3 (R, K, S, Y, T), Unit 4 (T), Unit 5 (T)
			Columbia River Basalt Group: Saddle Mountains Basalt flow members and interbeds	Saddle Mountains unit (SM)
			Mabton interbed	Mabton unit
			Wanapum Basalt flow members and interbeds	Wanapum unit (WN)
			Vantage interbed	Vantage unit
			Grande Ronde Basalt flow members and interbeds	Grande Ronde unit (GR)

Table 2. Correlation chart showing the regional relation between generalized geologic units and hydrogeologic units basin-fill and Columbia River Basalt Group units, and bedrock units for the Yakima River basin aquifer system, Washington.—Continued

BEDROCK UNITS				
ERA	**PERIOD**	**EPOCH**	**SIMPLIFIED GEOLOGIC UNITS**	**HYDROGEOLOGIC UNIT**
CENOZOIC	**Quaternary**	Holocene to Pleistocene	Old Snowy Mtn. andesite, Mount Rainier andesite, Tieton andesite, Russel Ridge andesite, Round Mtn. andesite, Pear Lake andesite, Jess Lake comjplex andesite, Deep Creek andesite, Deer Lake Mtn. andesite, Swampy Meadow andesite, Signal Peak andesite, South Butte andesite, Hellroaring and Big Muddy Creek complex andesite, Mt. Adams volcanics, Tumac Mtn. basalt, Rimrock Lake basalt, Lava Creek basalt, Kincaid Lake basalt, Hogback Mtn. basalt, Canyon Creek basalt, Paradise Falls basalt, Outlaw Creek basalt, McClellan Meadows basalt, White Chuck cinder cone basalt, Walupt Lake basalt, Two Lakes Basalt, Trout Lake Creek basalt, Thomas Lake basalt, TIllicum Creek basalt, Twin Buttes basalt,Sleeping Beauty basalt, Sawtooth Mtn. basalt, Red Lake basalt, Riley Creek basalt, Rush Creek basalt, Red Butte basalt, Mosquito Creek basalt, Lakeview Mtn. basalt, Little Goose Creek basalt, Loaf basalt, Lake Comcomly basalt, Lone Butte basalt, Indian Viewpoint basalt, Indian Heaven basalt, Ice Cave basalt, Hidden Lake basalt, Goat Butte basalt, Green Canyon basalt, Gotchen Creek basalt, Glaciate Butte basalt, Flattop Mtn. basalt, East Canyon Creek basalt, Dead Horse Creek basalt, Deep Lake basalt, Camas Prairie basalt, County Park basalt, Chenamus Lake basalt, Burnt Peak basalt, Bunnell Butte basalt, Bird Mountain basalt, Blue Lake basalt, Badger Peak basalt, Placid Lake basalt, Spiral Butte dacite, Clear Fork dacite, Snyder Mtn. dacite, Olallie Lake dacite	**QB** **Quaternary Bedrock unit** (for area that overlays the older bedrock deposits and lies outside the Grande Ronde Basalt extent)
		Pliocene	Simcoe Mtn. volcanics, Devils Horns volcanics, Goat Rocks andesite, Bee Flat andesite, Lincoln Plateau basalt, Hogback Mtn. basalt, Devils Washbasin basalt, Dalles Ridge basalt, Bald Mtn., pluton, and Bethel Ridge basalt	**TB** **Tertiary Bedrock unit**
	Tertiary	Miocene	Conglomerate Point breccia, Howson andesite, Council Bluff volcanics, Clear West rhyolite, Silver Creek tonalite, Skyscraper Mtn. volcanics, Eagle tuff, Palisades tuff, Ellensburg Fm. volcanics, Cooper Pass volcaniclastics, Stevens Ridge Fm. volcanics, Fife's Peak Fm. volcanics, Tatoosh pluton, Bumping Lake pluton, Box Canyon gabbro, Carbon River stock, Nisqually diorite, Jug Lake sills, Snoqualmie Batholith, Snipes Peak Fm. sediments, White River pluton, Box Canyon volcanics	
		Oligocene	Wenatchee Fm. sediments, Chumstick Fm. andesite, Grotto Batholith, Mount Daniel volcanics, Index Batholith, Mount Aix volcanics, Eagle Gorge volcanics, Ohanapecosh Fm. volcanics, Mill Creek basalt, Wildcat Creek volcaniclastics, Spencer Creek volcaniclastics, Rattlesnake Creek tuff, Bumping River tuff, and Burnt River tuff	
		Eocene	Tukwila Fm. volcanics, Tiger Mtn. Fm., sediments, Roslyn Fm. sediments, Renton Fm. sediments, Naches Fm. sediments and volcanics, Barlow Pass volcanics, Swauk Fm. sediments and volcanics, Manatash Fm. sediments, Lookout Creek sandstone, Tenaway Basalt, Camas Land diabase, Goat Mtn. dacite, Banks Lake dacite, Fuller Mount polug, Cooper Mtn. Bartholith Raging River Fm. sediment, Mt. Persis volcanics, Taneum andesite, Summer Creek basalt, Peoh Point andesite, Frost Mtn. basalt, Chumstick Fm. volcanics, and Spencer Creek tuff, Tieton Pass basalt, Discovery Creek basalt	
		Paleocene	Swawilla Basin granite, Coffee Lake granite	
MESOZOIC	**Cretaceous**	Early to Late	Nason Ridge Gneiss, Mount Stuart Batholith, Bald Mtn. pluton, Arbuckle Mtn. tonalite, Russell Branch Fm. sediments, Ten Peak pluton, Sloan Creek pluton	**MB** **Mesozoic Bedrock unit**
	Jurassic	Early to Late	Lookout Mountain Fm. metamorphics and metavolcanics, Alta Lake metamorphics, Ingalls tectonic complex metamorphics and metavolcanics, eastern and western melange elt metavolcanics, Indian Creek complex diorite, Sarrington phyllite, Shuksan greenschist, Quartz Mtn. stock tonalite	
	Triassic	Early to Lake	Tonga Fm. metamorphics and metavolcanics, Gibraltar Rock migmatite, Chiwaukim schist	
PALEOZOIC	**Permian**	Early to Late	North Peak metavolcanics	**PB** **Paleozoic Bedrock unit**



Base from U.S. Geological Survey and (or) Washington Division of Geology and Earth Resources digital data, 1983, 1:100,000
Universal Transverse Mercator projection, Zone 10
Horizontal Datum: North American Datum of 1927 (NAD 27)
Vertical Datum: North American Vertical Datum of 1988 (NAVD 88)

Figure 7. Location of six structural basins and geologic structure, Yakima River basin aquifer system, Washington.

In the Kittitas, Selah, and Yakima basins (fig. 7), an alluvial unit (Unit 1), an unconsolidated unit (Unit 2), and a consolidated unit (Unit 3) were mapped (pl. 1); the Yakima basin also is locally called the Ahtanum-Moxee basin (Ahtanum and Moxee subbasins). In the Roslyn basin, a surficial coarse-grained aquifer unit (Unit 1), a lacustrine confining unit (Unit 2), and a confined coarse-grained aquifer unit (Unit 3) underlying the confining unit were mapped (table 2, pl. 1). In the Toppenish basin, five units were mapped that consist of coarse-grained and fine-grained unconsolidated deposits and three units comprised of part of the Ellensburg Formation. An unconsolidated unit (Unit 1) and a consolidated unit (Unit 2) were mapped in the Benton basin. In addition, in the eastern part of the Benton basin (area shown on plate 1 as a subbasin), the unconsolidated unit (the only unit present in this area) was further divided into six subunits (Jones and others, 2006). These six subunits, which occur together only in this part of the study area, consist of fluvial and lacustrine deposits of the Ringold Formation, glaciofluvial sediments of the Hanford Formation, and pre-Missoula Flood gravels. Typically, the consolidated units contain deposits that are more consolidated (for example, sandstone) than the deposits of the unconsolidated units (for example, recent sand and gravel).

The structural-basin boundaries (fig. 7) are slightly different than those shown in Jones and others (2006) in order to provide an improved framework for constructing a groundwater flow model. These differences include an extension of the Toppenish basin to include the western part of the Benton basin where both unconsolidated and consolidated units are present. Also, the eastern part of the Benton basin containing the six unique subunits was extended eastward to the Columbia River, to include about 217 mi^2 outside the Yakima River basin. This area generally coincides with the Hanford Site and is part of what is herein called the extended study area. The extents of the six eastern Benton basin subunits were not mapped during this study but were obtained from S.P. Reidel and P.D. Thorne (Battelle Institute, written commun., 2003 and 2005). To distinguish these extended areas from the previously defined structural basins, they are referred to as the extended Toppenish basin and the Eastern Benton basin (fig. 7).

The unconsolidated units include alluvial, alluvial fan, terrace, glacial, loess, lacustrine, and flood (Touchet Beds) deposits that range from coarse-grained gravels to fine-grained clays, with some cemented gravel (Thorp gravel and similar unnamed gravels). Most of the unconsolidated units consist of coarse-grained deposits.

The deposits that constitute the consolidated units are principally deposits of the Ellensburg Formation, but also include some undifferentiated continental sedimentary deposits. These units include continental sandstone, shale, siltstone, mudstone, claystone, clay, and lenses or layers of uncemented and weakly to strongly cemented gravel and sand (conglomerate). These clastic deposits are one of the most stratigraphically complex parts of the aquifer system. In the structural basins where these deposits overlie bedrock they were either mapped as one hydrogeologic unit (called the consolidated unit, for example, Unit 3 in the Kittitas, Selah, and Yakima basins and Unit 2 in the Benton basin) or they were subdivided into several hydrogeologic units as in the Toppenish basin (Jones and others, 2006, table 2). Except for the Mabton and Vantage Interbeds, where the Ellensburg Formation is interbedded with the CRBG, it was considered part of the CRBG hydrogeologic units (Jones and Vaccaro, 2008). The Mabton and Vantage Interbeds are considered separate hydrogeologic units and the depth to the top of these units was mapped by Jones and Vaccaro (2008).

The lithology of the consolidated units varies from the Kittitas basin to the extended Toppenish basin; consolidated units are absent in the Roslyn and Eastern Benton basins. The variations are due to spatial-temporal variations in deposition, erosion, and structural deformation (faulting and folding). Examples of the spatial variations in lithology based on seven well logs are presented below. In the Kittitas basin, at well 17N/18E-01C01, which yields more than 1,000 gal/min, the consolidated unit (Unit 3) is overlain by the unconsolidated unit (Unit 2) that is about 400 ft thick. This 1,209-ft deep well penetrates about 800 ft of clay, sand, and gravel of the consolidated unit, and only 6 ft of shale was penetrated near the bottom. In contrast, in the Selah basin, a 1,955 ft well (15N/17E-36A01) with a yield of 900 gal/min penetrates more than 1,800 ft of the consolidated unit (Unit 3). In the upper section of this well is a 411-ft thick sandstone section underlain by about 1,110 ft of mainly clay with lenses of sand and gravel and some conglomerate; about one-half of the remaining deposits are sandstone. A nearby 1,200-ft well (15N/17E-27A) penetrates predominantly sandstone (with some cemented gravel layers) from about 38 ft to the bottom of the well; the driller reported only a trace of water (5–6 gal/min) at 1,100 ft. In the Yakima basin, the consolidated unit in a 1,200-ft deep well (13N/18E-35K01) is overlain by about 650 ft of sand, gravel, and cobbles with minor amounts of clay. From about 650 to 1,032 ft, the consolidated unit consists principally of clay and layers of sand and gravel, and only a 5-ft interval in which a shale/sandstone layer was penetrated. Clay and sandstone were penetrated from 1,032 to 1,075 ft before the CRBG was reached. In contrast, in the Moxee subbasin of the Yakima basin, a 637-ft well (12N/20E-8F01) penetrated clay, sandstone, and shale to its final depth and only minor amounts of interbedded sands and gravels were present. Lastly, near the central part of the Toppenish basin, a 1,471-ft well (11N/18E-26M03), which reached basalt at 960 ft, penetrated about 110 ft of gravel, boulders, and sand with some clay before penetrating 851 ft of clay with layers of sand and gravel intermixed with clay (shale and sandstone where absent). Well 11N/20E-12H01 (1,168-ft deep) in the Toppenish basin penetrated mainly shale, sandstone, and clay before reaching basalt at 758 ft.

The other two categories of units are the Miocene age CRBG units and Paleozoic to Quaternary bedrock units (table 2). The CRBG units form the major, productive volcanic-rock part of the aquifer system. The generally older, non-CRBG bedrock units (herein called bedrock units) are present mainly along the northern, northwestern, and western margins of the basin (fig. 5) and include volcanic, intrusive, marine and nonmarine sedimentary, and metamorphic rock units. In a few areas at the western margins of the basin in the High Cascades, younger rock units composed principally of Quaternary volcanics units are present (table 2). These include the Tieton andesite and the Mount Adams volcanics that were derived from the volcanically active Cascade Range.

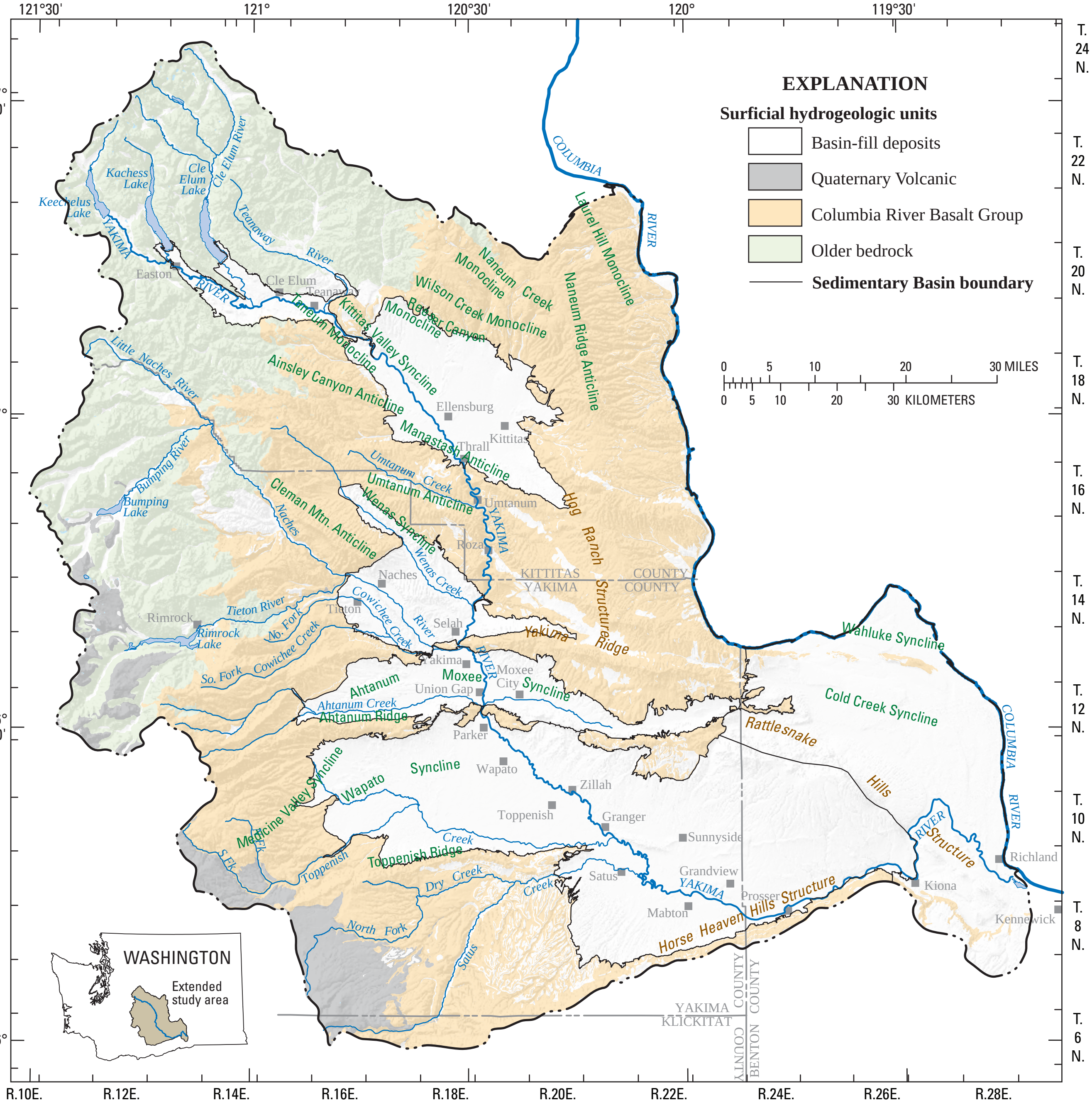
The CRBG contains three formations in the study area, from oldest to youngest, the Grande Ronde, Wanapum, and Saddle Mountains Basalts (table 2). The Grande Ronde Basalt is the most extensive, consisting of 85 to 88 percent of the total volume of the CRBG and the Saddle Mountains Basalt is the least extensive, consisting of less than 2 percent of the total volume of the CRBG (Reidel, 1982; Tolan and others, 1989). The younger Saddle Mountains Basalt contains the thickest and most sedimentary interbeds, especially in the Yakima Fold Belt, because of its episodic eruptions and proximity to source material. For example, a 773-ft deep well (11N/21E-34C01) penetrated the Saddle Mountains Basalt from 17 to 70 ft and then penetrated clay, sand, sandstone, and gravel to 587 ft, followed by 34 ft of basalt that was underlain by 152 ft of interbeds. A nearby well (11N/21E-36J01) penetrated 400 ft of sand and clay before penetrating 50 ft of the Saddle Mountains Basalt; 270 ft of sand and clay underlies the basalt to the final depth of 720 ft. There are locations where the interbeds in the Saddle Mountains Basalt comprise more than 50 percent of the total thickness of the formation.

Each CRBG formation was defined as a hydrogeologic unit (table 2). The formations are called units because they include interbeds. Each unit includes a thick sequence of basalt flows that overlap and intermingle. In turn, individual basalt flows comprise the members of the formations. The depth to the top of each unit was mapped by Jones and Vaccaro (2008), and the configuration of the top of each unit is very complex, as indicated by the altitude of unit tops (pl. 2). The Mabton and Vantage Interbeds also were defined as hydrogeologic units and depths to their tops were mapped by Jones and Vaccaro (2008). The Mabton unit lies between the Saddle Mountains and Wanapum units and the Vantage unit separates between the Wanapum and Grande Ronde units. Neither the Mabton nor the Vantage units are present throughout the extent of the overlying CRBG units. The maps showing the depth to the tops include the area east of the Yakima River basin boundary to the Columbia River, an area of about 700 mi^2. This additional area, combined with the Yakima River basin, is called the extended study area and encompasses about 6,900 mi^2. The extended study area thus is bounded on the east by the Columbia River, a hydrologic boundary in the regional groundwater flow model that is being constructed as part of this study.

Discontinuous and generally small patches of CRBG formations were not combined with underlying formations because WaDOE generally manages withdrawals from the CRBG on the basis of formations. Within the extended study area, the lateral extent of the Grande Ronde unit is about 5,383 mi^2 (its thickness was not estimated due to insufficient data), the extent of the Wanapum unit is about 3,444 mi^2 and has an average thickness of about 600 ft, and the Saddle Mountains unit extent is about 2,289 mi^2 and has an average thickness of about 550 ft (table 1).

Four bedrock units were defined on the basis of their age. These units are called the Paleozoic unit, the Mesozoic unit, the Tertiary unit, and the Quaternary bedrock unit. The Paleozoic unit consists of metamorphic rocks and the Mesozoic unit has three subdivisions consisting of metamorphic, volcanic, and plutonic rocks. The Tertiary unit also has three subdivisions consisting of sedimentary, volcanic, and plutonic rocks. The Quaternary bedrock unit consists principally of volcanics, but in a few small areas it consists of sediments (pl. 2). Where not overlain by younger deposits, the surficial extents of these bedrock units were obtained from the digital GIS database for Washington (Washington State Department of Natural Resources, 2002). Neither thickness nor vertical subdivisions for these units were mapped due to a lack of information. The formations that comprise these units, such as the nonmarine sandstone of the Eocene age Roslyn and Swauk Formations, which are separated by the Teanaway Basalt, and the interbedded sandstone-basalt Naches Formation are as much as 5,000 ft thick (Kinnison and Sceva, 1963; Campbell, 1989). The combined lateral extents of these units where they abut the CRBG are about 1,520 mi^2. Within this area, these units either crop out (pl. 2) or underlie unconsolidated and (or) consolidated deposits. The Mesozoic and Tertiary units also underlie the CRBG and the Quaternary bedrock unit overlies either the CRBG or the Mesozoic and Tertiary units.

The lateral extents of the hydrogeologic units are shown on plates 1 and 2, and their surficial extents are shown in figure 8. For the CRBG units, plate 2 also shows the altitude (a shaded relief map) of the top of the unit, and where a unit outcrops, its altitude will be land surface. Note that the basin-fill units in the structural basins are aggregated for display purposes (detailed extents of these units are shown on plate 1 and the mapped thickness of units are shown in Jones and others (2006)). The area shown on plate 2 and figure 8, which is the 6,900 mi^2 extended study area, includes both the Eastern Benton basin and the area outside of the Yakima River basin where the basalt units were mapped. Deposits that are similar to the basin-fill deposits outside of the six structural basins are included on figure 8 to show their limited extent. These deposits are considered part of the underlying hydrogeologic unit and they are not important components of the regional groundwater flow system. Information on the area and thickness of the units and their components (table 1) and the correlation chart (table 2) indicate the complex make-up of the Yakima River basin aquifer system.



Base from U.S. Geological Survey and (or) Washington Division of Geology and Earth Resources digital data, 1983, 1:100,000
Universal Transverse Mercator projection, Zone 10
Horizontal Datum: North American Datum of 1927 (NAD 27)
Vertical Datum: North American Vertical Datum of 1988 (NAVD 88)

Figure 8. Extent of surficial hydrogeologic units, Yakima River basin aquifer system, Washington.

Hydraulic Characteristics of Units

The ability of sediments and rocks to store and transmit groundwater (their hydraulic characteristics) determines how a groundwater-flow system functions. Knowledge of the hydraulic characteristics also is necessary to evaluate how the flow system responds to stresses such as pumpage. These characteristics include lateral and vertical hydraulic conductivity and the storage coefficient. Estimates of characteristics from previous studies and this study are described below by categories of hydrogeologic units. Information described below provides a general overview of the range and median of hydraulic characteristic values for the hydrogeologic units.

Lateral Hydraulic Conductivity

Lateral hydraulic conductivity (referred to here as K_h) is a measure of a material's ability to transmit water laterally. It is expressed in units of cubic feet per square feet per day—simplified to feet per day (ft/d). Values of K_h can be estimated from specific-capacity data reported on drillers' logs, or determined from aquifer tests or groundwater flow modeling. Numerous studies conducted in the Yakima River basin or in the surrounding area have calculated or compiled information on K_h (table 3).

Basin-Fill Units

Basin-fill deposits are diverse in lithology and, thus, so are their hydraulic characteristics (table 3). The deposits, which consist of unconsolidated and consolidated material of fluvial, glacial, lacustrine, and volcanic origins, form important water-bearing units, as well as semiconfining to confining units. Estimates of effective K_h for the complete thickness of the basin-fill deposits ranged from 0.02 to 150,000 ft/d. This large range in estimated values is due to the large variation in grain size, depositional regimes, and age of the deposits. The deposits include shale, sandstone, clay, sand, gravel, and cobbles.

The median reported values were in the range of 40 to 240 ft/d. K_h of loess, which mantels part of the study area, is between 1 to 10 ft/d. The Touchet Beds, which also mantel part of the study area, have estimated K_h values between 8 and 11 ft/d. Values of K_h for the alluvium range from 6 to 100,000 ft/d and generally average between 100 and 800 ft/d. The large range is due to the variation in grain size (silty sands to cobbles) of the alluvium. In turn, grain-size varies complexly throughout the area.

Estimates of K_h for the Pasco gravels that occur in the Eastern Benton basin ranges from about 48 to 73,000 ft/d, with a median reported value from about 880 to 1,250 ft/d. The reported values for the upper, middle, lower, and basal Ringold Formation range over several orders of magnitude, which reflects the wide-range in local conditions. Effective median values range from about 0.1 to 25 ft/d for the upper Ringold, from about 40 to 250 ft/d for the middle Ringold, and from about 0.1 to 4 ft/d for the lower Ringold; the median K_h for the basal Ringold is about 200 ft/d.

K_h of the water-producing parts of the Ellensburg Formation ranges from about 0.01 to 2,265 ft/d. Average values reported for the Ellensburg range from about 0.1 to 72 ft/d. The large range in the average values is due to the variations in the types of materials composing the water-producing zones in the Ellensburg Formation. These materials range from sandstone to uncemented sands and gravels.

No information is available on the K_h of the claystone or clay sections of the Ellensburg Formation but Neuzil (1994) indicates that values may be as small as 10^{-7} to 10^{-4} ft/d. Neuzil (1994) also indicates that typical values of K_h for shale, which is present in parts of the Ellensburg Formation, is on the order of 10^{-8} to 10^{-7} ft/d.

Columbia River Basalt Group Units

Hydraulic characteristics vary greatly within and between the individual basalt flows, members, and hydrogeologic units (table 3). Upper zones of the flows were exposed to weathering processes and were broken by subsequent flows, resulting in the formation of conductive "flow tops." These flow tops, when combined with the base of the overlying basalt flow, form interflow zones that generally exhibit high K_h (Lindolm and Vaccaro, 1988). In general, the flow tops are brecciated and (or) vesicular and the flow bases are brecciated and may contain pillow complexes if the basalt was extruded within or flowed into water. The interflow zones are separated by the less transmissive entablature and colonnade in which the fractures are typically vertically oriented (Tomkeieff, 1940; Waters, 1960; MacDonald, 1967; Swanson and Wright, 1978; Sublette, 1986; Hansen and others, 1994; Whiteman and others, 1994). The fractures are a result of differential contraction during cooling of basalt flows (MacDonald, 1967; Long and Wood, 1986) and of later folding and faulting. The greatest density and lowest K_h generally occur in the interior or middle of a basalt flow, typically the entablature (Wood and Fernandez, 1988; Reidel and others, 2002). Observations of exposed colonnades, which typically are three- to eight-sided columns of basalt, suggest that there would be lateral connectivity along the columns; springs have been observed emanating from exposed colonnades. Many of the physical characteristics of the CRBG described above are similar to those of the Snake River Group (Lindolm, 1996), but K_h is much higher for the Snake River Group because the older CRBG flows have been affected by fracture filling with mineral precipitates (such as smectite and clinoptilolite) that decreases K_h (Wood and Fernandez, 1988; Steinkampf and Hearn, 1996). Fractures tend to be filled with these alteration products, whereas vesicles typically are only partly filled (Steinkampf and Hearn, 1996). Such alteration products are well documented (Ames, 1980; Benson and Teague, 1982; Hearn and others, 1985; Steinkampf and Hearn, 1996).

Table 3. Summary of previous estimates of lateral hydraulic conductivity, Yakima River basin aquifer system, Washington.

[**Abbreviations**: ft, foot; –, not applicable; >, greater than]

Unit or deposit	Basin or area	Lateral hydraulic conductivity, in feet per day					Source
		Minimum	Median	Mean	Maximum	Model derived	
Alluvium	Toppenish	95.0	–	–	778	–	Bolke and Skrivan, 1981
	Ahtanum-Moxee	–	–	–	–	284	Golder Associates, 2002
	Pasco/Benton	10	–	275	1,000	–	Zimmerman, 1983
Alluvium, fluvial gravels, coarse sands	Hanford	6	–	–	100,000	–	Vermeul and others, 2001
Alluvium, coarse to medium grained (some fines and caliche)	Hanford/Black Rock	–	–	.85	–	–	Didricksen, 2004
Alluvium, unconfined	Ellensburg	–	–	2,000	–	–	GeoEngineers, 1999
Alluvium, unconsolidated	Ahtanum Valley	668	–	13	802	–	Foxworthy, 1962
Alluvium, young	Lower Satus Creek	–	–	–	–	864	Prych, 1983
Alluvium/Ellensburg (upper)	Ahtanum-Moxee	–	–	–	–	75	Golder Associates, 2002
Alluvium, old	Lower Satus Creek	–	–	–	–	86	Prych, 1983
Overburden aquifer	Columbia Plateau	.04	43	99	1,100	–	Hansen and others, 1994
	Columbia Plateau	.02	240	–	150,000	–	Whiteman and others, 1994
Loess	Columbia Plateau	–	–	–	–	5	Lum and others, 1990
	Columbia Plateau	1	–	–	10	–	Vaccaro, 1999
Plio-Pleistocene Unit	Hanford	.02	–	–	57	–	Khaleel and Freeman, 1995
Sediments	Horse Heaven Hills	–	–	24,192	–	–	Davies-Smith and others, 1989
Ellensburg, upper (old Alluvium)	Lower Satus Creek	.9	72	–	3,024	–	Prych, 1983
Ellensburg, upper	Ellensburg	2	–	–	–	53	GeoEngineers, 1999
Ellensburg, upper (all)	Ahtanum-Moxee	.6	–	13	2,265	–	Golder Associates, 2002
Ellensburg, upper (upper)	Ahtanum-Moxee	–	–	–	–	7.5	Golder Associates, 2002
Ellensburg, upper (upper 300 feet)	Ellensburg	–	–	–	–	32	GeoEngineers, 1999
Ellensburg, upper (upper and middle)	Ahtanum-Moxee	.6	–	12	2,265	–	Golder Associates, 2002
Ellensburg, upper (middle)	Ahtanum-Moxee	–	–	–	–	.002	Golder Associates, 2002
Ellensburg, upper (lower)	Ahtanum-Moxee	38	–	60	96	–	Golder Associates, 2002
	Ahtanum-Moxee	–	–	–	–	8	Golder Associates, 2002
	Ahtanum-Moxee	–	–	–	–	.1	Golder Associates, 2002
	Ahtanum-Moxee	–	–	–	–	.01	Golder Associates, 2002
Touchet beds	Pasco/Benton	8	–	–	11	9	Drost and others, 1997
Soil buried horizon, caliche, gravel	Hanford	47	–	–	625	–	Vermeul and others, 2001
Soils, Palouse	Hanford	.2	–	–	.4	–	Khaleel and Freeman, 1995

Table 3. Summary of previous estimates of lateral hydraulic conductivity, Yakima River basin aquifer system, Washington.—Continued

[**Abbreviations**: ft, foot; –, not applicable; >, greater than]

Unit or deposit	Basin or area	Lateral hydraulic conductivity, in feet per day					Source
		Minimum	Median	Mean	Maximum	Model derived	
Pasco gravel	Pasco/Benton	48	880	–	73,000	880	Drost and others, 1997
	Pasco/Benton	–	1,250	–	–	–	Meyers and others, 1985
Hanford Formation	Hanford	3	–	–	3,281,000	–	Wurstner and others (1995): Thorne and Newcomer (1992)
Ringold	Hanford/Black Rock	–	2.6	–	–	–	Didricksen, 2004
Ringold, upper	Pasco/Benton	2	25	–	210	5	Drost and others, 1997
	Hanford	–	–	–	–	.02	Vermeul and others, 2001
	Hanford	.001	–	–	.3	–	Wurstner and others (1995); Thorne and Newcomer (1992)
Ringold, middle	Pasco/Benton	200	–	–	250	–	Bergeron and others, 1986
	Pasco/Benton	8	180	–	5,000	200	Drost and others, 1997
	Pasco/Benton	–	41	–	–	–	Newcomb and others (1972); Graham (1981); Meyers and others (1985)
	Hanford	.1	–	–	5,500	–	Vermeul and others, 2001
	Hanford	.03	–	–	281	–	Vermeul and others, 2001
	Hanford	.03	–	–	.3	–	Vermeul and others, 2001
	Hanford	.3	–	–	656	–	Wurstner and others (1995); Thorne and Newcomer (1992)
	Hanford	.001	–	–	.3	–	Wurstner and others (1995); Thorne and Newcomer (1992)
	Hanford	.3	–	–	656	–	Wurstner and others (1995); Thorne and Newcomer (1992)
Ringold, middle, gravel facies	Hanford	2,000	–	–	6,400	–	Vermeul and others, 2003
Ringold, middle, sand facies	Hanford	–	–	1,300	–	–	Vermeul and others, 2003
Ringold, middle, silt facies	Hanford	–	–	2	–	–	Vermeul and others, 2003
Ringold, lower	Pasco/Benton	2	46	–	230	4	Drost and others, 1997
	Hanford	.001	–	–	.3	–	Wurstner and others (1995); Thorne and Newcomer (1992)
Ringold, lower, basal, clay/mud	Hanford	–	–	–	–	.00003	Vermeul and others, 2001
Ringold, basal	Pasco/Benton	200	–	–	250	–	Bergeron and others, 1986
	Hanford	.3	–	–	656	–	Wurstner and others (1995); Thorne and Newcomer (1992)

Table 3. Summary of previous estimates of lateral hydraulic conductivity, Yakima River basin aquifer system, Washington.—Continued

[**Abbreviations:** ft, foot; –, not applicable; >, greater than]

Unit or deposit	Basin or area	Lateral hydraulic conductivity, in feet per day					Source
		Minimum	Median	Mean	Maximum	Model derived	
Ringold, basal, sand/gravel	Hanford	0.03	–	–	700	–	Vermeul and others, 2001
Rattlesnake Ridge Interbed	Hanford/Black Rock	–	–	0.80	–	–	Didricksen, 2004
	Hanford	.05	–	8.4	210	–	Strait and Mercer, 1987
Selah Interbed	Hanford/Black Rock	–	–	2.7	–	–	Didricksen, 2004
Saddle Mountains	Horse Heaven Hills	–	–	18	–	–	Davies-Smith and others, 1988
	Pasco/Benton	.007	2	–	3,200	–	Drost and others, 1997
	Columbia Plateau	.2	1	1	3	–	Hansen and others, 1994
	Columbia Plateau	.007	2	–	1,892	–	Whiteman and others, 1994
	Pasco/Benton	–	–	–	–	1	Zimmerman, 1983
	Hanford/Black Rock	.04	–	–	2.6	–	Didricksen, 2004
	Hanford	.04	2	10	53	–	Schmidt and others, 2007
Saddle Mountains anticline	Horse Heaven Hills	.03	–	–	.5	–	Packard and others, 1996
	Horse Heaven Hills	.2	–	–	78	–	Packard and others, 1996
Mabton Interbed	Hanford/Black Rock	–	–	.025	–	–	Didricksen, 2004
Wanapum	Horse Heaven Hills	–	–	170	–	–	Davies-Smith and others, 1988
	Pasco/Benton	1.1	11	–	310	–	Drost and others, 1997
	Columbia Plateau	.9	–	–	1	–	Lum and others, 1990
	Columbia Plateau	.09	3	3	8	–	Hansen and others, 1994
	Horse Heaven Hills	.1	–	–	259	–	Packard and others, 1996
	Columbia Plateau	.007	5	–	5,244	–	Whiteman and others, 1994
	Pasco/Benton	–	–	–	–	5	Zimmerman, 1983
Wanapum, interflow	Hanford	–	2.8	–	–	–	Reidel and others, 2002
Wanapum, flow interior	Hanford	–	.000003	–	–	–	Spane and Thorne, 1985
Grande Ronde	Horse Heaven Hills	–	–	65	–	–	Davies-Smith and others, 1988
	Columbia Plateau	1	–	–	12	–	Lum and others, 1990
	Columbia Plateau	.005	5	–	5,222	–	Whiteman and others, 1994
	Pasco/Benton	–	–	–	–	.02	Zimmerman, 1983
Grande Ronde, top	Hanford	.000003	0.1	–	500	–	Strait and Mercer, 1987
Grande Ronde, 1–2,000 ft thick	Columbia Plateau	.1	1	2	9	–	Hansen and others, 1994
Grande Ronde, flow interior	Hanford	5×10^{-9}	.0000002	–	.001	–	Strait and Mercer, 1987
	Hanford	1×10^{-8}	.00000014	–	.000017	–	Eslinger, 1986

Table 3. Summary of previous estimates of lateral hydraulic conductivity, Yakima River basin aquifer system, Washington.—Continued

[**Abbreviations**: ft, foot; –, not applicable; >, greater than]

Unit or deposit	Basin or area	Lateral hydraulic conductivity, in feet per day					Source
		Minimum	Median	Mean	Maximum	Model derived	
Grande Ronde, interflow	Hanford	–	0.03	–	–	–	Reidel and others, 2002
Grande Ronde, > 2,000 ft thick	Columbia Plateau	0.1	1	2	6	–	Hansen and others, 1994
Basalt	Ahtanum-Moxee	–	–	–	–	5	Golder Associates, 2002
	Columbia Plateau	.9	–	–	9	–	Bolke and Vaccaro, 1981
	Columbia Plateau	.005	5	–	6,100	–	Whiteman and others, 1994
Basalt, top	Pasco/Benton	–	35	–	–	–	Drost and others, 1997
Basalt, center	Pasco/Benton	–	.0000002	–	–	–	Drost and others, 1997
Basalt, interflow	Hanford	.0000003	–	–	2,800	–	Reidel and others, 2002
Basalt, flow interior	Hanford	3×10^{-10}	.0000003	–	.003	–	Reidel and others, 2002
Basalt, weathered	Ahtanum-Moxee	–	–	–	–	.003	Golder Associates, 2002
Basalt, undisturbed	Pasco/Benton	.03	–	–	8	–	Zimmerman, 1983
Basalt, anticline	Pasco/Benton	.004	–	–	.3	–	Zimmerman, 1983
Basalt, fault	Pasco/Benton	.0001	–	–	.01	–	Zimmerman, 1983
Basalt, barrier (flow impediment)	Pasco/Benton	.002	–	–	.05	–	Zimmerman, 1983
Bedrock, crystalline	General	.000003	–	–	.3	–	Trainer, 1988
Bedrock, unfractured metamorphic and igneous rocks	General	.00000003	–	–	.00003	–	Freeze and Cherry, 1979

The Saddle Mountains K_h has been estimated to range from 0.007 to 3,200 ft/d, but results from previous studies indicate that the median is about 1 to 2 ft/d. The Wanapum had a slightly larger range (0.007 to 5,244 ft/d) than the Saddle Mountains, and the median reported K_h for the Wanapum ranges from about 3 to 11 ft/d. The range in Grande Ronde K_h values was similar to that for the Wanapum, from 0.005 to 5,222 ft/d. Median values for the Grande Ronde range from about 0.1 to 5 ft/d. Previous work indicates the CRBG K_h is one to two orders of magnitude smaller along anticlines and in deeper parts of the Grande Ronde (Hansen and others, 1994; Packard and others, 1996; Reidel and others, 2002) than in other parts of the units due to overburden pressure and secondary mineralization. For anticlines, Hansen and others (1994) reduced K_h values by multiplying the values by factors ranging from 0.01 to 0.018. For all types of flow barriers, the median multiplication factor was 0.18 (Hansen and others, 1994).

Strait and Mercer (1987) found K_h in basalt flow tops to be as much as five orders of magnitude greater than in basalt flow centers. Poeter (1980) reported that representative effective porosities ranged from about 0.02 in the flow interiors to about 0.14 in the interflow zones, further indicating the difference between K_h in interflow zones and flow interiors. Based on the median K_h values for the CRBG units as a whole (described above), assuming that about 10 percent of a unit consists of interflow zones, and using a depth-weighted average yields a K_h for interflow zones of about 10 to 100 ft/d. If the interflow zones comprised 5 percent of a unit then those estimates would double (20–200 ft/d). Hydraulic testing of interbeds in the basalts has been limited, but reported values range from 0.05 to 210 ft/d and appear to average between about 0.1 and 8 ft/d.

Bedrock Units

The headwater areas of the Yakima River basin in the Cascade geologic province include metamorphic (crystalline), sedimentary, volcanic, and intrusive and extrusive igneous rocks (table 2). Relatively little groundwater development of the bedrock units has occurred and therefore, little is known about their hydraulic characteristics. In general, the older, plutonic, metamorphic, volcanic, and sedimentary bedrock units have lower values of porosity and permeability than the basin-fill and CRBG units. Water-producing zones are variable, but are present in the bedrock units.

Typical values of K_h for unfractured metamorphic and igneous rocks range from 3×10^{-8} to 3×10^{-5} ft/d (Freeze and Cherry, 1979). K_h values for fractured metamorphic and igneous rocks can be five orders of magnitude larger than for unfractured rock (about 0.001–1 ft/d). Joints within crystalline rock are of limited lateral extent but are numerous enough to increase permeability locally, and such fractures are commonly tighter and less abundant with increasing depth due to the state of stress in the earth's crust (Trainer, 1988). The latter conditions may be important in the study area because of the existing tectonic stress that would tend to decrease fracture openings. Previous studies of well yields in crystalline rocks suggest networks of open joints are found principally within 300–500 ft of the surface and decline lognormally with increasing depth (Trainer, 1988). Trainer also reported that K_h values for crystalline rocks range from $3\text{x}10^{-6}$ to 0.3 ft/d. Thus, K_h of the Paleozoic and Mesozoic units (predominantly metamorphic rocks—gneiss, schist, phyllite, and amphibolite) likely is quite small; many well yields are reported on driller's logs as less than 1 gal/min or 'no water.'

The older sandstone and basalt parts of the Tertiary units generally yield less water than most of the sandstone of the Ellensburg Formation and the basalt of the CRBG units. Similar to the well yields in the Paleozoic and Mesozoic units, many yields for the sandstone and basalts also are reported on driller's logs as less than 1 gal/min or 'no water.' For the sandstone, proximity to the Cascade uplift and volcanism may have reduced pore space through heating, resulting in reduced K_h values. For the older basalts, tectonic stress and secondary mineralization also may account for reduced K_h values. Thus, on the basis of information for the CRBG and the Ellensburg Formation, and values in Freeze and Cherry (1979), K_h for the sandstone and basalt are likely on the order of 0.001 to 1 ft/d.

Few wells are completed in the Quaternary bedrock unit to assess its K_h values. The unit probably is permeable, however, based on information for similar rock types of similar ages, for example, the Quaternary basalts of the eastern Snake River Plain aquifer system. The Snake River basalts have K_h values ranging from 1 to more than 7,000 ft/d (Garabedian, 1986; Lindolm, 1996). A smaller range in values, typically from 1 to 20 ft/d, for the Quaternary basalts in Oregon has been estimated by Ingebritsen and others (1992) and Manga (1996, 1997).

Estimates from Specific-Capacity Data

Lateral hydraulic conductivities were estimated as part of this study using available data for water-level change (drawdown) and discharge rate (well yield) for wells pumped for periods that ranged from less than 1 hour to 200 hours. Data from wells that had a driller's log containing a discharge rate, duration of pumping, drawdown, static water level, well-construction data, and lithologic log were used. The methods and assumptions for calculating K_h are described in appendix A. Assumptions for calculating K_h from specific-capacity data generally are not strictly met resulting in values that may be considered rough estimates. Spatial variations can provide indications of patterns of K_h, however, and the availability of many values allows for a reasonable estimate of a median value. The limitations of using specific-capacity tests are described by Gannett and Lite (2004) and Meier and others (1999).

Well-yield data are needed not only to estimate K_h but also to improve the understanding of the potential productivity of a hydrogeologic unit. Available data indicate that yields in the bedrock units generally are about an order of magnitude smaller than those in the basin-fill and CRBG units (fig. 9). The average yield per foot of total well depth for the bedrock, CRBG, and basin-fill units was 0.1, 0.9, and 2.8 (gal/min)/ft, respectively. The largest yields in the basin-fill sediments and CRBG are found along the boundaries of the Yakima, Toppenish, and Benton basins. Most of the wells with large yields are irrigation wells and secondarily, municipal wells, because the well depth, well diameter, and number of water-producing zones are optimized to meet the large water demands of irrigation and public supply.

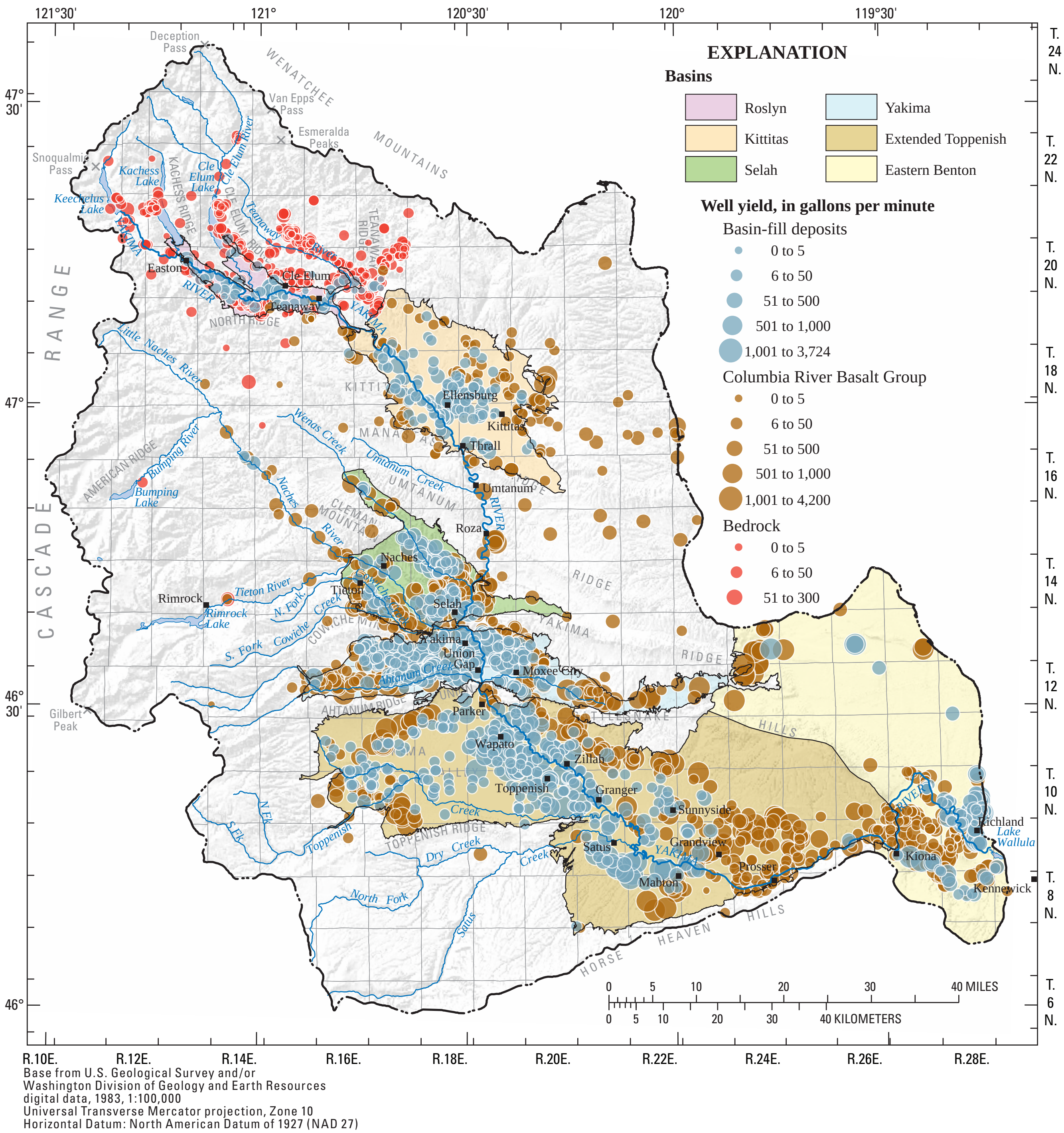


Figure 9. Reported yields of wells, Yakima River basin aquifer system, Washington.

Statistical summaries of K_h values calculated from the specific-capacity data, by hydrogeologic unit, are shown in table 4. The median values are similar in magnitude to values reported by Freeze and Cherry (1979) for similar materials and to values described previously. The information in table 4 also indicates that, overall, the values for the basin-fill and CRBG units are similar. Note that the bedrock values are based on a limited number of tests, and data for wells identified as 'dry' or little production were not included, resulting in a likely bias to larger values. In addition, bedrock values were not separated by hydrogeologic unit and different bedrock types have different effective K_h values. For example, the phyllite rocks of the Mesozoic unit are known to exhibit very low $K_{h,}$ with well yields generally less than 1–2 gal/min. The minimum values illustrate that zones of low K_h are present in most units, and the large range in K_h (fig. 10) indicates substantial heterogeneity in the units.

Table 4. Summary of lateral hydraulic conductivity values, estimated from specific-capacity data, Yakima River basin aquifer system, Washington.

Hydrogeologic unit	Number of wells	Hydraulic conductivity, in feet per day					
		Mean	Median	Minimum	25th percentile	75th percentile	Maximum
Basin-fill units	882	182	6	0.01	0.4	49	17,715
Columbia River Basalt Group units	833	182	3	.1	.4	25	48,787
Bedrock units	9	13	3	.01	.3	33	40

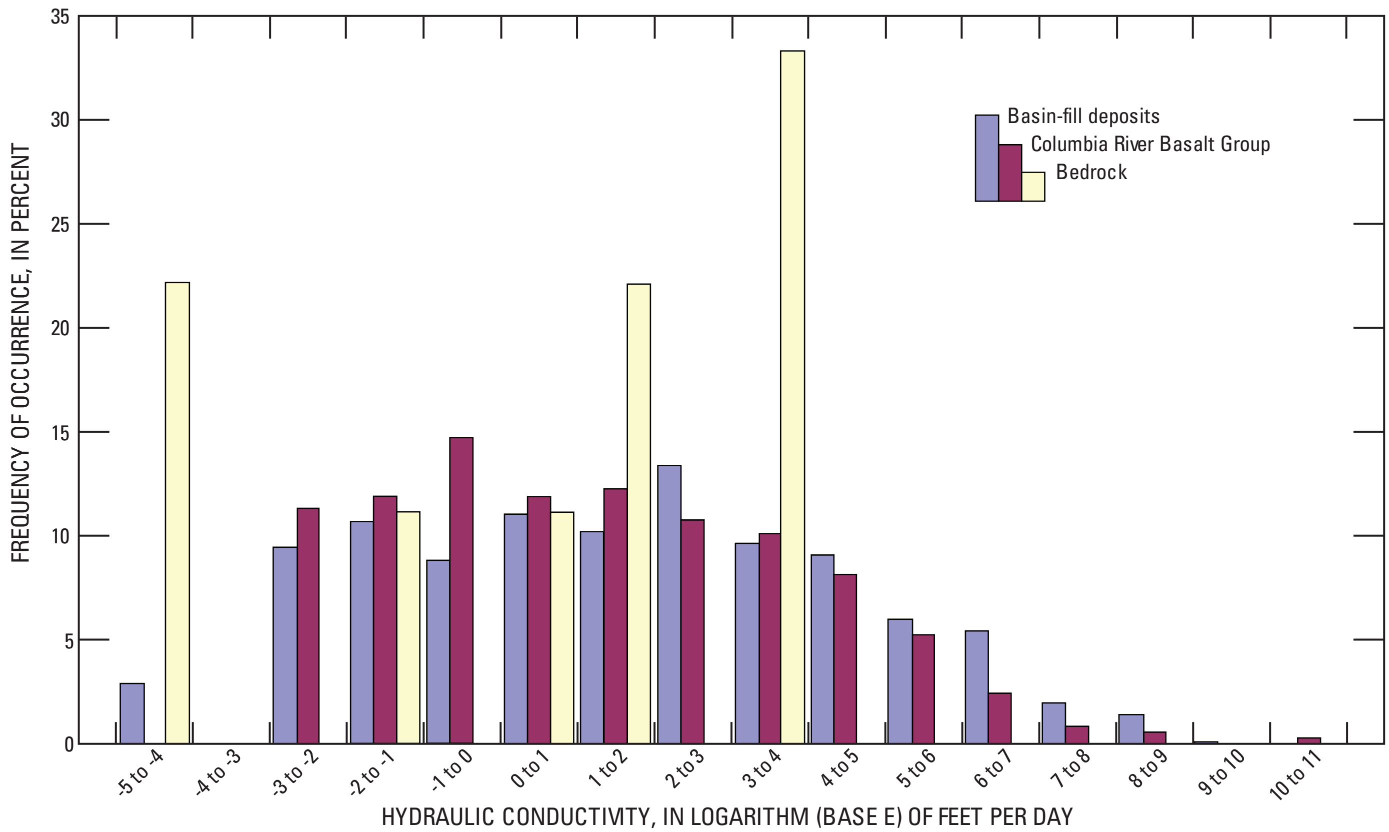


Figure 10. Frequency distribution of lateral hydraulic conductivity values estimated from specific-capacity data, Yakima River basin aquifer system, Washington.

Estimates from Aquifer Tests

K_h values were calculated using aquifer-test data for 99 wells (49 basin-fill wells and 50 CRBG wells). In all tests, drawdowns were for the pumping well. Calculations were made using the Cooper-Jacob time-drawdown (Cooper and Jacob, 1946), and Theis and Theis recovery (Theis, 1935) (see also Bentall, 1963) methods. All methods have certain assumptions, many of which are not strictly met. The full theoretical background of each method, including assumptions and limitations, can be found in the original documentation.

The mean and median K_h values for the basin-fill units were 167 and 26 ft/d, respectively, and for CRBG units they were 36 and 7 ft/d, respectively. Box plots of the calculated values (fig. 11) show much narrower ranges than values calculated from specific-capacity data, although the CRBG median value is similar to both the median of the specific-capacity derived value and previously reported values. For the basin-fill deposits, the diversity in lithology results in a larger difference in the median value between the methods. For example, some specific-capacity data are for coarse-grained materials such as cobbles, whereas others are from less productive sandstone; most of the aquifer test data, however, are for high-yield irrigation wells completed in coarse-grained material.

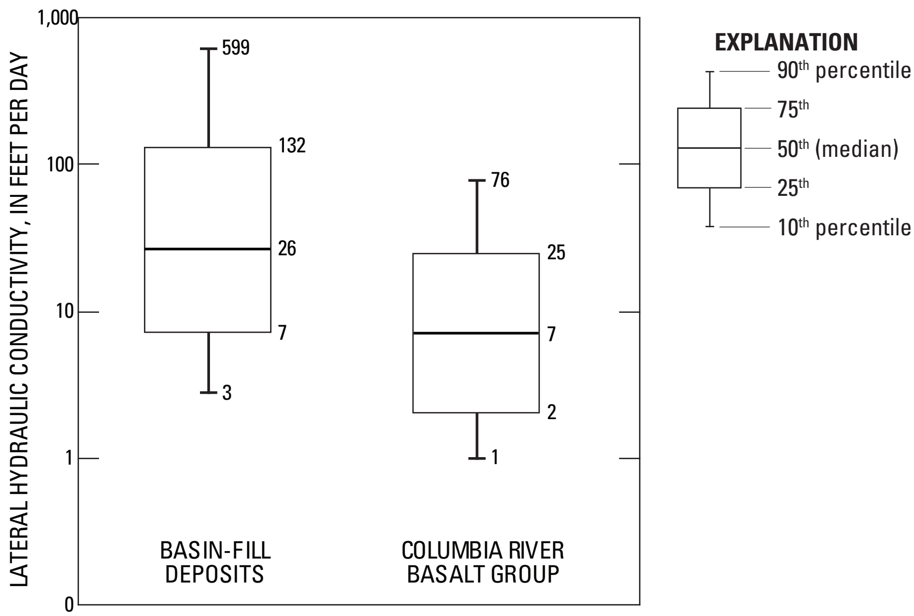


Figure 11. Distribution of lateral hydraulic conductivity estimated from aquifer test data, Yakima River basin aquifer system, Washington.

Vertical Hydraulic Conductivity

Vertical hydraulic conductivity (K_v) is an important hydraulic characteristic that is difficult to determine. It is a measure of a material's ability to transmit water vertically (the impedance to downward/upward flow) and is expressed in units of feet per day. K_v is a major control on the movement of water in the flow system—lateral and vertical variations in K_v affect vertical hydraulic gradients, and therefore, flow rates into, within, and out of units. Except for the work of Drost and others (1997), who used measured canal seepage rates to calculate K_v, nearly all of the previous estimates of K_v were derived using groundwater flow models.

Basin-Fill Units

The large differences in the types of basin-fill units results in a large variation in estimates of K_v. Fine-grained units, such as shale and clay, have values as small as 10^{-10} to 10^{-6} ft/d. K_v values between about 10^{-3} to 10^{-2} ft/d were estimated for units that contain fine-grained material such as loess and parts of the Ringold Formation. In contrast, more coarse-grained units had estimated values ranging from about 10^{-1} to 10 ft/d. Overall, K_v values ranged from 10^{-7} to 12 ft/d.

Bolke and Skrivan (1981) assigned a constant K_v of 9×10^{-3} ft/d in a groundwater model of the Toppenish alluvial aquifer. For the lower Satus Creek basin, Prych (1983) assumed a K_h:K_v anisotropy ratio of 1,000:1 for the old alluvium and upper Ellensburg Formation, yielding a K_v of 0.09 ft/d in a groundwater model. For the Touchet Beds in the lower Satus Creek basin, Prych's (1983) estimate of K_v ranged from 0.1 to 12 ft/d. On the basis of the previously described K_h values, effective regional K_v values for the basin-fill units are likely on the order of 0.1 to 1 ft/d.

For the sedimentary material overlying the Saddle Mountain unit in the Eastern Benton basin, Zimmerman (1983) assumed K_v to be controlled by the clays of the Ringold Formation, and used a constant value of 9×10^{-3} ft/d in a groundwater flow model.

Smoot and Ralston (1987) and Lum and others (1990) estimated a value of 0.05 ft/d for loess on the basis of a vertical anisotropy of 100:1. Hansen and others (1994) estimated K_v ranging from 4×10^{-7} to almost 1,400 ft/d for the sediments overlying the CRBG; the median value was 2 ft/d and the average vertical anisotropy was 25:1.

Drost and others (1997) used irrigation canal-seepage rates to estimate K_v in the Pasco basin. This method resulted in values of 0.4, 0.7, and 0.4 ft/d for the Touchet Beds, Pasco gravels, and upper Ringold Formation, respectively.

The K_v of the fine-grained (such as claystone, mudstone, and shale) parts of the consolidated units also is unknown but is likely as small as 10^{-10} to 10^{-6} ft/d assuming a K_h:K_v ratio of 1,000:1. Bredehoeft and others (1983) estimated K_v of the Pierre Shale and Cretaceous shale in South Dakota to be about 4×10^{-6} and 0.0005 ft/d, respectively. They also noted that values greater than 4×10^{-6} for the thick Cretaceous shale would not support an underlying flowing-artesian aquifer system due to increased vertical leakage. That is, the known upward vertical leakage through the shale has a threshold that is limited by the value of K_v. For the thick upper part of the late-Cretaceous Hell Creek Formation (shale dominated) in Montana, Hotchkiss and Levings (1986) used aquifer test data to estimate a maximum range in K_v from 0.00005 to 0.001 ft/d from aquifer tests. Model-derived K_v values for the shale-dominated confining units in a 42,000 mi^2 area of eastern Montana and northeastern Wyoming were about 0.00002 ft/d (Hotchkiss and Levings, 1986). Taken together, the summarized information indicates that the K_v of thick shale probably ranges from about 10^{-6} to 10^{-3} ft/d.

Columbia River Basalt Group Units

Zimmerman (1983) examined two possible controlling mechanisms of K_v for different basalt zones. In the “base case,” dense basalts were assumed to be the primary control on vertical flow. In the “alternate base case,” interbeds were the primary control. In both cases, dense basalts were assumed to control vertical flow in undisturbed zones. Zimmerman assigned vertical anisotropy ratios (K_h:K_v) of 1,000:1 to all groundwater model nodes in undisturbed zones, which yielded K_v values ranging from 3×10^{-5} to 8×10^{-5} ft/d. Fault and barrier zones for both cases were simulated as isotropic (K_h:K_v = 1), which may be similar to the anisotropy of a basalt flow interior. Differences between the two cases occurred along anticlines where interbeds rather than dense basalts were assumed to control vertical flow. Davies-Smith and others (1988) also estimated that the Mabton unit, in contrast to basalt, exerted more control on the vertical movement of water in the Umatilla basin in Oregon, where the Mabton is composed of fine-grained deposits. Vertical anisotropy for Saddle Mountains and Wanapum were assigned values of 8,333:1 and 2,063:1, respectively, yielding values for K_v ranging from 4×10^{-7} to 1.5×10^{-4} ft/d (Davies-Smith and others, 1988).

Vertical anisotropy for the Wanapum was estimated to be 500:1, with K_v ranging from 8.0×10^{-4} to 1.2×10^{-3} ft/d in the Pullman-Moscow basin in Washington and Idaho (Smoot and Ralston, 1987; Lum and others, 1990). These investigations estimated the Grande Ronde anisotropy to range from 2,000:1 to 5,000:1, with K_v ranging from 1×10^{-4} to 2.5×10^{-3} ft/d.

Whiteman and others (1994) reported that K_v of the Columbia Plateau aquifer system was largely unknown, but estimated that values ranged from 5×10^{-4} to 4 ft/d, with vertical anisotropy of 1,000:1 to 100:1. Hansen and others (1994) estimated that K_v in the system ranged from 5×10^{-5} to 7 ft/d, with a median value of 1×10^{-3} ft/d. Typical vertical anisotropy was 1,500:1 to 1,000:1.

Packard and others (1996) estimated K_v for two zones (anticlines and synclines) in a groundwater flow model of the Horse Heaven Hills on the southeastern border of the basin. A K_v value of 5×10^{-4} ft/d was calibrated for the anticline zone for the Saddle Mountains and Wanapum, whereas along the synclines the average value was larger (0.01 ft/d). The anisotropy ratio for the Wanapum was similar to that of Lum and others (1990).

Drost and others (1997) estimated K_v to be 0.3 ft/d for the Saddle Mountains on the basis of canal seepage losses. Although this K_v is larger than most other reported values, it was based on water-level changes with time resulting from a known amount of seepage (recharge).

Bedrock Units

The authors have been unable to locate any previous estimates of K_v for the bedrock units in the study area. K_v values for these units likely range over several orders of magnitude and vary by the type of materials comprising a unit, for example, schist in contrast to sandstone. For the Paleozoic, Mesozoic, and Tertiary units as a whole, the average vertical anisotropy would be large, perhaps on the order of 2,000:1 to 10,000:1. On the basis of information described above, K_v for the shale/clay parts of the sedimentary units may range from 4×10^{-6} to 0.0005 ft/d.

Storage Coefficient

The storage coefficient is a measure of a unit’s ability to store and release water, and is defined as the volume of water that a unit will absorb or release from storage per unit surface area per unit change in head. It is expressed in units of cubic feet per cubic feet, a dimensionless quantity. Storage coefficients for a confined aquifer can range from 5×10^{-5} to 0.005 and values for an unconfined aquifer (referred to as specific yield) are much larger, and can range from 0.01 to 0.30 (Freeze and Cherry, 1979).

Analyses of aquifer tests in the Pasco basin yielded unconfined storage coefficient values of 0.1 for the middle Ringold Formation and 0.15 to 0.2 for Pasco gravels (Drost and others, 1997). Aquifer tests in the confined parts of the Pasco basin yielded storage coefficients of 0.03 to 0.07 for Pasco gravels, 7×10^{-5} to 0.06 for the middle Ringold Formation, 0.002 to 0.05 for the lower Ringold Formation, and 1×10^{-6} to 0.006 for the CRBG (Drost and others, 1997). Drost and others, (1997) also reported that previous modeling studies calibrated CRBG storage coefficient values of 0.0001 to 0.01. Vermeul and others (2001) estimated storage coefficients for the coarse-grained sediments of the unconfined parts of the Hanford and Ringold Formations as 0.07 and 0.2, respectively. Model-derived storage coefficients for all confined units were 1×10^{-6}.

On the basis of grain size, Drost and others (1997) estimated storage coefficients for the Touchet Beds (0.08), upper Ringold Formation (0.07 to 0.2), and lower Ringold Formation (0.02 to 0.2). Prych (1983) used a specific yield of 0.1 for the Touchet Beds in the lower Satus Creek basin. Specific yield for the Toppenish alluvial aquifer has been estimated to be 0.2 (U.S. Geological Survey, 1975).

Golder Associates (2002) estimated a storage coefficient of 2×10^{-3} for the lower part of the upper Ellensburg Formation on the basis of buildup and drawdown data during aquifer tests. Converse Consultants NW (1991) used data from a 24-hour aquifer test to estimate a storage coefficient of about 7×10^{-4} for the lower part of the upper Ellensburg Formation.

Whiteman and others (1994) indirectly estimated storage coefficients for the CRBG units on the basis of specific storage. Median values for the Saddle Mountains, Wanapum, and Grande Ronde were 2×10^{-5}, 3×10^{-5}, and 2×10^{-4}, respectively. For the sediments overlying the CRBG, they estimated values ranging from 2×10^{-4} to 0.2.

Previously-estimated storage coefficients were initially used by Hansen and others (1994, table 4) in the Columbia Plateau aquifer system model. Model-derived specific yields for the overburden aquifer ranged from 0.1 to 0.2. The estimated median storage coefficients for the Saddle Mountains, Wanapum, and Grande Ronde were 4×10^{-5}, 4×10^{-5}, and 2×10^{-4}, respectively. Packard and others (1996) initially assigned CRBG storage coefficients on the basis of previous studies in a groundwater model of the Horse Heaven Hills. Transient calibration of the model yielded values of 1×10^{-3} for the Grande Ronde and Wanapum and 1×10^{-2} for the Saddle Mountains.

Values of unconfined and confined storage coefficients for the sandstone layers of the bedrock units are unknown. Leake and others (2005) estimated values of 10^{-4} and 0.06 (confined and unconfined values, respectively) for the Permian Coconino Sandstone of the Coconino Plateau, Arizona. Wells completed in this sandstone generally yield about 1 gal/min, which is similar to many reported well yields from the sandstone in the bedrock units in the study area. Values for the basalt part of the bedrock units are likely smaller than those of the younger CRBG due to fracture infilling and may range from about 10^{-6} to 10^{-4}. Values also are unknown for the Quaternary bedrock unit but should be similar to those reported for the Quaternary basalts of the Snake River Plain aquifer that has reported specific yield values of 0.01 to 0.20 and storage coefficient values of 10^{-5} to 10^{-3} (Lindolm, 1996).

Hydrochemistry

Patterns in the variations in the stable isotopes of hydrogen and oxygen, carbon isotopes, and concentrations of dissolved constituents, including methane and noble gases, can be used to infer information on the hydrogeologic framework and groundwater movement, including vertical movement of groundwater in both the CRBG units and the overlying basin-fill units. After recharge of isotopically distinct water, the extent of water-rock interaction increases with increased contact time and temperature, such that the concentrations of many dissolved constituents can either increase or decrease as groundwater flows from areas of groundwater recharge to areas of groundwater discharge. Dissolved chloride, fluoride, and helium are essentially non-reactive solutes in the dilute basaltic groundwater of the aquifer system. Thus, these constituents, along with sodium (because of its limited reactive character), tend to accumulate in groundwater along flow paths and are indicators of the extent of water-rock interaction. The concentration of other solutes, such as dissolved oxygen, carbon-14, and magnesium tend to decrease along flow paths. The concentration of dissolved inorganic carbon, which is a reactive component in groundwater in the aquifer system, also can be analyzed in conjunction with variations in carbon-13 to provide additional information on the conceptual model of groundwater in the aquifer system.

Hydrochemical data available for analysis include data collected by the USGS during this study, unpublished data from the Toppenish basin previously collected by the Yakama Nation (appendix B), USGS data reported by Bortleson and Cox (1986), Steinkampf and Hearn (1996), and the National Water-Quality Assessment Program, and stored in NWIS, and data for the lower Pasco basin (in particular the Hanford Site) obtained by the U.S. Department of Energy (Early and others, 1986; Reidel and others, 2002). All USGS data were collected using standardized USGS techniques and the water samples were analyzed by the methods of analysis of the USGS (for example, Fishman and Friedman, 1989, Brenton and Arnett, 1993, and U.S. Geological Survey, 2006). Data reported by others also were collected and analyzed using published protocols and methods of analysis. Some of the hydrochemical data for the deeper basalt wells outside the Hanford Site were collected from large-capacity production wells with multiple contributing zones that provide suboptimum geochemical sampling conditions. However, the work of Hearn and others (1985), McKinley (1990), and Steinkampf and Hearn (1996) shows the value of those data in assessing the hydrogeologic framework, geochemical evolution of groundwater, and groundwater movement.

Variation in Major-Ion Concentration

Concentrations of solutes in groundwater change with the residence time of the water and its flow path. Groundwater typically evolves from an oxygenated calcium-magnesium-bicarbonate type in recharge areas to an anoxic sodium-potassium-bicarbonate type at deeper, downgradient locations. Long flow paths extending deeply into the central Pasco basin lead to anoxic groundwater of a sodium-chloride type (Reidel and others, 2002), the penultimate end-member water type along the deepest flow paths.

The carbon present in the aquifer system is primarily in the bicarbonate form. Bicarbonate concentrations in recharge are a function of carbonic acid solution of carbon dioxide generated by soil respiration. Other dissolved solutes in basaltic groundwater are derived principally from the dissolution of basaltic glass and to a lesser extent plagioclase feldspar (Hearn and others, 1985). Chloride, sodium, and fluoride are essentially non-reactive solutes in basaltic groundwater of the Yakima River basin and tend to accumulate and provide information on the flow system. For example, information on sodium and chloride concentrations can be used to make rough estimates of residence times based on dissolution rates (Steinkampf and Hearn, 1996). Sodium and chloride concentrations greater than about 20 and 5 mg/L, respectively, tend to be associated with higher concentrations of accumulated fluoride, indicating residence times greater than 5,000 years (Steinkampf and Hearn, 1996). In turn, sodium concentrations in the CRBG units are significantly related to pH and further indicate evolution of the hydrochemistry along flow paths. A more detailed description of the chemical evolution of groundwater in the CRBG units can be found in Hearn and others (1985), Steinkampf and Hearn (1996), and Reidel and other (2002).

Deuterium and Oxygen-18

The isotope ratios of hydrogen ($^{2}H/^{1}H$) and oxygen ($^{18}O/^{16}O$) are expressed in delta notation as δD and $\delta^{18}O$ and are reported in units of per mil (or parts per thousand, ‰). The δD and $\delta^{18}O$ in groundwater and surface water in the basin plot near the Global Meteoric Water Line (GMWL) of Craig (1961) (fig. 12*A*), as well as Local Meteoric Water Line for Washington State (Coplen and Kendall, 2000; Kendall and Coplen, 2001). The slope of the regression line fitted to all the groundwater and surface-water data is 7.8, which closely approximates the 8.0-slope of the GMWL. These data indicate that the source of groundwater in the Yakima River basin is meteoric water derived from atmospheric precipitation that is largely unaffected by evaporation. This finding is consistent with that of Hendry and others (1992). Small differences were observed in the slope of meteoric water lines for different hydrogeologic units and components of the hydrologic system, indicating local variability within the large regional system. The isotopic composition of precipitation is quite variable compared to groundwater and surface water (fig. 12*B*; Hendry and others, 1992). The larger variations in the isotopic composition of precipitation are related to differences in altitude and the seasonality of precipitation; these types of variations are common for precipitation (Gat, 1980; Clark and Fritz, 1997, Coplen and others, 2008). The slope of the precipitation meteoric water line is close to 7 (regression line and equation shown in fig. 12*B*), which is similar to the precipitation slope from the Hanford Site (Early and others, 1986).

Isotopic data for the alluvial aquifers (unconsolidated units) cluster in a narrow range, and are similar to the surface-water data (figs. 12*C*, *D*). Areas where surface-water recharge is a substantial fraction of groundwater recharge are evidenced by a shift in the δD and $\delta^{18}O$ values. For example, infiltration of surface water from irrigation canal leakage and the application of surface water to croplands accounts for large quantities of local groundwater recharge (Vaccaro and Olsen, 2007a), and thus, the similarity in isotopic composition. The distribution of isotopic compositions of groundwater from alluvial aquifers overlaps those for the Yakima River samples, whereas isotopic compositions of groundwater from the consolidated (composed of the Ellensburg Formation) and CRBG units are typically more depleted in the heavier isotopes (more negative delta values) and more variable than the isotopic compositions for both the Yakima and Naches Rivers (figs. 12*C*, *D*). The $\delta^{18}O$ of samples from the Yakima and Naches Rivers cluster between -12 to -14.5 ‰ and the δD ranges between -87 to -108 ‰. The groundwater samples with more negative delta values have a larger range, with $\delta^{18}O$ ranging from -14 to -19.5 ‰ and δD ranging from -105 to -145 ‰. Thus, substantial differences are apparent in the individual isotopic compositions of surface waters and their ranges compared to those of groundwater from most consolidated and CRBG units (figs. 12*C*, *D*). The δD and $\delta^{18}O$ values (-135 and -16.2 ‰, respectively) reported for a deep (1,163 ft) Ellensburg Formation well in the Ahtanum subbasin (Golder Associates, 2001) are in this range and are similar to values for other deep wells (appendix B). The larger variation in the isotopic compositions of groundwater from consolidated and CRBG units as compared to surface water is indicative of a more extensive groundwater recharge area in the uplands but in some instances also may be related to cooler paleoclimatic conditions of the most recent glacial period (Larson and others, 2000). The difference in the isotopic compositions of water from the consolidated and CRBG units as compared to the alluvial (unconsolidated) units is clearly shown by the regression lines and equations of figure 12*D*.

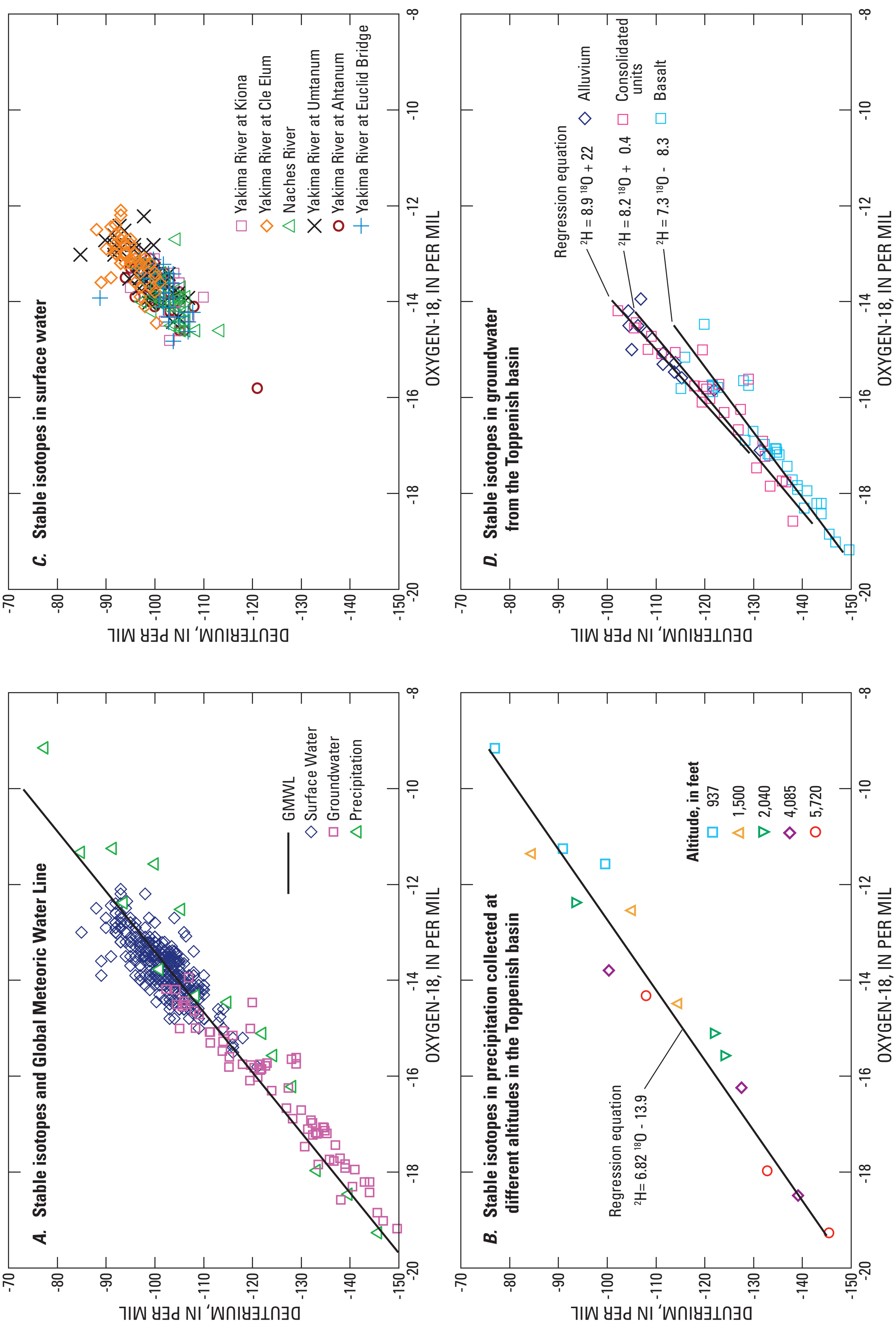


Figure 12. Relation between stable isotopic composition (deuterium and oxygen-18) samples in (*A*) all water samples, (*B*) precipitation in the Toppenish basin, (*C*) surface water in the Yakima River basin, and (*D*) groundwater in the Toppenish basin, Washington.

Methane

Dissolved methane has previously been reported in groundwater from semi-confined to confined aquifers in the CRBG (Hammer, 1934; Walsh and Lingley, 1991; Johnson and others, 1993; and Reidel and others, 2002). The concentration of dissolved methane measured in 80 percent of samples from consolidated and CRBG unit wells (appendix B) were larger than could be attributable to atmospheric sources; current atmospheric concentration of methane is about 1.2 mg/L. Concentrations as large as 50 mg/L were measured in samples analyzed by gas chromatography and actual concentrations may be larger because the gas-chromatography analysis of dissolved concentrations of nitrogen showed substantial alteration from atmospheric concentrations—indicting partial stripping of dissolved gases during sampling.

The stable carbon isotopic compositions of carbon ($\delta^{13}C_{methane}$; $^{13}C/^{12}C$) of dissolved methane in groundwater varied from -9.9 to -53.0 ‰, with all but one value ranging between -32.7 to -53.0 ‰ (fig. 13). These values are typical of methane produced by thermocatalytic breakdown of higher mass hydrocarbons by heat ($\delta^{13}C_{methane}$ ranging from -25 to -50 ‰) as opposed to biogenic production ($\delta^{13}C_{methane}$ ranging from -50 to -80 ‰) of methane from microbial reduction of buried organic matter (Clark and Fritz, 1997). A biogenic source of dissolved methane could be produced within the consolidated and CRBG units by microbial methanogenesis, whereas thermocatalytic sources would indicate methane migration from outside the consolidated-CRBG part of the aquifer system. Microbial methanogenesis and oxidation would substantially influence the $\delta^{13}C_{DIC}$ isotopic ratio of dissolved inorganic carbon (DIC), which is also used to age date groundwater. According to Johnson and others (1993), the isotopic composition of methane present in groundwater on the Hanford Site, located at the terminus of regional groundwater flow paths, indicates that the source of the methane was a mixture of biogenic and thermogenic processes, and its occurrence in groundwater in the deeper CRBG units was largely due to entrainment by upwelling groundwater from coal beds in Tertiary sedimentary rocks underlying the CRBG. Biogenic methane also has been reported in groundwater from south-central Washington with $\delta^{13}C_{methane}$ values of -63.6 to -88.2 ‰ (Rice, 1993). These values are consistent with the fact that highly-reducing geochemical conditions necessary for methanogenesis are present in the groundwater system in the Yakima River basin (appendix B) and carbon, in the form of peat, is present in some areas of the Grande Ronde unit (Steinkampf and Hearn, 1996).

The observed concentrations of methane are not uniformly distributed (fig. 13). For example, large concentrations of methane suitable for production of gas, such as in the Rattlesnake Hills Gas Field described by Hammer (1934), do not exist elsewhere in the basin. Several hydrocarbon exploration wells drilled through the Grande Ronde unit in eastern Washington, however, have been reported to have concentrations of methane approaching quantities necessary for commercial development. Johnson and others (1993) concluded that the areal distribution of methane in the CRBG on the Hanford Site is controlled by preferential flow near vertical faults. For example, major structural features such as the Cold Creek Fault have been cited as the source of the uneven vertical and lateral distribution of methane near a Hanford Site well (Reidel and others, 2002). However, most of the occurrences of methane measured in groundwater samples from wells outside of the Hanford Site were found at locations several miles distant from mapped structural fault features, indicating upward vertical movement of predominantly thermogenic methane from the underlying bedrock may be more widespread than previously thought and (or) the more general occurrence of unmapped fault structures. In either case, upward migration of methane through the CRBG from the underlying bedrock units is indicated by the data.

Carbon and Tritium Isotopes

Tritium (^{3}H) and carbon ($^{13}C_{DIC}$ and $^{14}C_{DIC}$) isotopic data can be used to infer groundwater residence time (age), or the length of time between recharge and collection of a groundwater sample. Large amounts of tritium and ^{14}C were injected into the atmosphere during or after the period of above-ground testing of thermonuclear weapons from 1953 to 1963.

Tritium is a radioactive isotope of hydrogen that has a half-life of about 12.3 years. It is present at elevated concentrations in groundwater that was recharged during or after the weapons-testing period. Groundwater with tritium concentrations greater than 0.5 tritium units (TU) are categorized as modern, and waters with concentrations less than 0.5 TU are categorized as pre-modern (recharged prior to 1952).

Carbon-14 is a radioactive isotope of carbon that is continually produced within the upper atmosphere by natural processes. It is incorporated into groundwater as dissolved inorganic carbon (DIC) when recharge water is in contact with the atmospheric gases. This process occurs mostly in the unsaturated zone where gaseous concentrations of carbon dioxide (CO_2) are substantially increased by microbial activity and root respiration. The half-life of ^{14}C is 5,730 years, and can be used to estimate groundwater ages as much as 30,000–35,000 years. Measurements of ^{14}C are compared to the concentration present in 1950 (before widespread thermonuclear weapons testing increased atmospheric concentrations of $^{14}CO_2$), which is referred to as 100 percent modern carbon (PMC). For example, groundwater with a ^{14}C concentration of inorganic carbon of 50 PMC is interpreted to have an estimated groundwater residence age of 5,730 years

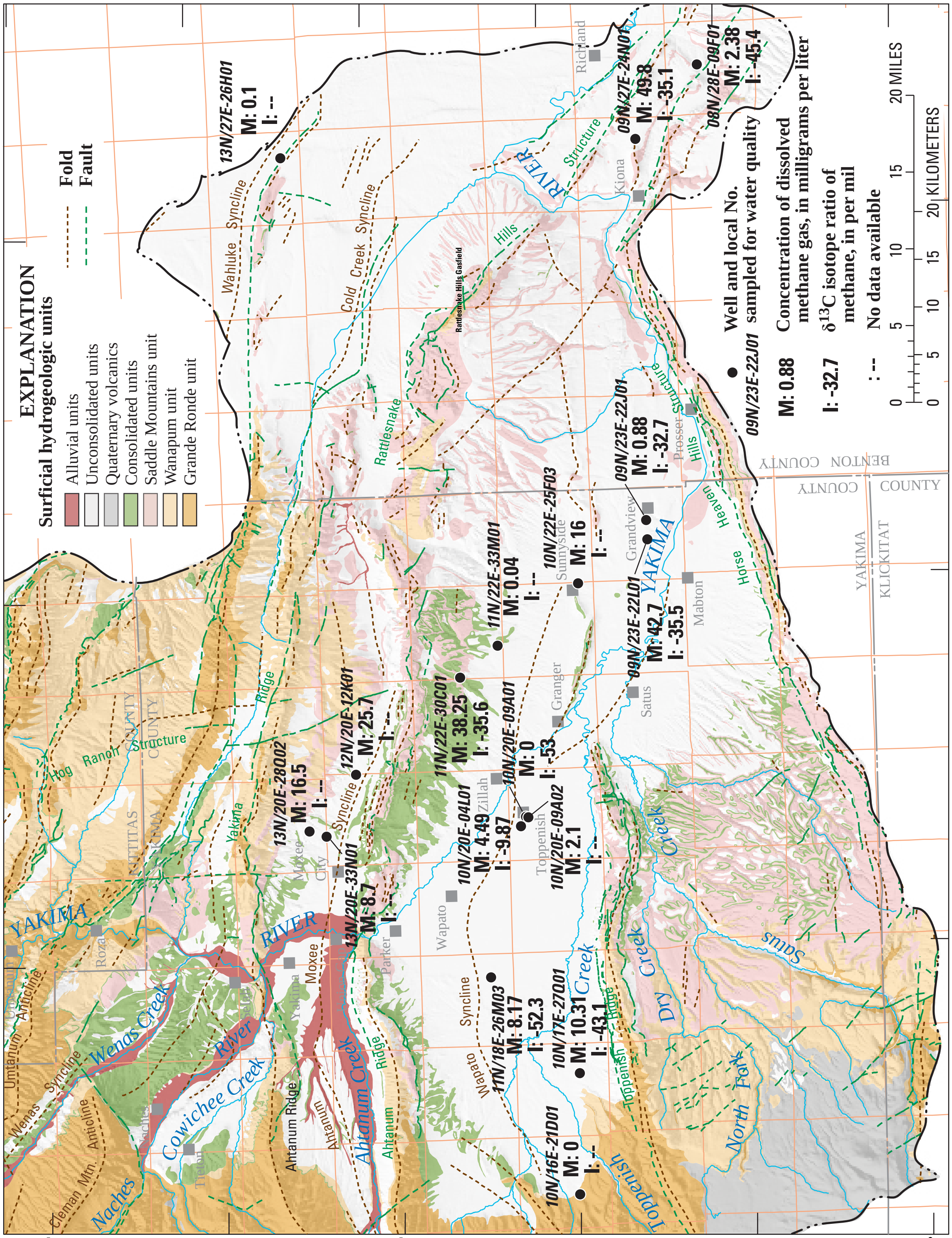


Figure 13. Spatial distribution of dissolved methane concentrations and carbon-13 ratios of dissolved methane in groundwater, Yakima River basin aquifer system, Washington.

if there were no indications of any geochemical reactions altering the isotopic concentration of DIC. However, within deep anaerobic aquifers, such as the more buried CRBG units, there can be many factors that influence the concentration of DIC, and thus affect the concentration of ^{14}C, which in turn will substantially influence estimates of groundwater age. The $\delta^{13}C_{DIC}$ concentrations are useful in interpreting geochemical alterations of $^{14}C_{DIC}$ and the hydrochemical history of groundwater, facilitating better estimates of groundwater age. Some samples that contain mixtures of old and young groundwater can be identified if both the concentration of tritium is greater than 0.5 TU and the $^{14}C_{DIC}$ concentration is substantially less than 100 PMC or less than the values measured in recently recharged (young) groundwater.

Large variations in ^{14}C and tritium concentrations are apparent (fig. 14*A* and appendix B). The large range in $\delta^{13}C_{DIC}$ (-17 to 2.8 ‰) is indicative of extensive interaction of dissolved carbon in the groundwater system and, hence, significantly complicates the application of $^{14}C_{DIC}$ concentrations in estimating groundwater residence time. The $\delta^{13}C_{DIC}$ values are more positive (enriched) in groundwater with larger concentrations of chloride and fluoride (fig. 14*B*, appendix B), demonstrating the relation between geochemical processes that accumulate conservative solutes as groundwater evolves along a flow path. The effect of these processes typically is to dilute the $^{14}C_{DIC}$ concentration, resulting in larger apparent groundwater ages.

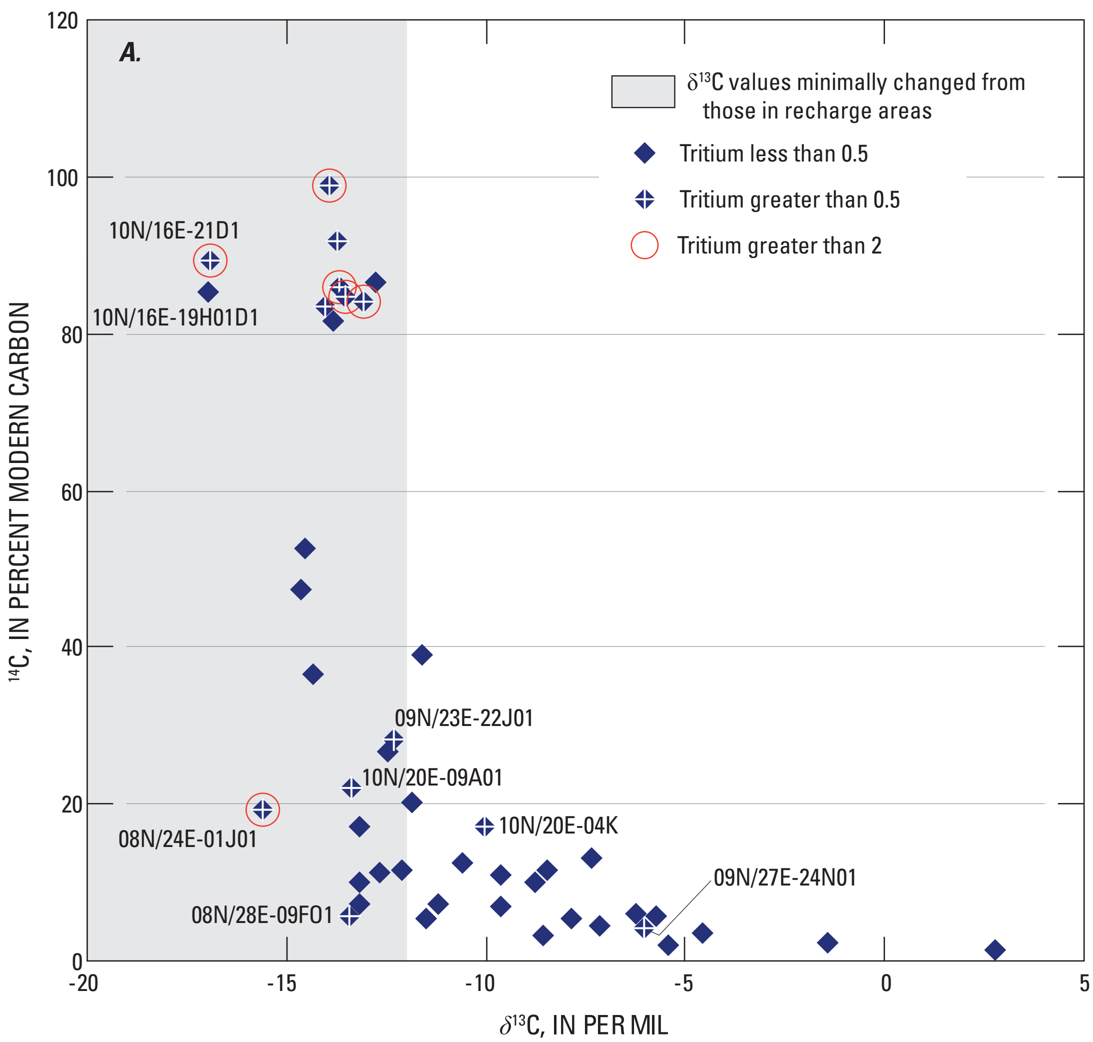


Figure 14. Relation between (*A*) tritium and $\delta^{13}C$ and carbon-14 (^{14}C) concentrations in dissolved inorganic carbon of groundwater samples from the Toppenish and Yakima basins, Washington and (*B*) $\delta^{13}C$ and fluoride and chloride concentrations.

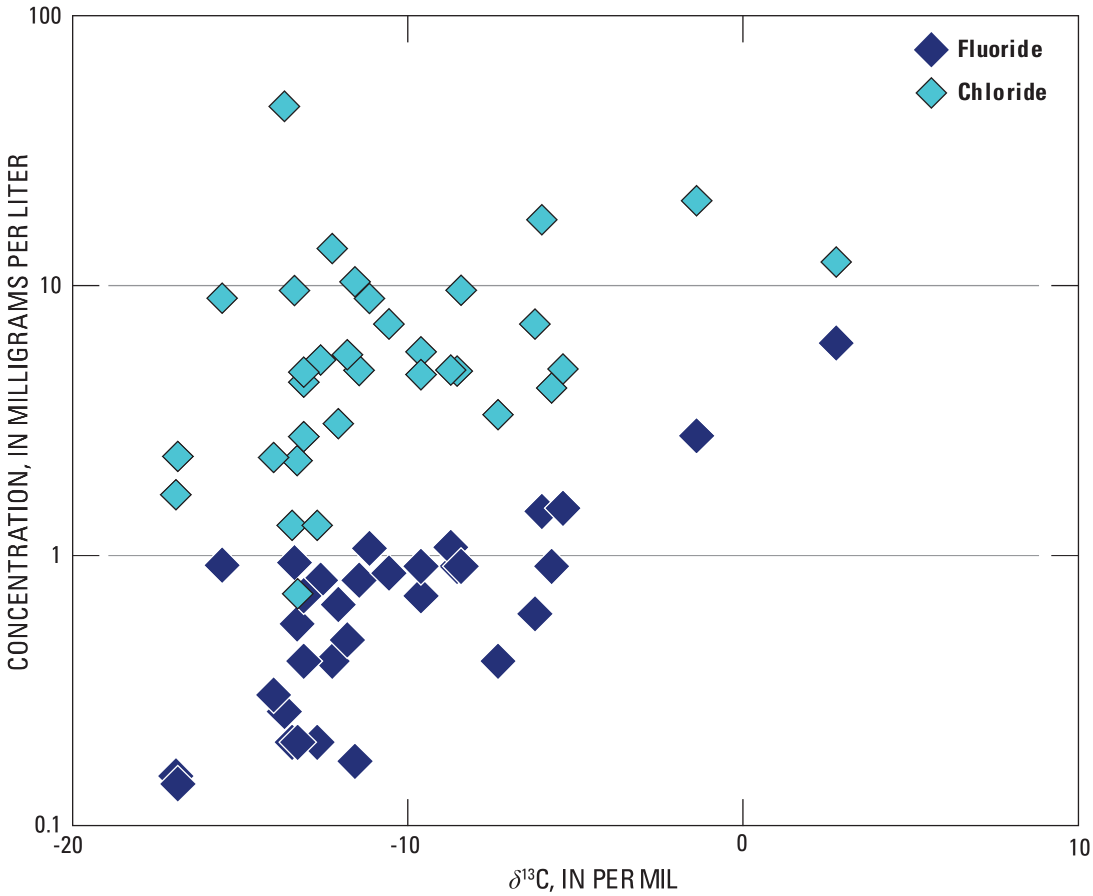


Figure 14.—Continued.

Based on tritium concentrations greater than 0.5 TU, the $\delta^{13}C_{DIC}$ and $^{14}C_{DIC}$ of recently recharged groundwater generally plot as a cluster (fig. 14*A*), with $^{14}C_{DIC}$ values typically being larger than 80 PMC and most $\delta^{13}C_{DIC}$ values ranging from -17 to -13 ‰. These groundwater samples are from wells that are relatively shallow (depths ranging from 150 to 300 ft) and are typically located in upgradient areas with respect to groundwater flow. Samples with the lower $\delta^{13}C_{DIC}$ values (about -16.9 ‰) were from two wells (10N/16E-19H01D1 and 10N/16E-21D1) that were sampled to provide information on the composition of relatively recent groundwater from the upgradient areas in the western uplands of the Toppenish basin near Ft. Simcoe. There was a substantial tritium concentration in the 158-ft deep well tapping the consolidated unit (21D1); however, the tritium concentration in the 425-ft deep well (19H01D1) tapping the underlying basalt unit was less than 0.5 TU.

The $\delta^{13}C_{DIC}$ and $^{14}C_{DIC}$ values vary in a generally consistent manner (fig. 14*A*) suggesting some basic patterns of change. Waters with higher $^{14}C_{DIC}$ values that also have a high tritium concentration likely include ^{14}C from weapons testing, demonstrating that the $^{14}C_{DIC}$ concentration in recently recharged groundwater unaffected by bomb-generated carbon is likely in the range of 80 to 85 PMC. Samples from greater depths below land surface and at downgradient locations typically had more enriched $\delta^{13}C_{DIC}$ values and diminished $^{14}C_{DIC}$ values—indicating that groundwater has undergone substantive geochemical alteration and in some instances has been within the groundwater system for many thousands of years. Samples from the deepest wells typically displayed the greatest differences in values compared to samples collected from wells near recharge areas.

The influence of geochemical reaction on $\delta^{13}C_{DIC}$ values is well demonstrated by the plot of $\delta^{13}C_{DIC}$ against the concentrations of chloride and fluoride (fig. 14*B*); both chloride and fluoride accumulate along flow paths in the CRBG (Steinkampf and Hearn, 1996). As fluoride accumulates in groundwater, there is a consistent increase in $\delta^{13}C_{DIC}$. The mechanism for the $\delta^{13}C_{DIC}$ increase is not well known. Microbial production of methane (methanogenesis) would result in a similar increase of $\delta^{13}C_{DIC}$, but the $\delta^{13}C$ of dissolved methane measured in groundwater samples typically indicates a non-biogenic source of methane. The mixed source of methane, combined with evidence of gas stripping, preclude accurate estimation of the extent of the effect that methanogenesis would have on the $^{14}C_{DIC}$ and thus, estimates of groundwater age. Alternatively, dissolution of calcite with higher $^{13}C_{DIC}$ values could result in an overall increase in $^{13}C_{DIC}$ of groundwater.

Groundwater represented by the data plotted in the shaded area on the left of figure 14*A* show minimally changed $\delta^{13}C_{DIC}$ values from those in the recharge area having a concurrent decrease in $^{14}C_{DIC}$ from about 80 PMC to somewhere in the range of about 10 to 20 PMC. If the change in $^{14}C_{DIC}$ is due to solely radioactive decay, it would represent about 2 to 3 half-lifes, which is equivalent to 11,500–17,000 years. Assuming groundwater flow path lengths from recharge area to sampled well locations are on the order of 15 to 30 mi, groundwater flow velocities would be in the range of 0.01 to 0.04 ft/d. These estimates would include a component of vertical flow because some of the sampled wells were in upward flow areas. Previous estimates of groundwater flow velocities in the deeper basalt units on the basis of numerical simulation were on the order of 0.1 to 0.2 ft/d (Vaccaro, 1999) and 0.001 ft/d on the basis of $^{14}C_{DIC}$ concentrations (Reidel and others, 2002).

Noble Gases

Helium and other noble gases (neon, argon, krypton, and xenon) are present in groundwater within the aquifer system, and their concentrations can provide information on the source or sources of the gases. Additionally, the isotopic concentrations of helium also provide information on its source or sources. To estimate if the source for any of the gases was upward vertical migration from the Earth’s mantle, water samples from nine wells were analyzed by isotopic mass-spectrometry (Stute and Schlosser, 2000).

Concentrations of helium, neon, argon, krypton, and xenon in groundwater in the samples are shown in figure 15 along with the calculated equilibrium air-saturation values for recharge occurring at 10 and 20°C (Mazor, 1991). With the exception of helium, the concentrations of the gases are

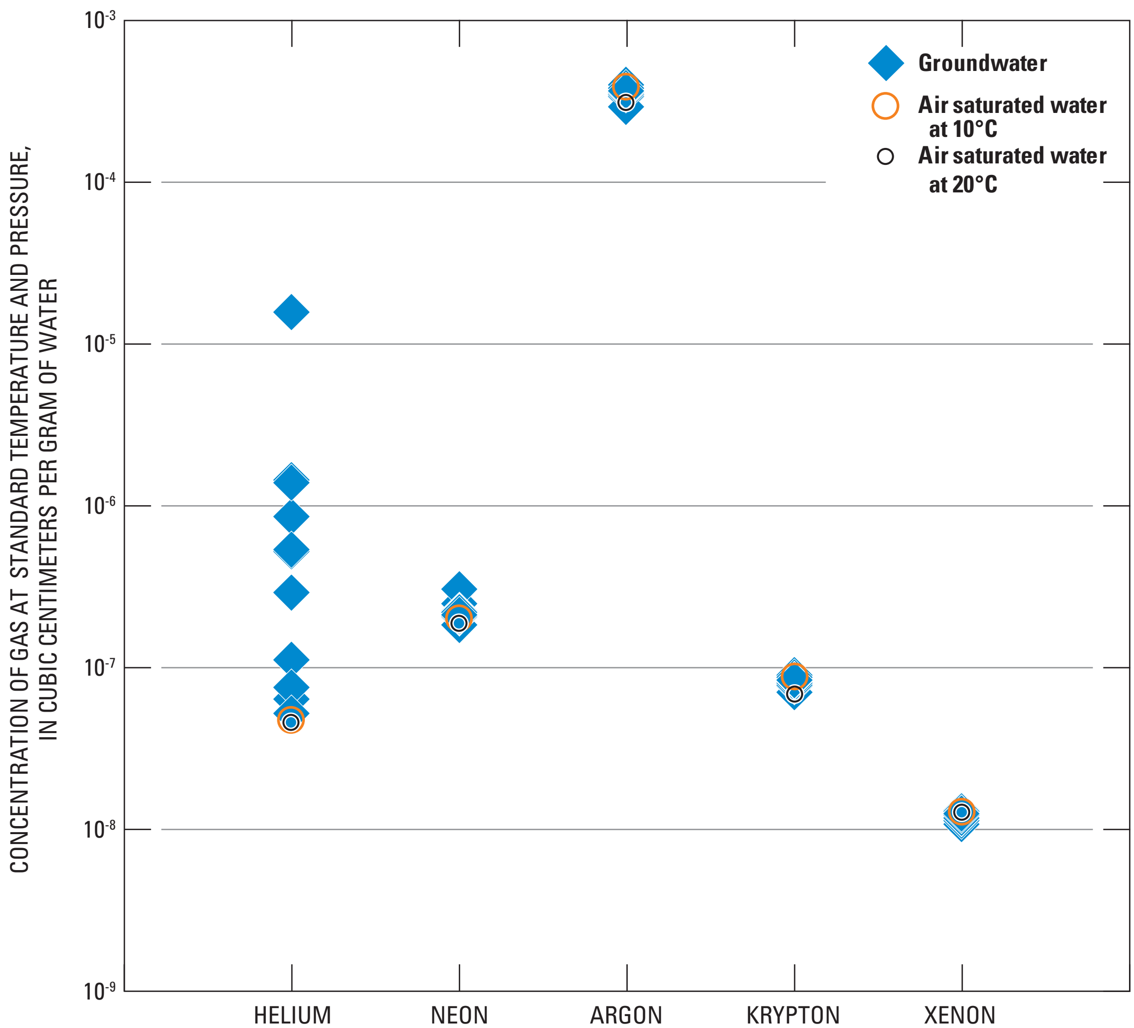


Figure 15. Concentrations of noble gases in groundwater samples from Yakima River basin aquifer system, Washington, and calculated concentrations of noble gases in air-saturated water.

within the expected solubility range for dilute natural waters saturated with atmospheric gases during the recharge process in the unsaturated zone. These measurements indicate that the atmosphere is the primary source of these noble gases detected in the groundwater of the basin.

Concentrations of helium (He), and to a lesser degree, neon, exceed air-saturated concentrations levels. Other than ‘excess air,’ which are air bubbles that can become trapped at the water table by rapidly changing water levels and create over-saturated concentrations (Heaton and Vogel, 1981), other sources of neon in groundwater are not expected to be significant. Additional sources of helium commonly found in groundwater include that produced by radioactive decay and externally produced helium transported into the aquifer. Sources of helium can be partially distinguished on the basis of the $^3He/^4He$ ratio in the sample normalized to the $^3He/^4He$ ratio in air as $R/Ra = (^3He/^4He)_{sample} / (^3He/^4He)_{air}$. Helium produced from radioactive decay of tritium (tritigenic helium, ^3He) has a R/Ra greater than 1, radiogenic helium generated from the decay of naturally occurring uranium and thorium in the rock matrix has a R/Ra less than 1, and helium transported from deeper mantle sources has a R/Ra greater than 1 (Lupton, 1983; Solomon, 2000). For water with a R/Ra greater than 1, one cannot distinguish the source of helium without other types of data. For R/Ra less than 1, one cannot explicitly determine if part of the helium is from a mantle source, only that the predominant source is radiogenic helium. But mantle helium is so enriched with ^3He that the likely source is mantle derived.

The normalized helium ratio in the samples ranged from 0.01 to 0.43, indicating that the predominant source of excess (above atmospheric concentrations) helium in the groundwater samples was radiogenic decay of uranium and thorium. Measureable increases in helium concentrations due to radiogenic production require on the order of thousands of years to accumulate (Solomon, 2000). Reidel and others (2002) suggest a helium accumulation rate for pore water in the CRBG of 1.7×10^{-11} cubic centimeter of helium (at standard temperature and pressure) per year per gram of water (cc He (STP)/yr per gram of water). After accounting for atmospheric concentrations and assuming the above accumulation rate, the helium data indicate that excess helium concentrations greater than 1×10^{-7} cc He (STP)/g water would be needed to yield estimates of groundwater ages on the order of thousands of years and more. It is of interest that the three groundwater samples containing less than 10^{-7} cc He (STP)/g water and having estimated ages in hundreds of years also contained tritium at concentrations 5–38 TU, which is indicative of modern groundwater. As with the carbon isotope data, this further shows the need for using multiple methods to improve our understanding of groundwater ages and the vertical connections in the flow system.

Groundwater in the Yakima River Basin

Water derived from rainfall and snowmelt recharges the aquifer system in the Yakima River basin and moves through the system from upland areas to the lowlands. Recharge derived from irrigation in the lowlands moves to farm drains and streams in the structural basins. The movement of the water is influenced by topography, geologic structure, and hydraulic characteristics of the sediments and rocks. This groundwater is present under unconfined (water table) to confined conditions. Changes in recharge and discharge conditions, and the withdrawal of groundwater for multiple uses affects water levels, streamflow, and water availability for both short and long periods.

Occurrence

Groundwater occurring in the hydrogeologic units of the aquifer system originated as recharge that was derived from rainfall, snowmelt, and, starting in the 1880s, irrigation. As part of this study, Vaccaro and Olsen (2007a) estimated daily recharge to the Yakima River basin aquifer system using the USGS Precipitation-Runoff Modeling System (Leavesley and others, 1983) and the Deep Percolation Model (Vaccaro, 2007) that are contained in the USGS Modular Modeling System (Leavesley and others, 1996). Recharge was defined as water leaving the active root zone or, for barren soils, the bottom of the mapped soil column, and is considered a ‘potential’ amount of recharge because a large part of the estimated recharge in the uplands does not enter the regional groundwater flow system but is discharged locally along short flow paths to streams. Recharge was estimated for water years 1950–2003 on the basis of a coverage of composite historical land use and land cover (herein called current conditions); recharge was also estimated assuming that there had been no human activities during that period (assumed predevelopment conditions). In addition, the monthly values of current-condition recharge that were aggregated from the estimated daily values by Vaccaro and Olsen (2007a) were refined and made available by Vaccaro and Olsen (2007b) in a spatially consistent and accessible format for water years 1960–2001—the period of interest in constructing a groundwater flow model. The monthly values were locally refined to account for decreased surface-water deliveries and application rates during the prorating years in the major irrigation districts with proratable rights (Vaccaro and Olsen, 2007b).

Potential mean annual recharge for current conditions for water years 1950–2003 was estimated to be 15.6 in. or 7,149 ft^3/s (about 5.2 million acre-ft), which is about 58 percent of the mean annual precipitation (Vaccaro and Olsen, 2007a). When predevelopment conditions were assumed, estimated recharge was 11.9 in. or 5,450 ft^3/s (about 3.9 million acre-ft) (Vaccaro and Olsen, 2007a).

Predevelopment recharge was about 97 percent of the estimated mean annual unregulated streamflow, indicating that only about 3 percent of the unregulated streamflow was supported by surface runoff, which is consistent with the results of Mastin and Vaccaro (2002).

Estimates of recharge were extended beyond the eastern boundaries of the Yakima River basin to the Columbia River to include an additional 808 mi^2. This area includes 700 mi^2 of the extended study area and another 108 mi^2 within the basin that was not included in the area analyzed by Vaccaro and Olsen (2007a); thus, the total recharge-estimation area is coincident to the extended study area. The methods used to estimate recharge for this 808 mi^2 area is described in appendix C. In this area, the mean annual recharge for current conditions for water years 1960–2001 was estimated to be about 1.55 in. or only 92 ft^3/s. This relatively small amount of recharge is due to the general lack of human activities (especially surface-water irrigation) and the low mean annual precipitation in this area. The mean annual precipitation for the 1971–2000 climate normal (Spatial Climate Analysis Service, 2006) in this area ranged from about 6 to 31 in., with more than 59 percent of the area receiving less than 10 in. The greater precipitation quantities mainly occur in the uplands in the northern-most part of this area near Park and Caribou Creeks.

The spatial distribution of current-condition mean annual recharge for water years 1960–2001 for the extended study area (fig. 16) shows the large spatial variability of recharge. Within the area of the Yakima River basin analyzed by Vaccaro and Olsen (2007a), mean annual recharge for this base period was about 15.2 in. or 6,820 ft^3/s (about 4.9 million acre-ft). These values account for the refinements in selected irrigation districts for the years with prorating and for the different base period (1960–2001 in contrast to 1950–2003). For the complete extended study area, mean annual recharge was estimated to be about 13.6 in. or 6,950 ft^3/s. Note that because much of the recharging water discharges locally, it is not available for use but supports streamflow.

The large quantity of recharge in the low-lying structural basins, where mean annual precipitation ranges from about 6 to 9 in., is due to the application of surface water to croplands and canal/lateral leakage. These factors account for the estimated 1,699 ft^3/s increase in recharge from predevelopment to current conditions within the area modeled (6,100 mi^2); the increase in recharge from predevelopment to current conditions for the 808 mi^2 area (described above) was only about 8 ft^3/s. The distribution of these large quantities of recharge is consistent with the location of the surface-water irrigation districts (fig. 17). The isotopic composition of surface water and shallow groundwater also confirms that irrigation is the source of increased recharge. In contrast, the large quantity of potential recharge in the uplands (where mean annual precipitation typically is greater than about 30–35 in.) is derived from rainfall and snowmelt. Small mean annual recharge quantities, generally less than 2 in. (fig. 16), occur where mean annual precipitation is low and there is no irrigation (primarily in the areas east of the structural basins). Mean annual recharge for current conditions was estimated to be less than 2 in. over about 37 percent of the extended study area (about 2,553 mi^2), and within this area recharge averaged about 0.45 in. or 86 ft^3/s (62,300 acre-ft).

The amount of the estimated potential recharge that actually enters the aquifer system is a function of the hydraulic characteristics of the surficial hydrogeologic units, the depth to the water table, and land-surface slope. For example, coarse-grained surficial units such as the alluvial units (table 1 and fig. 8) can readily accept recharge. In contrast, most of the recharge in the bedrock-floored, humid, forested-upland slopes of the Cascade geologic province (where the bedrock units occur) discharges as shallow subsurface groundwater flow to upland streams (Mastin and Vaccaro, 2002), and thus is not available to recharge the deeper parts of the aquifer system. Recharge in some of these upland areas, especially in Kittitas County, is limited because the bedrock generally has a much lower hydraulic conductivity and infiltration capacity than the overlying soils and (or) unconsolidated deposits, which also are generally thin or missing (Jones and others, 2006; fig. 8). On a long-term basis, only about 34 percent of the potential mean annual recharge (percentage varies on an interannual basis) was estimated to have entered the regional groundwater flow system in the bedrock units in the upper Yakima River basin in Kittitas County (Vaccaro and Olsen, 2007a; Mastin and Vaccaro, 2002, fig. 6). In addition, the steep slopes in these areas, combined with the conductivity contrast between the soils (and any developed porous regolith) and the bedrock, result in a relatively large lateral component of flow that limits vertical movement (Polubarinova-Kochina, 1962)—the steeper slopes increase the probability of potential recharge moving laterally to discharge as shallow subsurface flow. In contrast, the surficial CRBG units generally can accept most of the potential recharge (Davies-Smith and others, 1988; Bauer and Vaccaro, 1990; Hansen and others, 1994). In the dissected outcrop areas of the CRBG, however, the potential recharge to deeper parts of the units would be limited due to local discharge from the shallow groundwater flow system to streams and "cuts" along canyon walls. This recharge supports local streamflow and evapotranspiration. In some areas, the water table may be near the land surface during some periods of the year, and thus part of the recharge may be rejected and would be expressed as shallow subsurface groundwater flow and (or) overland flow (surface runoff).

The recharge that reaches the water table may have traveled for less than a day from below the root zone in areas with shallow water tables to as long as tens or even hundreds of years where the water table is deep. In the case of a shallow water table, the pulses of recharge are reflected by seasonal to event variations in groundwater levels. In areas with a deep water table, such pulses are attenuated with depth and approach a constant rate that is best approximated by an estimate of long-term mean annual recharge. In these latter areas, variations in water levels typically are caused by pumpage and (or) vertical leakage.

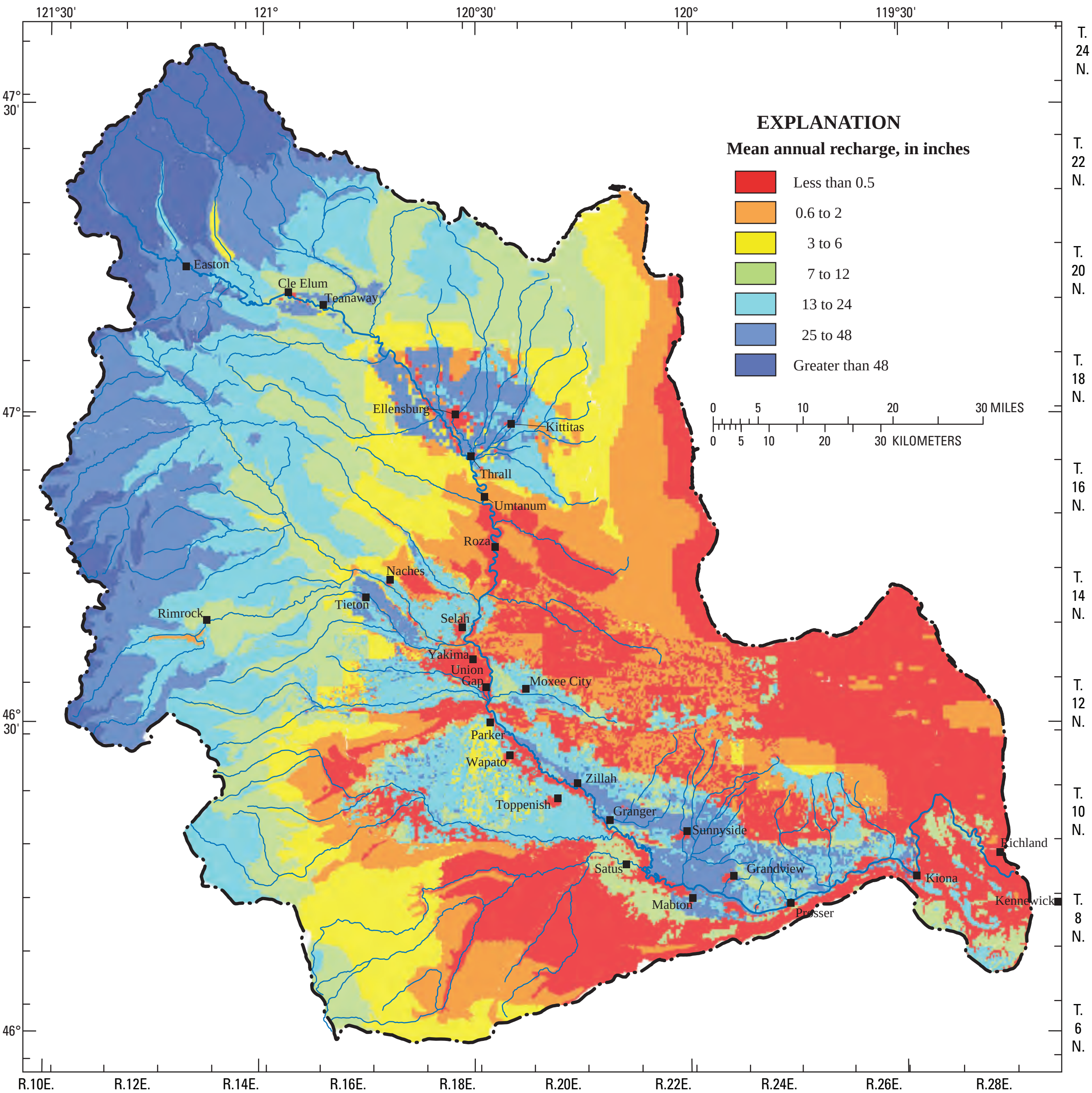


Figure 16. Spatial distribution of mean annual recharge for current conditions, 1960–2001, Yakima River basin aquifer system, Washington.

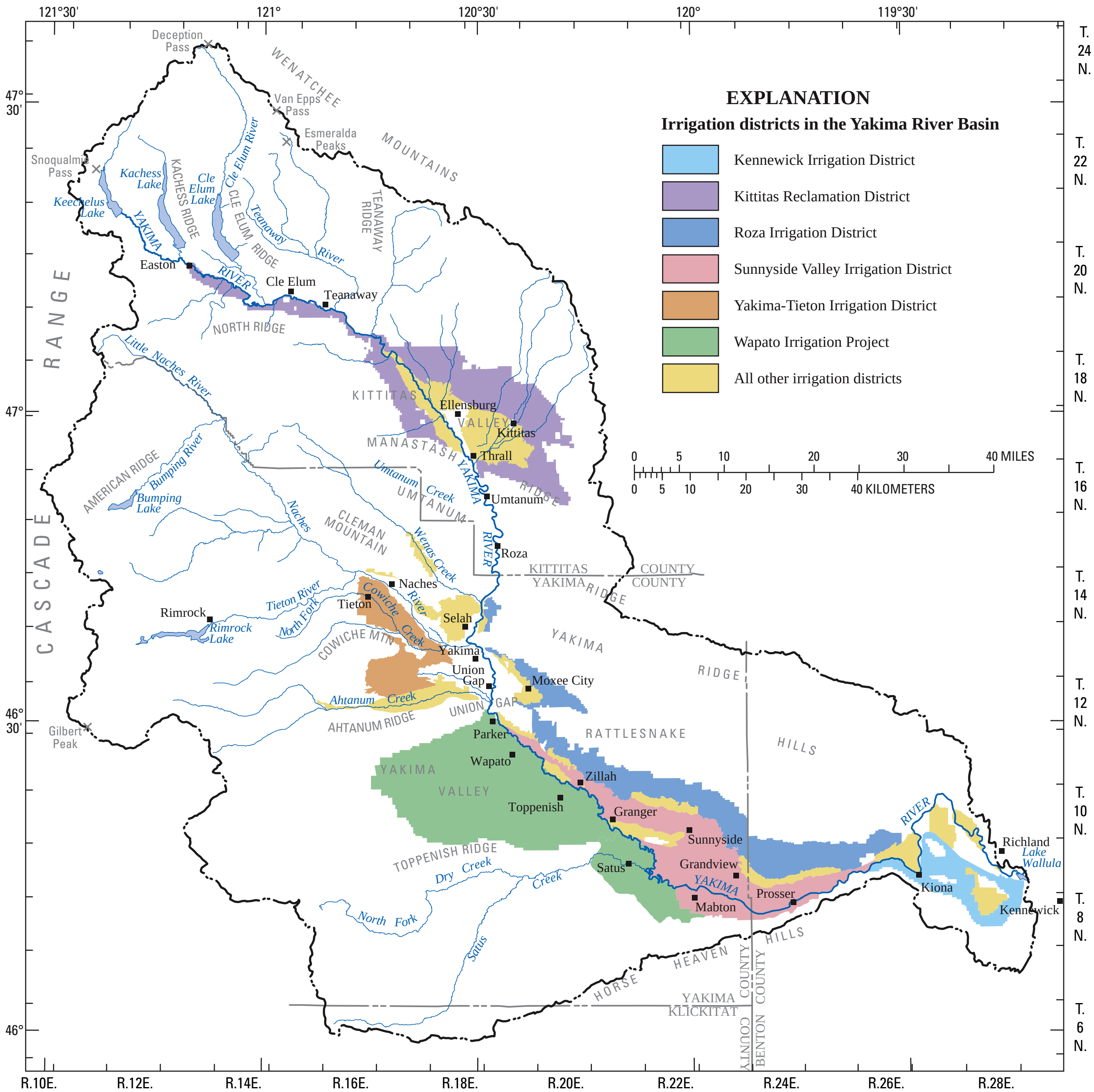


Figure 17. Surface-water irrigation districts, Yakima River basin, Washington.

Within the Yakima River basin, the percentage (rounded) of recharge that occurs in the outcrop areas of the hydrogeologic units is about 0.29 percent for the basin-fill units, 0.007 percent for the Saddle Mountains, 0.008 percent for the Wanapum, 0.13 percent for the Grande Ronde, and 0.57 percent for the bedrock units. After recharge enters the aquifer system, groundwater flows along local to regional flow paths. Local flow paths are on the order of feet to a few miles, and regional flow paths generally are on the order of 5–50 mi. Isotope analysis of groundwater samples (described in the previous section; see also Hearn and others, 1985; Hendry and others, 1992; Wagner and Lane, 1994; Steinkampf and Hearn, 1996; Reidel and others, 2002; Gazis and others, 2007) indicates that groundwater in the surficial units is of recent age; groundwater becomes progressively older with depth, and in parts of the more deeply buried units in the basin, groundwater is as much as and in some cases, more than, 10,000 years old. These variations in groundwater age highlight the differences between the local and regional components of groundwater flow.

Groundwater occurring in the different hydrogeologic units moves along various flow paths at rates that depend on the hydraulic characteristics of the units and the hydraulic gradients. Hydraulic characteristics and hydraulic gradients are highly correlated because gradients tend to be steeper in poorly transmissive units and flatter in more transmissive units. Water moves readily in the unconsolidated units because of the large interconnected pore spaces (interstices) in the silty sands to gravels and cobbles. Conversely, the rate of movement is slow through the small and less interconnected pore spaces in the clays present in the units. The semi-consolidated to consolidated sandstone of the consolidated units commonly has well-connected but small pore spaces, but compressional forces and (or) heat from historical volcanism can reduce the pore spaces in sandstone. The shale, mudstone, siltstone, and claystone in these units have limited interconnected pore spaces, which impedes and slows the movement of water. The unconsolidated to conglomerate lenses/layers within the consolidated units can have well-connected pore spaces. Wells completed in the consolidated units generally derive their water from sandstone and (or) layers and lenses of unconsolidated and conglomerate deposits, where most of the groundwater in these units occurs and moves.

In the CRBG units, water occurs in fractures, interconnected vesicles, and basalt flow-top breccia, scoria, and clinker. Most of the water in these units moves through the vesicles, rubble, and other interflow material, so that the direction of groundwater flow is predominantly parallel to the structural gradient/dip of the basalt. The occurrence and movement of groundwater between basalt interflow zones is along crevices and fractures in the entablature and colonnade. Water movement in this part of a basalt flow is predominantly vertical but may be reasonably connected laterally in areas with an established colonnade. Although individual flows of the CRBG can range in thickness from a few inches to more than 300 ft, the interflow zones (where most of the groundwater occurs and moves) constitute, on average, only about 5 to 10 percent of the thickness of an individual basalt flow (Swanson and others, 1979a), which limits the occurrence and movement of groundwater in the basalts. That is, large quantities of groundwater can move through parts of the CRBG units (large K_h values), but their overall ability to store water is small (small storage coefficient).

In the bedrock units, groundwater occurs in and moves through fracture systems in the Paleozoic metamorphic rocks (gneiss, schist, phyllite, and amphibolite) present in the Manastash and Taneum Creek drainages and the Mesozoic ultramafic rocks in the upper Cle Elum and Teanaway River drainages. In other bedrock units, groundwater occurs in and moves through interconnected interstices in shales and sandstones, and through fractures and vesicles in basalt and andesite. Except for the Quaternary unit, the bedrock units generally are not as conductive as the CRBG units or parts of the consolidated units (table 4). However, joints within the crystalline rocks are locally numerous enough to provide sufficient permeability to support some groundwater withdrawals, and groundwater in the interstices in the sandstone layers generally can also support withdrawals. In some areas, the Tertiary sedimentary unit is composed predominantly of clay and shale and the availability of groundwater is limited. However, usable quantities of water can be obtained from some layers that are recharged in an outcrop area and (or) by older water entering the layers by vertical flow through overlying or underlying units. The Quaternary bedrock unit, which generally is near the crest of the Cascade Range, probably is permeable on the basis of information for similar rock types of similar ages; for example, the Quaternary basalts of the eastern Snake River Plain aquifer system where groundwater occurs in fractures, flow tops, and vesicles have been shown to be highly permeable (Garabedian, 1986; Lindolm, 1996). Gannett and Lite (2004) also show that these young volcanics are highly permeable. However, too few wells are completed in this unit to provide data to assess its permeability and occurrence.

The amount of groundwater in storage can be roughly estimated for the basin-fill, bedrock, and CRBG units based on the volume of units and an estimate of their effective porosities. The total thickness used in the volume calculation for the bedrock and CRBG units was limited to 3,000 ft for this analysis. Due to the large variations in lithology of all units and the paucity of information on effective porosity, single values were used in the calculations. These values were selected to derive 'conservative' estimates of the volume of water in storage in the aquifer system. A value of 0.1 was selected as an estimate for the clay, sand, gravel, sandstone, and shale in the basin-fill units. For the CRBG units, a conservative value of 0.04 was selected. Sublette (1986) reported that the effective porosity for different parts of the basalt flow ranges from 0.002 to 0.41, with representative values between 0.02 and 0.14. The lower representative values are for the basalt flow interiors and thus, the selected effective porosity value of 0.04 for the CRBG units represents the volume of basalt consisting primarily of flow interiors. For the older bedrock units, a value of 0.01 was selected for the analysis. The large range in lithology, as well as the effects of diagenesis, fractures, and overburden pressure likely result in spatial variations in the effective porosity values of the bedrock-unit ranging from 0.001 to 0.45. On the basis of these porosity estimates, the amount of water in storage in the basin-fill, CRBG, and bedrock units is 43.5, 277, and 52.6 million acre-ft, respectively. Several factors should be noted for these estimates. First, they are generalized and should be considered order-of-magnitude estimates. Second, they are considered to be the same for both predevelopment and current conditions due to the small change in storage caused by groundwater pumpage.These volumes should not be interpreted as available in their entirety to meet water-supply needs; dewatering of any aquifer is environmentally and economically undesirable. The amount of water that may be available is related to water quality and environmental, economic, and legal constraints. Indeed, estimated recharge is only about 5 million acre-ft (about 2 million acre-ft after accounting for discharge to streams in the bedrock uplands), and the recharge supports in-stream and out of stream uses. Water in storage is important, however, because it is the groundwater that moves within and through the aquifer system.

Conditions of Occurrence

Groundwater in the aquifer system occurs under water-table to confined conditions. The water table occurs in the surficial hydrogeologic units throughout the study area. Groundwater in units below the water table occurs principally under semiconfined to confined conditions, but is locally unconfined where a perched water table exists. The depth to the water table (DTW) was mapped in the structural basin and in areas outside the basins where data were sufficient (fig. 18). The generalized map shows that the water table is shallowest in the topographically smooth (low slope) parts of the structural basins and is deeper in other parts of the system (fig. 18). However, in arid areas such as the Hanford Site (fig. 1), the low recharge quantities results in a large DTW in the low slope area.

The DTW reflects the relation between the quantity of recharge, hydraulic characteristics of units, and topographic setting. The DTW shown on figure 18 is generalized and does not represent any particular year or season but rather identifies typical ranges. Locally, DTW is highly variable because of topographic variations. In the areas where the DTW is not shown, there was a lack of water-level data, and in some areas the interval boundaries of 81–200 ft and greater than 200 ft are highly generalized due to sparse data. These factors are due to the fact that of the more than 10,000 water levels analyzed, most were in the structural basins, and secondarily, many of the remaining water levels were not representative of the water table due to the groundwater conditions of the unit or units in which a well was completed.

The water table is within about 0–20 ft of land surface in the low slope parts of the structural basins that receive recharge from surface-water irrigation and in the alluvial river valleys with perennial streams (fig. 18). The proximity of the water table to land surface is highly variable throughout the remainder of the aquifer system, ranging from a few feet to more than 500 ft. The shallow water table in parts of the surface-water irrigated areas is due to the increased recharge (described previously) that started in the late 1890s. For example, in describing the relation between irrigation and drainage, Jayne (1907) presents information for nine wells in the Sunnyside area dug between 1890–1900 that shows that by 1902 there were water-level rises from 14 to 75 ft (the differences in the magnitude of the rises were due to the location of a particular well). Such water-level rises throughout the irrigated areas during this period resulted in the formation of drainage districts for the purpose of constructing drainage ditches, also called drains (Jayne, 1907). The shallow water table in the structural basins suggests a readily available supply of groundwater, but much of this irrigation-derived water in the shallow system discharges to drains and streams and is relied on to meet downstream uses (both in-stream flows and diversions with entitlements). Areas with high water tables (shallow groundwater) also are more susceptible to contamination from sources at the land surface (Sumioka, 1998).

The complexity of DTW is due to not only local topographic variations but also to local variations in the properties of the hydrogeologic units. For example, in some areas there are numerous domestic wells with depths between 200–400 ft and a DTW on the order of 150 ft. However, there also are some interspersed, shallow (less than 50 ft deep) wells in which the DTW is about 20 ft. In such areas, the shallow wells may be tapping a localized perched aquifer that locally overlies a low-conductivity layer. Such perched zones were described by Newcomb as early as 1959 (Newcomb, 1959). Perched water tables generally indicate a combination of limited recharge and an underlying impeding layer. For the CRBG units, perched water would be most prevalent in the low recharge areas east of the Yakima River. Thus, the mapped DTW in some of these areas may be most representative of perched water.

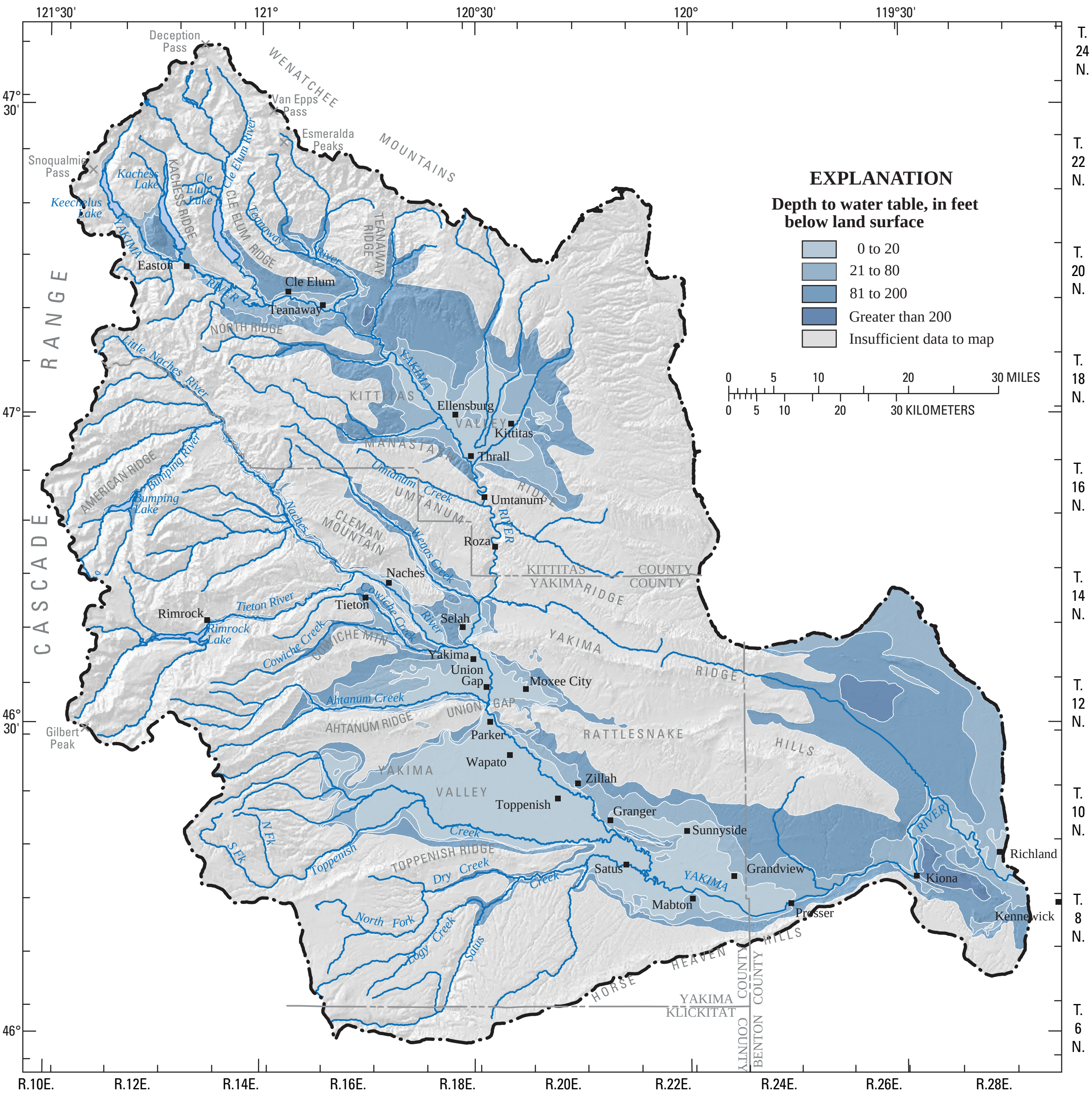


Figure 18. Generalized depth to water table for parts of the Yakima River basin aquifer system, Washington.

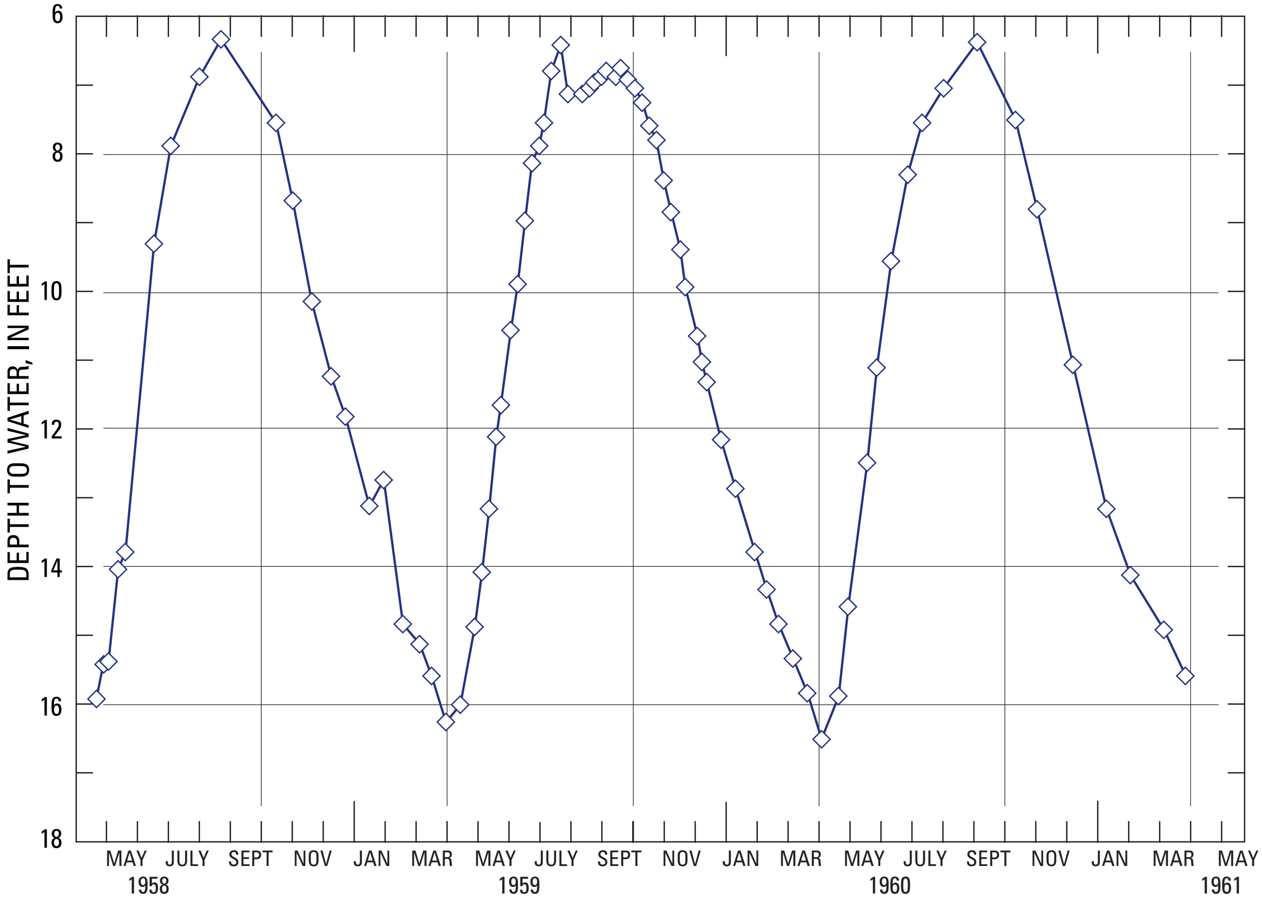
DTW fluctuates on both a seasonal and interannual basis due to natural variations in precipitation and (or) human activities. For example, a hydrograph (fig. 19*A*) for well 10N/19E-05R01 (a 20.5-ft deep well) that is completed in Unit 2 of the Toppenish basin shows distinct fluctuations, on the order of 10 ft, due to seasonally varying recharge. The fluctuations are out of phase with typical expected fluctuations due to recharge from precipitation because most of the recharge in this area (the Wapato Irrigation Project [WIP], fig. 17) is derived from canal/lateral leakage and applied irrigation surface water. About 80 percent of the mean annual recharge at the location of this well (where mean annual precipitation is about 7.5 in.) occurs during the April–October period. Annual recharge was estimated to range from 19.7 to 22.2 in. between 1958 and 1960, nearly three times larger than annual precipitation. The water-level fluctuations and estimated recharge quantities indicate that the specific yield of the unit is about 0.16–0.19, which is consistent with the deposits described on the drillers' log and on logs for nearby wells.

Similar water-level fluctuations are observed in most other irrigation districts. For example, Drost and others (1997) present seasonal hydrographs for wells in the surface-water irrigated areas in eastern Benton County that peak from August through December and reach their low from March to May. The seasonal rises also are attributed to the combined effects of canal/lateral leakage and infiltration of irrigation water. The median seasonal changes in water levels, which were classified on the basis of the dominant influence that produced the seasonal trend, for numerous wells ranged from 1 to 6 ft with maximum changes greater than 31 ft. The largest rises were associated with the wells that were classified as being most influenced by leakage from canals/laterals, especially unlined canals and wells completed in the Pasco gravels.

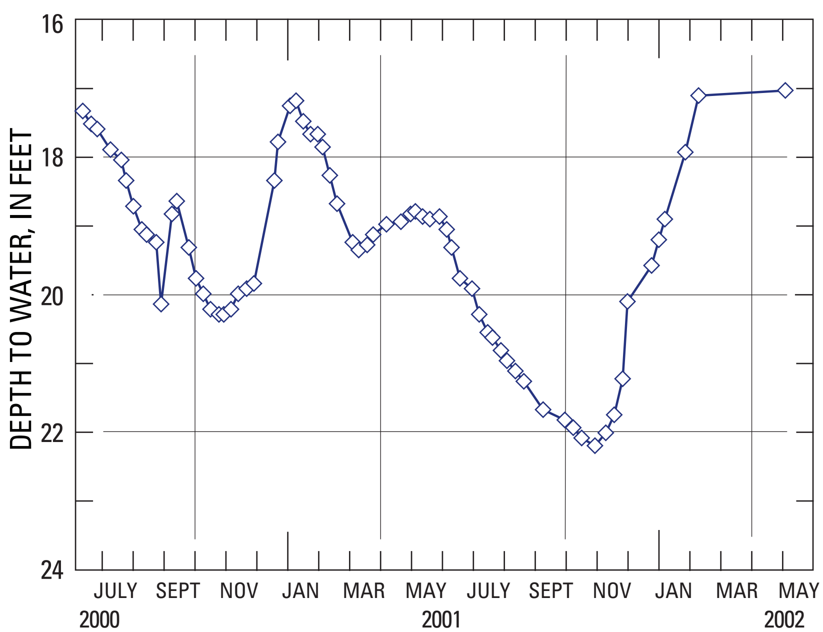
From late August through at least the beginning of the next irrigation season, the water table in surface-water irrigated areas declines due to groundwater discharge to streams, tile drains, drains, and wasteways. Typically, the drainage continues until the next irrigation season, with the lowest water levels generally occurring between mid-March to early May (fig. 19*A*; Drost and others, 1997). The variations in timing of the highs and lows are a function of location, local crop type, and year (interannual climatic variations). Although recharge can occur during the winter, the rate of groundwater drainage generally is more than the winter recharge rate. Under natural conditions, soil moisture in the irrigated areas would be near or at wilting point by about October, but the moisture added to the soil column during the irrigation season allows for winter recharge in these dry areas. However, in the southern structural basins the winter recharge quantities are smaller than the irrigation season quantities because winter precipitation is generally less than 8 in. In addition, K_h values of the surficial deposits in the irrigation districts generally are large enough to allow for groundwater drainage to exceed recharge during the winter.

Water-level hydrographs for water-table wells in areas where recharge is derived from precipitation (rainfall and snowmelt) have a shape similar to those affected by irrigation, but the highest water levels occur in the winter-summer period (January–July, depending on location and year) and the lowest in the fall-winter period (September–February, also depending on location and year). For example, a hydrograph for a 30–ft deep well (10N/16E-24E01) completed in coarse-grained deposits near Toppenish Creek shows distinct seasonal fluctuations with highs between January and February and lows in October (fig. 19*B*). The fluctuations are due to recharge from precipitation and are smaller than fluctuations due to surface-water irrigation. Additional water-level fluctuations correspond to runoff variations (stream losses) in Toppenish Creek. Water-level hydrographs in the wetter upland areas with a large seasonal snowpack generally have later seasonal highs, typically between April and June, but sometimes in July.

Hydrographs for wells in or near major irrigation pumping centers are similar to hydrographs for water-table wells in areas where recharge is derived from precipitation, such as shown in figure 19*B*. Hydrographs of irrigation wells in these areas show the effects of seasonal pumpage. For example, in the hydrograph for a 1,510-ft deep composite CRBG unit well (10N/17E-27Q01, fig. 19*C*) the highest water levels are in late March to early April and the lowest in late September, which is typical of CRBG wells throughout the Columbia Plateau (Hansen and others, 1994; Whiteman and others, 1994). The seasonal water-level range in this well is about 23 ft, but Drost and others (1997) show that the seasonal range in basalt-irrigation wells can be as much as 95 ft. Typically, in the areas of major basalt-irrigation pumpage, the water levels decline throughout an area encompassing the pumping centers during the irrigation season and begin to rise after the end of the irrigation season; these seasonal fluctuations generally are consistent over large areas. For the deeply buried basalt units that do not outcrop, the rises can be attributed to vertical leakage from overlying/underlying basalt flows or units. In areas experiencing long-term groundwater-level declines, the levels do not rise to those observed prior to the irrigation season because locally the pumpage is larger than vertical leakage, lateral inflow, and (or) recharge.

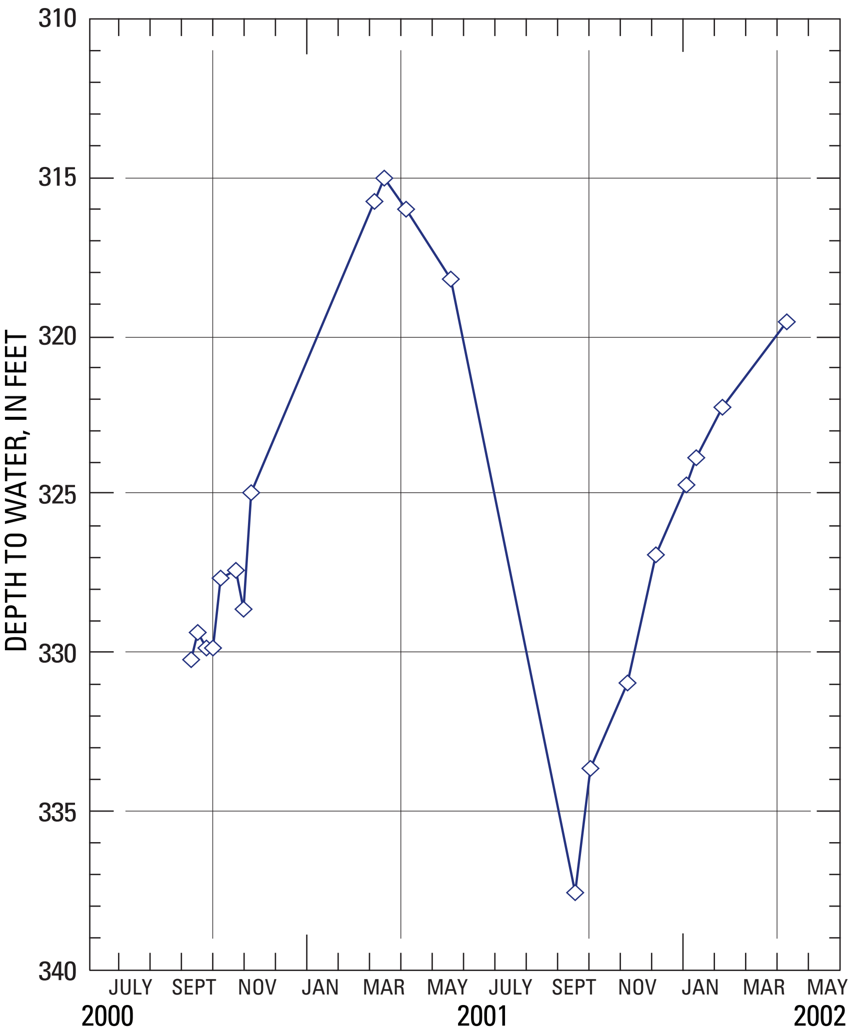


***A.* 10N/19E-05R01**

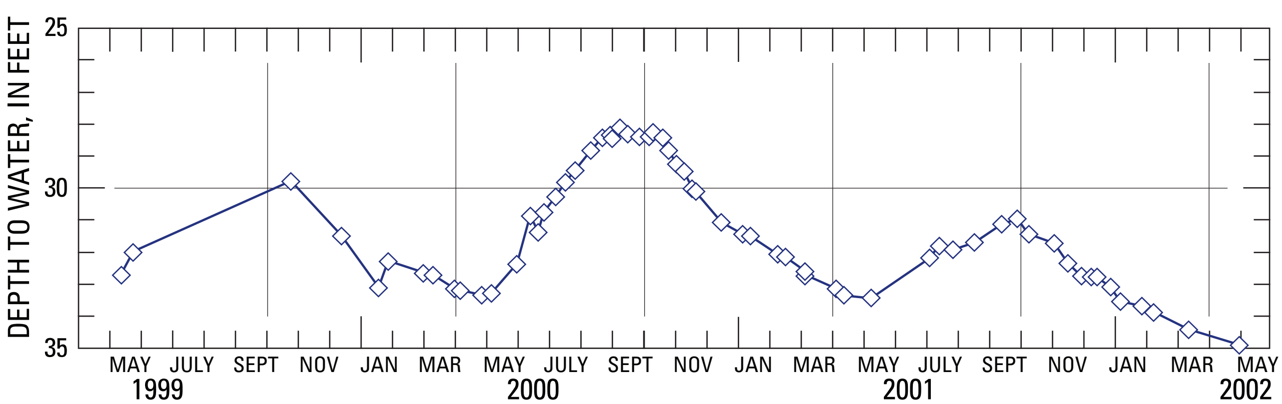


***B.* 10N/16E-24E01**

Figure 19. Depth-to-water in four wells, Yakima River basin, Washington.



C. **10N/17E-27Q01**



D. **8N/22E-03M01**

Figure 19.—Continued.

The hydrograph for well 8N/22E-03M01 (fig. 19*D*), which is completed in unconsolidated deposits of Unit 1 of the Benton basin near the boundary of WIP, also shows distinct seasonal fluctuations due to recharge. This 167-ft deep well is adjacent to sloping uplands, and water-table wells located in such areas, especially where coarse-grained deposits are overlain by fine-grained deposits such as the Touchet Beds (table 1), have smaller seasonal fluctuations than wells tapping the water table in the flatter, central parts of irrigation districts (Hansen and others, 1994). Fluctuations in the central part of WIP can be more than 15 ft (Kinnison and Sceva, 1963; U.S. Geological Survey, 1975). The hydrograph also shows a distinct interannual variation due to the interannual variation in recharge. For example, water year 2001 was an extreme proratable (about 37 percent) drought year and as a result, not only was natural recharge much less than average, but the application of surface water to crop lands in WIP (more than 50 percent of its surface-water entitlement is proratable) also was diminished in comparison to non-proratable years, as was the amount of water present in the canals, laterals, and drains—resulting in a deeper water table and a smaller seasonal fluctuation (fig. 19*D*). The 3-ft deeper water table, combined with the specific yield for this area described previously, indicates a decrease in recharge in WIP of at least 6 in. This decrease represents a relatively large quantity of recharge; for example, in the study area mean annual recharge was estimated to be less than 6 in. over about 3,175 mi^2 and less than 2 in. over about 2,553 mi^2.

Unconfined conditions are highly variable in the study area. For example, water levels in two closely spaced wells (530-ft deep 8N/30E-22R03 and 52-ft deep 8N/30E-22R04) that are completed in the Saddle Mountains are about the same and information from the drillers' logs and seasonal water levels indicate that both function as water-table wells; interbeds are minimal in this location. In other areas, however, water-level data indicate that confined conditions exist at much shallower depths than 530 ft. These types of differences are principally due to the presence or absence and the highly variable nature of interbeds in the CRBG units, the hydraulic characteristics of individual basalt flows, and (or) the spatial variations in the lithology of the consolidated units. For the latter, spatial variation is explained by basin-fill deposits that had a variety of sources and were emplaced under a wide range of depositional-climatic regimes, and the basin-fill deposits and the CRBG units were subsequently affected by erosion and structural deformation. In areas where the CRBG units are at or near land surface and the quantity of recharge is small, groundwater also can occur under perched conditions.

Variations in unconfined conditions also occur in the unconsolidated units in the structural basins. For example, wells deeper than 400 ft in parts of the Toppenish basin are water-table wells, whereas an 8-ft deep well in the Ahtanum-Moxee basin taps a confined flowing artesian zone. In some areas, wells tap shallow unconfined groundwater in cemented gravels that function as an unconfined aquifer, whereas nearby wells that penetrate through the cemented gravels tap an underlying confined aquifer confined by the cemented gravels. In a few areas, these differences may be attributed to fine-grained terrace deposits overlying the gravels. These types of variations appear to be most prevalent in the Ahtanum subbasin (Foxworthy, 1962).

There are many occurrences of flowing artesian zones in the aquifer system (fig. 20*A*). Groundwater in these zones occurs under confined conditions, which are prevalent in many parts of the aquifer system. Water levels in wells in these zones are higher than the land surface, and thus uncapped wells tapping these zones will flow. These conditions in the basin were described as early as 1901 (Smith, 1901) and were later discussed by Waring (1913) for parts of the basin, by Molenaar (1961), and by Foxworthy (1962) for the Ahtanum subbasin. Artesian conditions generally occur where a water-bearing unit dips beneath a unit of low conductivity, such as a clay, shale, or a tight sandstone. For flowing artesian zones in confined aquifers, the water-bearing unit crops out (or is overlain by thin or conductive deposits) and is recharged in an area of higher altitude than the overlying low-conductivity unit. The outcrop area is the source of high-pressure head for the artesian zone. Discharge from the zones also is impeded due to the presence of the fine-grained overlying unit. For example, upward vertical discharge from the Saddle Mountains and Unit 5 in the Toppenish basin is impeded by the overlying fine-grained Unit 4 (table 1). These conditions are present particularly in basin-like structures, which are represented by synclines in the study area, and such conditions exist in many areas of the aquifer system (fig. 20*A*).

The highly variable character and wide-spread distribution of artesian zones is indicated by the location and depth of more than 500 flowing wells (fig. 20*A*). The wells represent both historical and existing flowing wells on the basis of information from Molenaar (1961), NWIS, and review of more than 10,000 well logs. The measured discharge for 251 of these wells ranged from 0.125 to 3,900 gal/min and averaged 230 gal/min. In some areas, such as the upper part of the Moxee subbasins, groundwater pumpage from these zones has resulted in the reduction of hydraulic head (pressure) and the elimination of flowing artesian conditions. As another example, hydrographs for wells 13N/24E-27K02 and 13N/24E-36D01 show the cessation of flowing conditions due to pumpage and well 13N/24E-27K03 (drilled to replace 27K02) shows a continual decline of hydraulic heads (fig. 21*A*). In addition, unpublished field notes from 1971 for a water-level measurement of a CRBG well indicate that four CRBG-unit wells (11N/17E-13A, -14Q, -15P, -22D; actual wells not identified) stopped flowing after the drilling and subsequent pumping of a nearby, deep basalt well used for irrigation. As a final example, flow from well 13N/24E-24E01 decreased from 3,900 gal/min in 1925 to 697 gal/min in 1951, and of 136 water levels measured between January, 1979 and February, 1983, only one was above land surface—essentially, the well no longer flows.

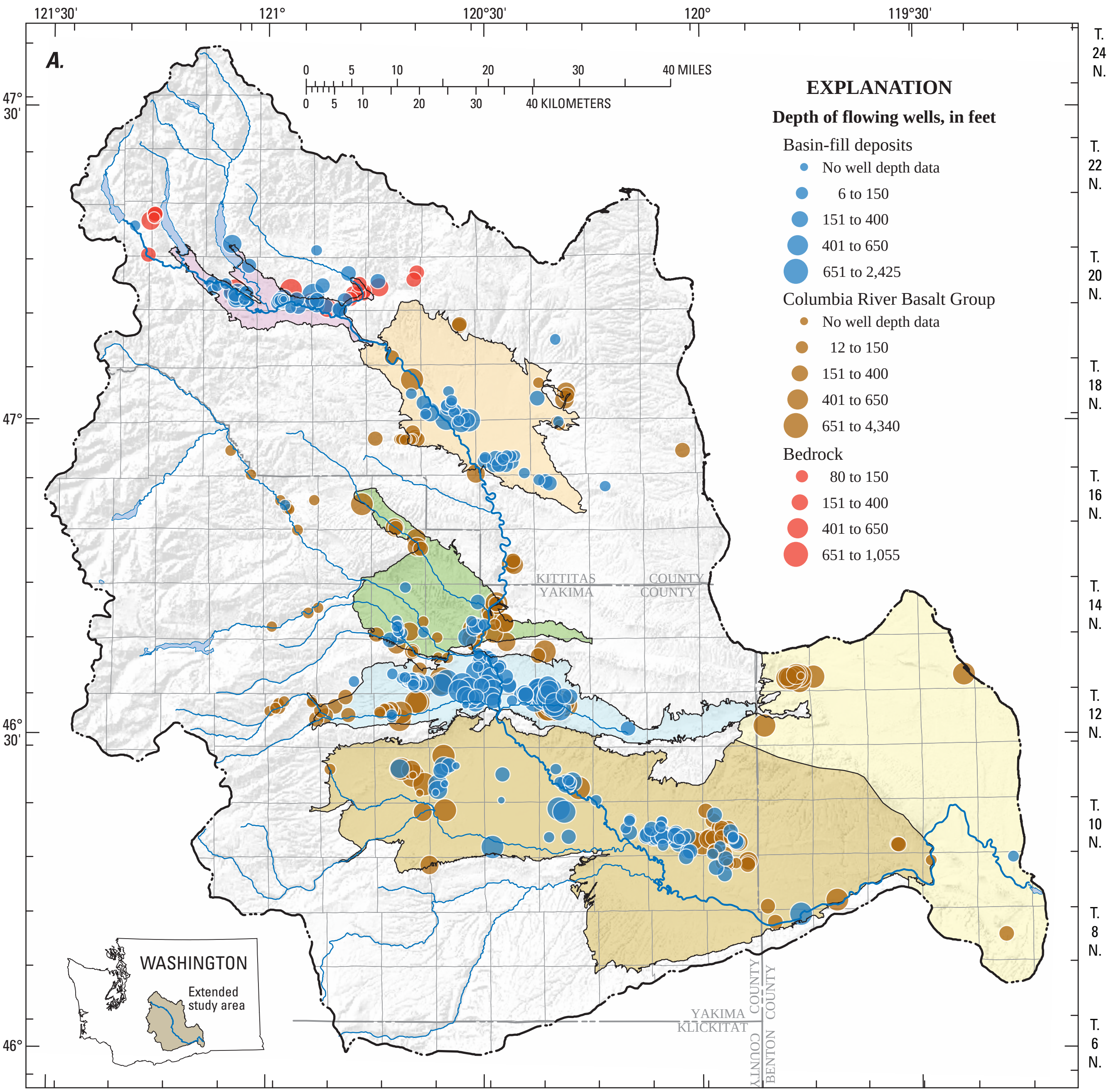
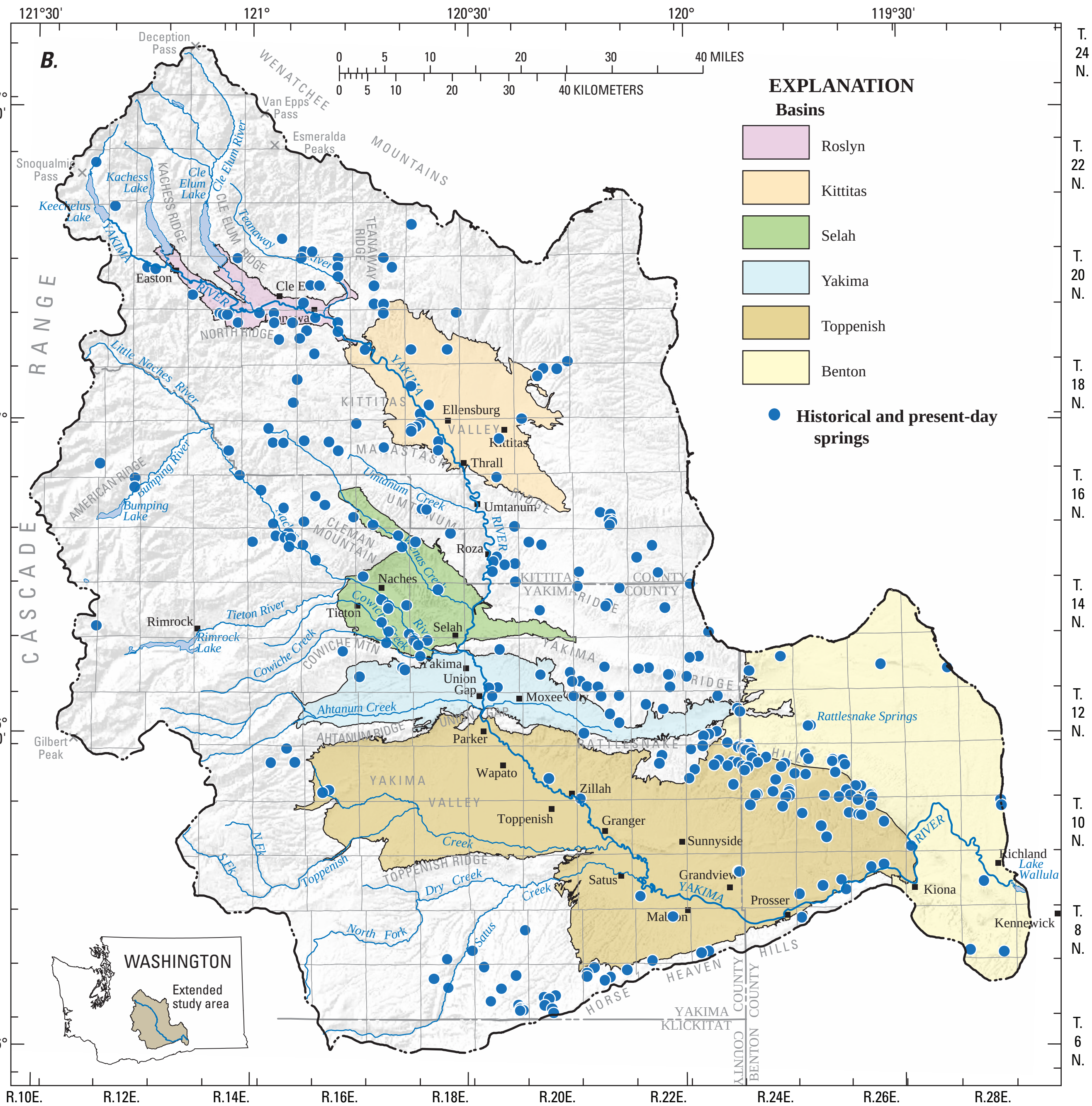
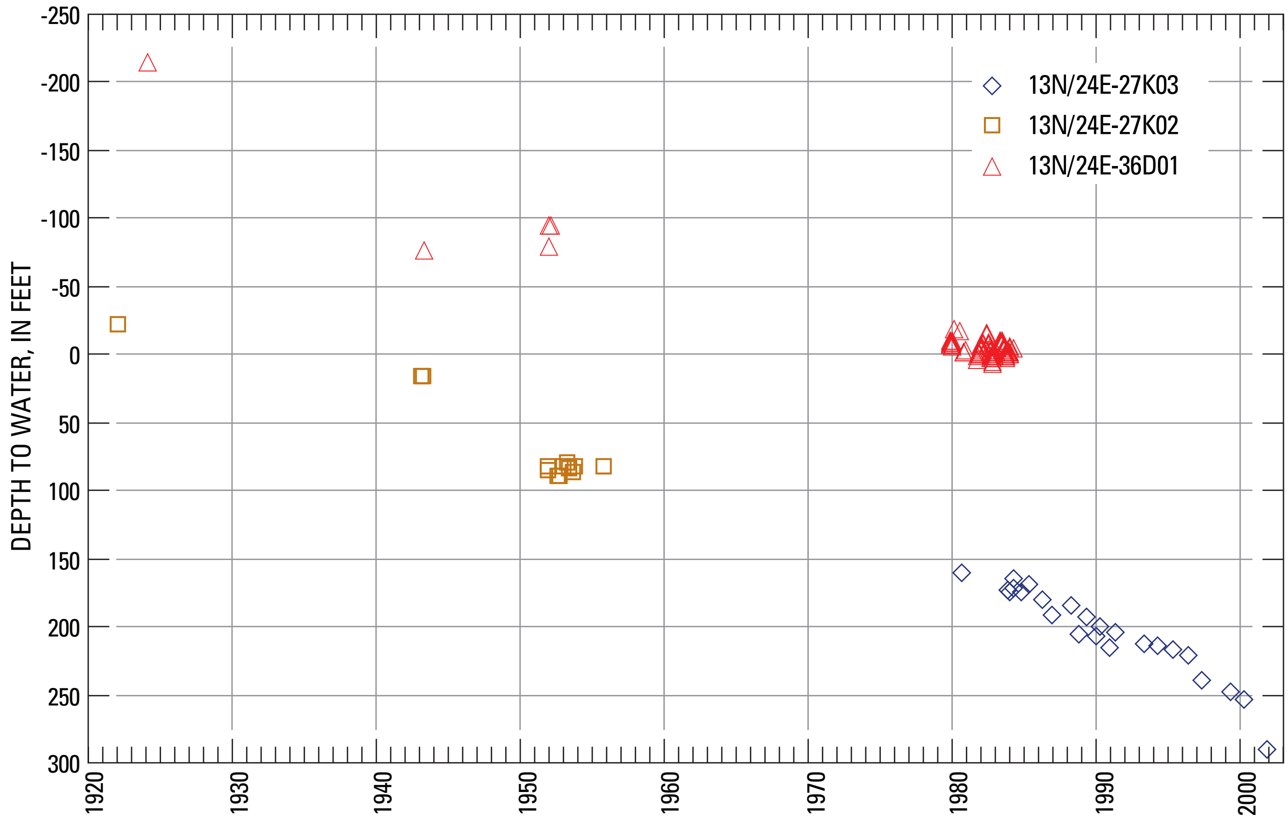


Figure 20. Location and depth of historical and present-day (*A*) flowing wells, and (*B*) location of historical and present-day springs, Yakima River basin aquifer system, Washington.

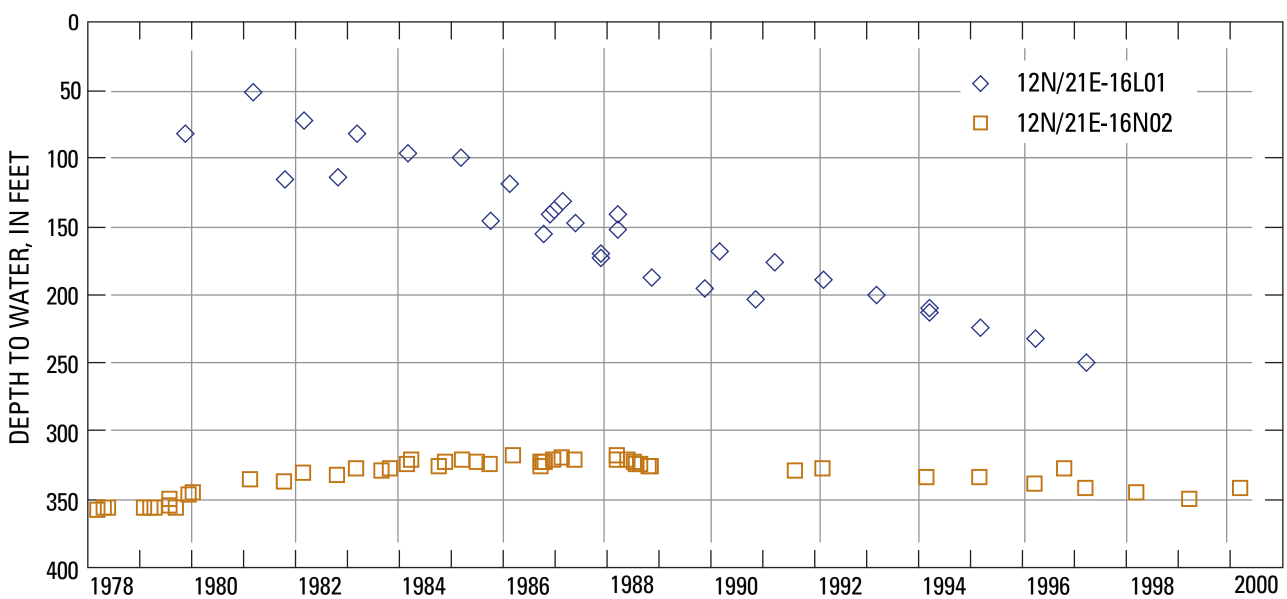


Base from U.S. Geological Survey and/or
Washington Division of Geology and Earth Resources
digital data, 1983, 1:100,000
Universal Transverse Mercator projection, Zone 10
Horizontal Datum: North American Datum of 1927 (NAD 27)

Figure 20.—Continued.



A. **Wanapum unit**



B. **Wanapum and Saddle Mountains units**

Figure 21. Depth-to-water in (*A*) wells completed in the Wanapum unit, and (*B*) wells completed in the Wanapum (12N/21E-16L01) and Saddle Mountains (12N21E-16N02) units, Yakima River basin aquifer system, Washington.

Flowing artesian conditions exist locally in some areas, whereas in other areas, such as the Yakima basin (Ahtanum and Moxee subbasins), there are large, areally-extensive productive artesian zones where groundwater occurs under confined conditions. Local flowing artesian zones occur in many locations, and these zones can be isolated from nearby units. For example in the Grande Ronde, well 19N/18E-10SW (quarter sections M, L, N, P) penetrated a flowing zone at 346 ft but the water level dropped after drilling through this zone and the final water level in this 763-ft well was 590 ft below land surface. In contrast, two nearby wells (-10D and -10F) drilled to depths of 520 and 380 ft, respectively, did not encounter this zone, and their water levels were between 275 and 290 ft below land surface. A 708-ft deep well (-10R) also did not encounter the flowing zone and had a water level of 592 ft below land surface. These two wells may have encountered this flowing zone but the hydraulic head may have been reduced by cross connections and a downward flow regime. (The water levels in the above wells also indicate downward flow in this area.) In the next section to the east (19N/18E-11), two wells (-11SW and -11L) tap a flowing artesian zone at depths of 322 and 225 ft, respectively, whereas a 310-ft well (-11G) is a water-table well. Together, this information suggests that the flowing zone is locally isolated from some of the surrounding deposits—further indicating the complexity of the aquifer system. Such complexity may be due to unknown structure that locally compartmentalizes the flow system.

Flowing artesian zones are most prevalent in the consolidated hydrogeologic units, the CRBG units, and in Unit 3 of the Roslyn basin, which is confined by an extensive lacustrine layer (Unit 2; table 2, pl. 1); only about 8 percent of identified flowing wells were completed in the bedrock units. Fine-grained lenses or layers in or overlying these units provide a continuous confining section of low conductivity. In addition, Smith (1901) hypothesized that compressional forces in some of the tighter structural folds, such as in the Moxee subbasin, compacted conductive layers (in this case sandstone) to reduce their conductivity, thus providing a continuous layer of low conductivity. As a result, the same sandstone layer can change from an aquifer to a semi-confining/confining unit over a distance of less than 10 mi. There also are occurrences of shallow wells tapping flowing zones in the surficial unconsolidated hydrogeologic units, as wells as in the bedrock units (fig. 20*A*).

In the CRBG units, a low-conductivity layer may consist of interbedded sediments or a basalt flow interior that is 'tighter' (exhibits low K_h and K_v values) than the underlying and (or) overlying flow (such as in the wells in 19N/18E-10, 11 described above). In some areas, water-level data suggest that the uppermost part of the Grande Ronde functions as semi-confining to confining layer, and it is not known if this is caused by the presence of a saprolite layer on top of the flow and (or) by the paucity of fractures. In addition, in the Wanapum in township 13N, range 24E, sections 25, 26, 27, and 36 (see fig. 21*A*, for examples), the flow top of the Priest Rapids member used to be a highly productive flowing artesian zone, suggesting that in this area the flow top functioned as an aquifer that is contained within lower conductivity zones (it is overlain by the Mabton unit that is mainly composed of shale in this location). Flowing artesian zones in tightly folded synclines within the CRBG in some parts of the basin are probably the result of compressional forces diminishing the interconnectedness of the basalt fracture system in the entablature and colonnade (resulting in even lower K_v values), similar to that described for sandstone. This is clearly evidenced by the exposed central part of folds of the ridges in the Umtanum Ridge, where the basalt is so compressed that there is an absence of well-defined interflow zones and colonnades.

Geologic structures such as faults can offset productive units/layers to create a flow barrier that results in a flowing artesian zone. These types of zones occur predominantly in the CRBG units. For example, some wells in 13N/24E-sections 25, 26, 27, and 36 (described above) that are upgradient (west) of the Cold Creek fault tapped flowing artesian zones in the CRBG, but downgradient of the structure water levels are lower. Thus, overlying fine-grained layers and geologic structure can combine to create a flowing zone. The effects of the Cold Creek fault on groundwater flow near the Hanford Site were summarized by Reidel and others (2002). Grande Ronde wells that range in depth from 62 to 280 ft in 17N/17E-sections 12-15 near Manastash Creek in the Kittitas basin tap a flowing zone that is upgradient of a mapped fault, but wells downgradient from the fault have lower water levels and do not flow. Two flowing basalt wells (15N/17E-12N01, Wanapum-Grande Ronde composite unit well, 550 ft deep and 15N/17E-11A01, Wanapum, 498 ft deep) that are just upgradient of a mapped fault had measured closed-in pressures of 167 and 85 lb/in^2, respectively. The artesian zone in the Wanapum that is tapped by well 11A01 (which was noted in the driller's log for well 12N01) is overlain by sandstone and shale. Well 12N01 taps artesian zones in both the Wanapum and Grande Ronde, which is overlain by a thick sequence of the Vantage interbed (principally sandstone). The presence of other such artesian zones is suggested by the location of the flowing wells and nearby structure perpendicular to the flow direction. However, structure can affect the movement of groundwater in the basalts without creating flowing zones. For example, along the Willow Creek monocline in north central Oregon, the lateral hydraulic gradient increases upgradient of the monocline and the vertical gradient changes from downward to upward near the monocline; just downgradient of the monocline, the gradient decreases and the vertical gradient changes to downward (Davies-Smith and others, 1988). Water levels downgradient of the monocline indicate that groundwater is moving through the monocline—it impedes but does not block flow.

Faults can impede lateral flow and can also provide an area where vertical movement of water can be enhanced or impeded (Newcomb, 1961, 1969). For example, Hammer (1934), in describing the origin and low pressure of methane in the Rattlesnake Hills gas field, where producing zones ranged from about 300 to more than 700 ft below land surface, shows that gas pressures increased in wet years due to downward migration of recharge (increased overlying pressure head) and decreased in dry years due to less vertical impedance (with more gas moving to the atmosphere). The effects of structure on groundwater flow in the CRBG was described early by Newcomb (1961, 1969) and was later discussed by Steinkampf and others (1985), Davies-Smith and others (1988), Hansen and others (1994), Whiteman and others (1994), Packard and others (1996), and Steinkampf and Hearn (1996).

Structurally controlled artesian zones also occur in both the young unconsolidated units and in the bedrock units. In the unconsolidated units, at least 15 shallow wells in 18N/18E-27 ranging in depth from 48 to 127 ft tap a local flowing zone composed of cemented gravels. The existence of nearby water-table wells suggests that there is some type of a localized structural control (groundwater flow barrier) in this area; such groundwater levels can be used to identify flow barriers (Steinkampf and others, 1985; Packard and others, 1996). In the bedrock units, the flowing zones likely are associated with either faults and (or) lateral and vertical changes in rock types, such as shale abutting sandstone.

Springs in the basin (fig. 20*B*) may be representative of flowing artesian zones, in which case they typically are related to geologic structure. The locations of 362 springs (fig. 20*B*) were obtained mainly from NWIS; additional spring locations were obtained from a file of springs that have water rights (B. Johnson, Washington State Department of Ecology, written commun., 2007). Most springs emanating from the bedrock units are probably fault-controlled, although some may be fracture-controlled. In the former case, the faults form lateral flow barriers but provide a conduit for vertical movement of groundwater. Springs also may be due to the truncation of a water-bearing unit, such as a basalt interflow zone in a stream-cut valley, or springs may be derived from precipitation recharge that discharges in seeps downslope from the recharge source (as described by Smith [1901] for many springs on the south slope of Rattlesnake Hills). A well upgradient of a spring may or may not be a flowing well.

Measured or estimated discharge for 207 of the 362 springs (fig. 20*B*) ranged from 0.25 to 2,400 gal/min, and averaged 76 gal/min with a median of 8 gal/min. The two largest springs are in the basalt units, as are most springs flowing at more than about 3–5 gal/min. The largest spring, the Clerf Spring (17N/20E-06A01S), discharges from the Grande Ronde near folds and a mapped fault (Smith, 1901), and had a measured discharge of 2,400 gal/min in 1953. Rattlesnake Springs (12N/25E-29B01S) discharge from the Saddle Mountains, and had a measured discharge of 585 gal/min in 1907, when it was used to irrigate 100 acres on the Benson ranch (Waring, 1913). Measured flow at these springs averaged about 209 gal/min over the period October 1990 to May 1993 (Dinicola, 1997), suggesting potential effects of upgradient pumpage on reduced spring flow. Spring flows of this magnitude discharge from the CRBG units throughout the Yakima Fold Belt, for example, in Klickitat County.

Although semiconfined to confined conditions generally exist below the water-table system, there are occurrences of perched groundwater zones. For example, in the Red Mountain-Badger Mountain area in eastern Benton County (the southern extension of the Rattlesnake Hills structure), most domestic and irrigation wells in the uplands are deeper wells and water levels (mapped by Drost and others, 1997) are much lower than land surface. However, there also are interspersed shallow wells with much higher water levels. The small number of these wells suggests a locally perched water table in this area. Similar situations are also noted throughout the study area, especially in parts of the Rattlesnake Hills, Rattlesnake Mountain, Ahtanum-Moxee basin, Selah basin, and in areas outside of the structural basins in Kittitas County, particularly to the east.

Flow System

Groundwater moves through the aquifer system from the uplands (high land-surface altitude—topographic highs) to surface drainage features in the lowlands, principally to the Naches and Yakima Rivers and to the Columbia River in the eastern part of the extended study area. Groundwater movement is affected by topography, geologic structure, natural recharge, discharge locations, hydraulic characteristics, recharge from the use of water (principally surface-water irrigation), and groundwater pumpage.

In the upper part of the flow system, groundwater levels are a subdued replica of the land surface. High-conductivity hydrogeologic units have water-table surfaces that are the most subdued replica of land surface; conversely, in a fine-grained unit (such as the Touchet Beds or low-permeability granite), water-table surfaces more closely mimic topographic variations. Departures from this general relation between water-level patterns and grain size occur where the water table is deep, which is related to the quantity of recharge. That is, where the quantity of recharge is small, for example less than 1 to 2 in/yr, the water table generally is deep and ‘flat,’ especially where recharge is less than about 0.5 in/yr.

The effect of recharge on groundwater movement decreases with distance from the outcrop or recharge area and with depth because movement of groundwater is affected by the depth of burial of a unit. Thus, flow in the deeper parts of the system (such as in Grande Ronde in the central Toppenish basin) is controlled primarily by hydraulic characteristics of the unit itself, the thickness and hydraulic characteristics of overlying units, and by regional discharge locations. Therefore, vertical leakage between hydrogeologic units is an important component of flow in the aquifer system. Flow in the deeper parts of the Grande Ronde also is affected by the topography of the underlying older Tertiary units. This topography, which would be expressed in the thickness of the Grande Ronde, is largely unknown, however, because only a few wells penetrate the unit.

As a result of the factors described above, the movement of groundwater to discharge locations can be on the order of hours to days in the shallow system to millennia in the deeper system, as is indicated by the large variations in groundwater ages estimated from carbon isotopes and concentrations of excess helium data.

Typical of groundwater flow systems, most of the movement of groundwater occurs in the upper part of the Yakima River basin aquifer system and decreases with depth within the system. This is due largely to the effect of recharge being derived from surface-water delivery and application in the lowlands and from precipitation in the humid uplands. For example, Hendry and others (1992) show the presence of tritiated water beneath WIP in the alluvial aquifer and the underlying aquifer unit contained in the Ellensburg Formation in the Toppenish basin, whereas water samples from the deeper CRBG were nontritiated. In addition, the δD and $\delta^{18}O$ values of surface water and shallow groundwater are similar due to the effects of the large recharge quantities in the surface-water irrigated areas but the values are different from those in water from deeper wells. Excluding the humid uplands, most of the flow or activity in the system occurs in the upper 100–300 ft of the system, whereas much less water moves in the more deeply buried parts of the system. Flow in the upper part of the system dominates the seasonal and annual water budget (groundwater flow). Overlain on this active part of the flow system is the larger regional flow system. Thus, an understanding of the water budget and movement of water in this upper part is needed to obtain a reliable water budget for the deeper part of the system because the water budget for the upper part of the system overwhelms the water budget of the deeper part.

Except for groundwater flow in the deeply-buried parts of the system, large-scale structural control compartmentalizes the flow system (Kinnison and Sceva, 1963; Hansen and others, 1994; Bauer and Hansen, 2000). The compartmentalization limits the length of the flow paths, resulting in relatively short paths for such a large aquifer system. Structural control is exerted primarily by the major ridges in the basin (figs. 7 and 8) such as, from north to south, Naneum Ridge, Ainsley Canyon anticline, Manastash Ridge, Umtanum Ridge, Cleman Mountain, Cowiche Mountain, Yakima Ridge, Rattlesnake Hills-Mountain, Ahtanum Ridge, Toppenish Ridge, Snipes Mountain, and the Horse Heaven Hills structure (Kinnison and Sceva, 1963; Newcomb, 1970; Hansen and others, 1994; Reidel and others, 2002; Jones and others, 2006). In addition, the Naneum Ridge anticline extends southward along the eastern boundary of the Yakima River basin to the Hog Ranch structure; together, they provide a north-south structural control on the flow system extending from Rattlesnake Hills to about the northern boundary of the basin. Combined, they also are called the Hog Ranch uplift (Smith, 1988) and the Hog Ranch-Naneum Ridge anticline. This structural control also is exerted locally. For example, Kirk and Mackie (1993) show the control of the Bird Canyon fault and the Hog Ranch-Naneum Ridge anticline on groundwater flow in the Saddle Mountains and Wanapum units in the Moxee-Black Rock valley. The structural/geologic history of the study area has been most recently summarized by Reidel and others (2002). As a result of this structural control throughout the system, down-valley groundwater flow in the basin-fill deposits in the structural basins generally is terminated with minimal underflow at all outlets except for the Eastern Benton basin, which does not have a structurally controlled outlet. Therefore, changes in recharge and discharge in five of the six structural basins principally are only promulgated down-valley through changes in streamflow leaving a basin.

Down-valley flow also is terminated in the upper parts of the bedrock and CRBG units in all but the Eastern Benton basin. Topographic and structural control also can lead to the termination of groundwater flow paths in the bedrock and CRBG units at greater depths. For example, Bauer and others (1985) and Hansen and others (1994) show groundwater flow paths in the Wanapum in the Yakima and extended Toppenish basins being directed towards the basin outlet and to the Yakima River, respectively. The control is represented by the topographic highs that are the major structural ridges, such as Ahtanum Ridge-Rattlesnake Hills and Toppenish Ridge. Topographic and structural controls on regional flow systems were described as early as 1940 by Hubbert (1940) and later by Toth (1963a, 1963b, 1978).

Groundwater in the basin-fill deposits in the Roslyn, Kittitas, and Yakima basins that is not lost to evapotranspiration from the water table ultimately discharges to drains and streams upgradient from the basin outlets; only minor amounts of groundwater discharges down-valley as underflow in the alluvial units. For the Selah basin, some discharge occurs as underflow along the Naches and Yakima River outlets. In the extended Toppenish basin, groundwater discharges from the basin-fill deposits by evapotranspiration and to drains, tile drains, and streams. Due to the lack of basin-fill deposits at the basin's outlet, most of the groundwater in the basin-fill deposits of the extended Toppenish basin has been discharged by about Prosser, and groundwater underflow is minimal downstream from the approximate location in the Yakima River of the Spring (Snipes) Creek complex. This complex is a relic drainage that functions as a drain that receives both surface water (spill and return flow) and groundwater (return flow) from irrigation.

Groundwater underflow mainly occurs in the thin alluvial deposits along the Yakima River near where it exits the basin. In the Eastern Benton basin, groundwater that is not lost to evapotranspiration discharges to the Yakima and Columbia Rivers and to drains and wasteways.

The water-level contours and generalized direction of groundwater flow in the surficial hydrogeologic units in the structural basins (pl. 3) and in the basalt units (pl. 4) highlight some of the above aspects. Actual water levels and flow directions are more complex than illustrated due to canal/lateral leakage, irrigation, drains, streams, pumpage, variations in recharge, spatially varying hydraulic characteristics, and topographic setting. Together, these factors locally affect the shape of the water-level contours and thus, flow directions and hydraulic gradients. These factors also affect the DTW, which locally is much more complex than illustrated on the generalized map (fig. 18). The water-level information described previously was used to construct these maps. Note that water-level information was most abundant for the upper part of the aquifer system and the number of measured water levels declined with depth.

The configuration of the water table in the surficial deposits in the Roslyn, Kittitas, and Yakima basins clearly shows groundwater moving towards the alluvial aquifer with streams and ultimately to the basins' outlets (pl. 3). This is especially true in the Kittitas and Yakima basins because of their shapes and narrow outlets with thin basin-fill deposits. For both the Kittitas and Yakima basins, major drains, such as Wilson and Cherry Creeks in the Kittitas basin and Wide Hollow Creek and Moxee drain in the Yakima basin, capture most of the shallow groundwater originating from surface-water irrigation. Although the Roslyn basin also has a narrow outlet, groundwater movement is toward the rivers and drains, rather than toward the basin outlet, throughout most of the basin because it is long and relatively narrow. Thus, most of the groundwater will have discharged to the Yakima River before it reaches the basin's outlet that abuts the Kittitas basin.

Groundwater flow also is directed to the outlet of the Selah basin, but underflow along the Naches River to the Toppenish basin is indicated by the contours (pl. 3); this underflow also was suggested by Kinnison and Sceva (1963). In addition, the contours show that the flow system is compartmentalized into three distinct flow systems—Wenas Creek in the northeast Selah basin, Naches River in the center, and Cowiche Creek (with additional compartmentalization between the North and South Forks in the Cowiche Creek drainage) due to topographic setting and geologic structure; such multi-compartmentalization will be described in more detail for the Eastern Benton basin. The contours and flow lines for the extended Toppenish basin show that a large part of the flow is directed towards the Yakima River by about Toppenish Ridge. After Toppenish Ridge, the basin becomes narrower and the flow contribution towards the river is similar to that in the Roslyn basin. Most of the remaining groundwater moves towards the Yakima River by Prosser and, on the basis of contributing area, only a small amount of the groundwater in the Toppenish basin moves in the surficial deposits between Prosser and the Spring (Snipes) Creek complex.

The flow system is more compartmentalized in the Eastern Benton basin than in the other basins (pl. 3) due to its location, topographic setting, and geologic structure. This area is bounded between Townships 8N and 9N and Ranges 27E and 28E, and includes the Rattlesnake Hills structure that is expressed as the uplands containing Red Mountain, Candy Mountain, and Badger Mountain. Except where the Saddle Mountains outcrops, these uplands are overlain by the fine-grained Touchet Beds. Groundwater moves both east and west in Badger Coulee from a groundwater divide that is east of Badger—westward toward the Yakima River near Benton City and eastward toward the Columbia River. Flow in the uplands moves towards, (1) Badger Coulee, (2) West Richland from the gap between Red and Candy Mountains, and (3) the Yakima River on the west. Except for the gap, the geologic structure in the uplands generally limits the northerly and northeasterly flow to the area north of the Red Mountain-Badger Mountain structure. The northerly flow from the uplands is bifurcated near West Richland with flow to the northeast towards the Columbia River and to the northwest to the Yakima River. Although few data are available, the southerly flow from the Rattlesnake Hills structure in the uplands may be redirected to the northwest by Goose Hills (an expression of an anticline and possible flow barrier in the southeast) and to the north and northeast by another topographic high in 8N/28E-17 that is also an anticline and close to a thrust fault. Groundwater also flows toward the Yakima River from the west and northwest from Rattlesnake Mountain from just north of Kiona/Benton City to about Horn Rapids; the river is incised in the Saddle Mountains in part of this reach. Drost and others (1997) concluded that the Saddle Mountains gains water from the Yakima River in part of this reach. Flow from about Horn Rapids to the the mouth of Yakima River is toward the Columbia River, and the water-level contours indicate that the Yakima River is losing water over part of this reach. This latter aspect is consistent with the findings of Brown (1979), Lindberg and Bond (1979), and Drost and others (1997).

Groundwater levels in the basalt units (pl. 4) generally parallel the land surface or, where a unit is buried, the dip of the basalt because most groundwater occurs and moves in the interflow zones. Where the units are deeply buried in the structural basins, water-level contours are smoother (lower hydraulic gradient) than those in the uplands, which are typically outcrop areas. The water-level contours mapped for the CRBG units are generalized due to sparse data in many locations and large variations in water levels with depth; two closely-spaced wells that are completed at different depths in a unit can have large differences (more than 200 ft) in water levels. The maps show the authors' representation of average water levels for a unit based on available data and hydrologic judgment. In areas of sparse data for a particular unit, water levels for an overlying unit, previously published maps (Bauer and others, 1985; Lane and Whiteman, 1989), and model-calculated groundwater levels (Hansen and others, 1994) were used to help guide the contouring of water levels to provide a more complete picture of the flow system. Thus, actual groundwater levels may be higher or lower than those estimated, especially in the Satus Creek basin and eastern Kittitas County. There are fewer Grande Ronde water-level contours compared to the Saddle Mountains and Wanapum due to the paucity of data, and the contours also are more generalized in most locations and primarily are representative of the upper part of the Grande Ronde. Spatial and vertical variations in hydraulic characteristics of both individual flows and interbedded sediments and the presence of geologic structure result in a much more complex flow system for each unit than is depicted by the water levels.

The groundwater levels for the CRBG units show topographic/structural control on the movement of water (pl. 4) similar to that inferred for flow in the basin-fill deposits. The Saddle Mountains flow system is distinctly compartmentalized and nearly isolated in the Selah and Yakima basins due to both the unit's extent and structural control. In the extended Toppenish basin, flow to about a north-south line (generally paralleling a line between Sunnyside and Mabton) is somewhat isolated from the Saddle Mountains flow system to the east. Similarly, most of the flow system in the Eastern Benton basin is isolated from the rest of the Saddle Mountains system by the Rattlesnake Hills structure and the Hog Ranch uplift. Flow in the Wanapum is similarly compartmentalized but with some connectivity between the Selah, Yakima, and Toppenish basins (pl. 4). Distinct isolation of the Wanapum flow system occurs in the Kittitas basin and in the areas east of the Hog Ranch uplift. Toppenish Ridge-Snipes Mountain also appears to be a general flow boundary for the Wanapum in the extended Toppenish basin, and the spatial extent of the flow system to the east is larger than that estimated for the Saddle Mountains. The boundary for this part of the flow system follows increased altitude and "flattening" of the top of the Wanapum to the east (Jones and Vaccaro, 2008) that approximately parallels the Sunnyside-Mabton boundary in the Saddle Mountains. However, there is a more distinct flow boundary in the Saddle Mountains compared to that in the Wanapum in the extended Toppenish basin. For example, for the 800-ft water-level contour, the Saddle Mountains contour along this boundary extends for about 10 mi to the southeast, whereas the Wanapum contour extends to the southeast for about 6 mi (pl. 4). For both units, there does not appear to be a large-scale regional flow system, which reflects the influence of the structural control. However, depending on the control of Snipes Mountain and its potential buried extension with depth, flow paths may be longer in the deeper parts of the Wanapum. The data are insufficient to determine this, however, and water-level contours were conservatively mapped to indicate compartmentalization. Generally, where both the Saddle Mountains and Wanapum are present, there is a reasonable correspondence in their flow systems. For example, groundwater levels on the north flank of Ahtanum Ridge-Rattlesnake Hills anticline in the Moxee subbasin show upgradient flexure at Konnowac Pass; the structural low at Konnowac Pass is where the ancestral Yakima River crossed the anticline (Smith, 1988).

The contours for the Grande Ronde are representative of the upper part of the unit because very few water wells penetrate more than 1,000 ft of the unit in the basin. Groundwater flow in the Grande Ronde also is compartmentalized (pl. 4) but not to the same extent as in either the Saddle Mountains or Wanapum. The large spatial extent of the Grande Ronde results in a large flow system with more interconnections than in the other two CRBG units, for example, in eastern Kittitas County and from the Selah basin to the Yakima basin. Where the Grande Ronde outcrops, the water-level contours mimic land-surface topography and they become a more subdued replica of topography as the unit becomes buried. In the more deeply buried parts of the unit, the contours are smoother than those for the other CRBG units. Similarly, water-level contours near geologic structure in the eastern part of the area are more subdued and smoother. The flow system in the Grande Ronde is controlled by the regional discharge locations along the Yakima and Columbia Rivers; that is, the regional flow (hydraulic head) in the Grande Ronde tends to the level of the major streams. There may be a regional flow system in the deeper part of the Grande Ronde but there are insufficient data to verify the presence of such a system. However, results of regional groundwater flow models (Hansen and others, 1994; Packard and others, 1996) indicate the presence of a larger-scale flow system in the Grande Ronde that is recharged in its outcrop area to the west. There also may be flow contribution to the Grande Ronde from the underlying bedrock units.

The water-level contours also show the lateral hydraulic gradient in the study area. For the water table (pl. 3), the hydraulic gradient ranges from about 7 ft/mi in the Toppenish basin to more than 400 ft/mi in the Roslyn, Selah, and Eastern Benton basins. Close spacing of contours (steep hydraulic gradients) is indicative of areas with units having low K_h and (or) steep terrain. K_h generally is large where the spacing is further apart (low hydraulic gradients), such as in the central part of the Toppenish basin. In this area, some of the surficial units consist of coarse-grained deposits with K_h values on the order of 800 ft/d (Bolke and Skrivan, 1981). The steep gradients typically occur in steep terrain near narrow river valleys. Except for parts of the Eastern Benton basin, gradients are low along the major stream channels. For example, in the Roslyn and Kittitas basins the gradients are as small as 12 and 17 ft/mi, respectively, along the channels. Low gradients, about 15–17 ft/mi, also are present along the lower Naches River and Wenas Creek valleys. In the Yakima basin, hydraulic gradients are as small as 21 ft/mi along the Yakima River. In the Eastern Benton basin, most gradients range from about 70 to 200 ft/mi. To the north and east of the Yakima River to the Columbia River, especially north of Richland on the

Hanford Site, hydraulic gradients are only about 9 ft/mi, which corresponds to flatter gradients in areas of deep water tables. In the central section of the eastern part of Badger Coulee, there is an area where the gradient is about 17 ft/mi, which is probably due to the decreasing thickness of the Touchet Beds and the large thickness, about 100 ft, of the coarse-grained Pasco gravels (Drost and others, 1997; Jones and others, 2006). For comparison, the lateral hydraulic gradient for the consolidated units (pl. 1) was estimated to range from about 15 to 50 ft/mi.

The lateral hydraulic gradient in the CRBG units also is highly variable (pl. 4). Similar to those in the water-table system, the steepest gradients are in areas of steep topography and the lowest gradients occur along the synclines in the topographically smooth parts of the structural basins. For example, the slope of the water-level contours for the Saddle Mountains and the slope of the land surface are generally within 5 degrees of each other over about 85 percent of the extent of the unit; this type of similarity was described by Toth (1978). This relation between hydraulic gradient and topographic slope also holds for the Wanapum and Grande Ronde. An overview of basalt lateral gradients is described on the basis of the Saddle Mountains water-level contours, and the gradients are generally representative of the other basalt units. The lowest hydraulic gradients occur in the Toppenish basin where they range from about 3 to 14 ft/mi. Low gradients, on the order of 7–10 ft/mi, also occur in parts of the Eastern Benton basin. Moderate gradients, 25–40 ft/mi, occur along the synclinal valley in the Yakima basin. In contrast, the lowest Saddle Mountains gradient in the valleys in the Selah basin is more than 50 ft/mi. In gently sloping uplands, such as in parts of the Satus Creek basin and parts of the Rattlesnake Hills, hydraulic gradients can be as small as 30 ft/mi. In steeper and more complex terrain, gradients are more than 125 ft/mi (such as along parts of the Rattlesnake Hills) and in some areas they are more than 800 ft/mi (such as along ridges in parts of the Selah basin). Thus, lateral hydraulic gradients tend to be similar to the overall topographic gradient, with the steepest gradients generally in zones associated with anticlines and (or) faults, and the smallest gradients in the flatter structural basins. The latter also suggests that where the basalt flow system meets the low-gradient flow system in the basin-fill deposits, groundwater in the basalt moves into those deposits, because, although the basalt hydraulic conductivity is larger in the structural basins than in the steep-gradient areas, somewhat steeper hydraulic gradients would otherwise be present in the basalts underlying the basin-fill deposits.

Groundwater moving along local-to-regional flow paths of varying lengths has both lateral and vertical flow components in a three-dimensional flow system. Depending on location (lateral and vertical), the movement of water may be predominantly downward, upward, or lateral. In large flow systems, flow direction is predominantly downward in the uplands, lateral in a transition zone, and upward in regional discharge locations. These three zones also are referred to as the zone of recharge, zone of lateral flow, and zone of discharge, respectively (Berger, 2000). Superimposed on these generalized large-scale patterns are complex variations. For example, for the bedrock units and Wanapum and Grande Ronde, high hydraulic heads in the wetter uplands compared to heads at lower altitudes suggest predominantly downward flow. However, large potential recharge quantities, limiting infiltration rates, and (or) lower K_h and K_v values in some areas also produce a large lateral-flow component in the upper part of the flow system. The lateral flow component follows local to intermediate flow paths to streams, and it can support as much as 50 percent of the streamflow during certain periods, especially in areas floored by bedrock units (Mastin and Vaccaro, 2002). In areas of regional groundwater discharge, which are typically along the major streams and near the outlets of the structural basins, flow directions are upward in the deeply buried confined units and transition to predominantly lateral in the upper part of the surficial water-table unit. For example, lateral flow is indicated in the surficial units in the extended Toppenish basin (pl. 3), whereas a flowing, 1,020-ft deep well near the Yakima River (10N/20E-04L01, completed in Unit 5 of the Toppenish basin) is in an area of upward flow—the lateral component of groundwater flow is minimal in this part of the system below the river.

The direction and magnitude of the vertical component of flow thus varies spatially and with depth. On a regional basis, vertical flow is downward in the uplands and transitions to upward near the boundaries of the structural basins. The downward component can be large; for example, an 830-ft downward head difference was noted in a 3,200-ft section of the Grande Ronde unit in a test well drilled on the crest of Rattlesnake Mountain (Raymond and Tillson, 1968). This large-scale pattern is similar to that described by Hansen and others (1994). Locally, there are complex variations in the direction and magnitude of vertical flow due to geologic structure and the spatial variations in K_v and K_h values recharge, and hydrogeologic-unit composition. In turn, the vertical hydraulic gradient that drives the vertical flow integrates these variations.

To improve the understanding of vertical gradients for the aquifer system as a whole, non-pumping water-level measurements were made, where possible, in two closely spaced wells with different depths. Seventy-three pairs of wells were identified for analysis of vertical gradients. The vertical distance over which the water-level differences were assumed to be occurring was based on the well depths because the wells were all open at the bottom, which is representative of the maximum opening depth. The difference in depths for the well pairs averaged 396 ft (median of 240 ft), with 90 percent of the values between 53 and 975 ft; the maximum depth difference was 2,094 ft. The differences in depth were significantly correlated (correlation coefficient 0.95) to the deeper well of the well pair. That is, deeper wells were associated with larger differences in depth, and thus, gradients were calculated over larger vertical distances. The median water-level difference for the pairs was about 35 ft and ranged from -174 (negative values indicate upward flow) to 485 ft;

the median of the absolute water-level difference was 80 ft. Vertical hydraulic gradients (in units of ft/ft) ranged from -1.45 to 3.28 and the median was 0.16 (median of 0.32 for absolute values) with a standard deviation of 0.71 (0.63 for absolute values). Eighteen of the pairs had negative gradients (upward flow), 6 pairs had gradients that were near neutral (generally because the errors of the measurements were greater than the calculated gradients and (or) both wells were in either a zone of lateral flow or a transition zone), and the remaining pairs had positive gradients (downward flow). Considering that (1) the well pairs were widely distributed, representative of numerous hydrogeologic units (including CRBG units), and in locations ranging from recharge areas to regional discharge areas, and (2) the difference in depths of the pairs of wells was as much as 2,094 ft, the consistency of vertical hydraulic gradients (fully 75 percent of the absolute values were less than 0.51) was unexpected. Most of the variation in vertical gradients occurred for differences in depths for pairs of wells that were less than 201 ft (median gradient of 0.36 for 34 absolute values) compared to opening-depth differences greater than 200 ft (median gradient of 0.30 for 39 absolute values); the range in gradients for the less than 201-ft category was more than 4 times the range for the greater than 200-ft category. These differences suggest that (1) the shallow part of the flow system is more active than the deeper part, and (2) the deeper system approaches equilibrium, probably due to base level water-level altitudes in the regional discharge areas and minimal effects from recharge. Depth differences also were significantly related (at the 0.01 significance level) to the vertical hydraulic gradient and the slope of the relation was negative—the deeper in the system the higher the probability of upward vertical gradients. Whiteman and others (1994) present basalt water-level hydrographs from piezometers in the Columbia Plateau aquifer system that show vertical hydraulic gradients ranging from 0.2 to 0.54. Packard and others (1996) estimated vertical gradients of 0.14 in the Wanapum along the axis of Horse Heaven Hills and identified previous work that calculated values of 0.18 for the Saddle Mountains in the same area, further suggesting that the vertical gradients in the CRBG units fall in a relatively narrow range. In one area in Texas for the Gulf Coast aquifer system, Williamson and Grubb (2001) report vertical gradients of 0.15 and 0.16. Vaccaro and others (1999) report vertical gradients ranging from 0.01 to 0.36 (average of 0.30) for the Puget Sound aquifer system; the lower value was for coarse-grained, alluvial valley water-table systems. Woodward and others (1998) report vertical gradients ranging from 0.02 to 0.13 for the Willamette Lowland aquifer system. Together, this information suggests that vertical gradients approach some equilibrium value that may be independent of a particular regional flow system, regardless of location within the system. Variations to the above aspect are found in confined units with high artesian pressure, further indicating that the magnitude of the vertical gradients is a function of the degree of confinement.

On the basis of water levels noted on driller's logs, data from piezometers, and logging of wells, water-level changes with depth can be abrupt for basalt wells, and indicate complex variations in vertical gradients. Packard and others (1996) describe and present examples of this for the Horse Heaven Hills. As an example of this phenomenon, the largest vertical gradient was observed in the USGS Medicine Valley observation well (11N/16E-15K01P1 and -15K01P3, both completed in the Grande Ronde). P3 is open at a depth of 525 ft and its water level is about 12 ft below land surface, and P1 is open at 674 ft and its water level is about 500 ft below land surface—indicating an abrupt water-level change due to a low-permeability layer between 525 and 674 ft. In addition, water levels in P1 have declined more than 50 ft since 1972, whereas water levels in P3 have declined by only about 3 ft. It is of interest that there is no pumping upgradient of the observation well and the nearest groundwater irrigation is several miles away, indicating the lower zone (in which P1 is completed) is well connected to a downgradient pumping zone and that the effects of the pumping are likely due to changes in storage. Gregg and Laird (1975) describe how this deeper water-producing zone is continuous under the structure (an anticline) that separates Medicine Valley (a syncline) from the Wapato syncline to the east. The shallower P3-zone has water levels that are similar to those in the unconsolidated deposits, and this zone is not continuous through the ridge (Gregg and Laird, 1975). As another example of abrupt changes in water levels with depth, during drilling of two deep wells (12N/17E-16R01, 1,075-ft deep, and 11N/16E-25Q01, 1,100-ft deep) water levels initially indicated downward flow, with an abrupt transition to upward flow at depth. At well 16R01, the flow was downward to 788 ft, reversed abruptly to upward, and then the well flowed from 851 to 1,075 ft. Similarly, the depth to water was 71 ft at 242 ft for 25Q01, declined to 214 ft at 880 ft, and rose to 64 ft at 1,020 ft. Well 9N/23E-26M01 (960-ft deep) had downward flow to 829 ft that reversed to upward flow at 858 ft. Also, in some sections in 10N/23E, various basalt wells have cascading water, indicating a large downward gradient and low K_h value for the flow interiors.

When groundwater moves in the three-dimensional flow system from recharge areas to discharge areas, the water moves between hydrogeologic units, and there are both losses and gains of water to a unit. On a long-term basis and under natural conditions, the losses from and gains to a unit would be approximately the same because the groundwater flow system is approximately in equilibrium. Under current conditions, groundwater pumpage and increases in recharge have changed both the quantities and directions in the movement of water between all units (Skrivan, 1987; Hansen and others, 1994; Bauer and Hansen, 2000). Water-level hydrographs for wells 12N/21E-16L01 (1,390 ft deep, cased to 1,017 ft, completed in Wanapum) and 12N/21E-16N02 (704 ft deep, open to 640–704 ft, completed in Saddle Mountains) (fig. 21*B*) in the upper Moxee subbasin illustrate how these changes occur due to pumpage. From about 1979 to 1981, differences in water levels between these two wells were about 275 ft, with the deeper well (12N/21E-16L01) having higher water levels, indicating an upward vertical gradient. The water level in 16L01 during this period was more than 500 ft above the top of the Wanapum. By 1997, that difference had been reduced to about 94 ft—a reduction in the vertical gradient from about -0.88 to -0.30. As a result, the upward movement of groundwater in this area may have decreased by about 76 percent between 1979 and 1997. If the downward trend in water levels of about 11.6 ft/yr has continued in 16L01, the water-level difference in 2000 may have been less than 60 ft and by 2005 there may have been a reversal in flow direction, from upward to downward. However, water levels in nearby wells completed in the Saddle Mountains have declined similar to levels in the Wanapum in well 16L01, and the vertical gradients remain upward. Thus, a reversal to downward vertical gradients may be confined to this particular area.

Water-level data and flow-system analyses indicate that the movement of water between and within basalt flows decreases with increasing age of the basalt due to overburden pressure (Hansen and others, 1994; Reidel and others, 2002) and secondary mineralization; the latter was documented by geochemical information (Ames, 1980; Benson and Teague, 1982; Hearn and others, 1985; Steinkampf and Hearn, 1996). As a result, vertical movement of groundwater between basalt flows in the Saddle Mountains and Wanapum units generally is less impeded than the movement between either Grande Ronde flows or between the Wanapum and Grande Ronde units. Thus, lateral water movement in the Grande Ronde, where it is deeply buried, is less than the other basalt units due to fracture infilling from secondary mineralization and overburden pressure; this aspect is especially true where there is large thickness of the Grande Ronde such as in the Eastern Benton basin. Where the Grande Ronde outcrops, it functions similar to the other basalt units in their outcrop areas. Depending on their hydraulic characteristics, interbeds in the basalts can cause either increased or decreased movement of groundwater between basalt flows. In some areas (for example, in 10N/23E in the Rattlesnake Hills), some high-yield irrigation wells obtain their water from the interbeds because locally they are more productive than the CRBG units. In other areas, for example in 13N/24E, the Mabton Interbed is composed of shale and sandstone and acts as a confining unit.

No data are available on the vertical movement of water in the bedrock units. However, on the basis of information for similar rock types and of similar ages elsewhere, the vertical movement would be greatly diminished with depth (Hollyday and Hileman, 1996). For example, in parts of Kittitas County with a large potential recharge rate, depths to water in many wells range from 150 ft to more than 400 ft, suggesting lower K_v values because large potential recharge rates are generally associated with a shallow water table.

Groundwater Use

Vaccaro and Sumioka (2006) estimated groundwater pumpage from the aquifer system for the period 1960–2000 for eight categories of use, and they estimated that more than 40,000 water wells withdrew about 312,284 acre-ft (about 430 ft^3/s) in 2000 for multiple uses and averaged about 234,000 acre-ft (about 325 ft^3/s) during that 41-year period; these quantities exclude estimates for standby/reserve irrigation pumpage. The 8 categories of use were: (1) municipal, including two larger Washington State Department of Health Group A systems (Terrace Heights and Nob Hill), and some systems operated by Yakima County, (2) irrigation, (3) commercial and industrial, (4) livestock, (5) fish and wildlife, (6) non-municipal Group A and B systems, (7) domestic, and (8) groundwater claims. Group A systems generally have 15 or more service connections and Group B systems generally have 2 to 14 connections. The first 5 categories (especially category 2) account for most of the 2,874 active groundwater rights in the study area; categories 6 and 7 also include some groundwater rights. The irrigation category includes standby/reserve rights. These rights were obtained by surface-water users in irrigation districts with proratable surface-water rights, and the wells would be pumped during prorating years. The total allowable quantities for the groundwater rights (certificates and permits) as of 2001 were (R. Dixon, Washington State Department of Ecology, written, commun., 2001):

	Number of certificates and permits	Entitlement		Irrigated area (acres)
		Instantaneous (gal/min)	Annual (acre-ft)	
Certificates	2,575	720,683	422,040	101,371
Permits	299	230,623	107,191	28,199

The estimates of pumpage for 1960–2000 (table 5; Vaccaro and Sumioka, 2006) show that the irrigation category accounts for about 60 percent of the total pumpage without standby/reserve estimates. However, it was estimated that about 32,430 acre-ft of the pumpage for the groundwater claims category (95 percent of the total) was for irrigating about 12,800 acres, indicating that about 70 percent of the 312,284 acre-ft of pumpage in 2000 was for irrigation. The information in table 5 indicates that the pumpage in 2000 for the non-irrigation categories falls within two general ranges, values less than 9,369 acre-ft and values ranging from 20,036 to 37,272 acre-ft. The total for the Public Water Supply (including municipal) and domestic categories was about 66,421 acre-ft or about 21 percent of the total (table 5).

Table 5. Summary of estimated annual pumpage, by pumpage category in 5-year increments, Yakima River basin, Washington, 1960–2000.

[Data from Vaccaro and Sumioka (2006). **Public water supply:** Pre-1960 includes some pumpage for 1960 because initial estimates were calculated for 1960 for Group A and B systems, and domestic. **Irrigation Standby/reserve:** Estimates based on pumpage that occurred in all years and is not the actual pumpage; pumpage occurred only in 1977, 1979, 1987, 1988, 1992, 1993, and 1994. **Groundwater claims:** About 95 percent of pumpage for claims is for irrigation. **Cumulative:** cumulative is total pumpage at end of time period and does not include standby/reserve estimates. Total pumpage for each category and for all categories combined is for 2000. Estimates based on pumpage that occurred in all years and are not the actual pumpage; estimates for total annual pumpage existing prior to 1960 are also shown. All values in acre-feet]

Years	Public water supply			Domestic	Irrigation			Livestock	Commercial and industrial	Fish and wildlife	Ground-water claims	Total for all categories	Cumulative without standby/reserve
	Municipal	Group A systems	Group B systems		Total	Primary	Standby/reserve						
Pre-1960	19,127	3,888	858	12,379	42,000	41,896	104	222	3,093	3	34,310	115,880	115,776
1960–64	628	331	73	1,187	9,299	9,299	0	208	438	0	0	12,164	127,940
1965–69	2,702	358	79	716	20,271	20,249	22	39	1,428	0	0	25,593	153,511
1970–74	1,379	389	85	988	25,010	23,743	1,267	3,189	612	2	0	31,654	183,898
1975–79	4,236	422	93	2,914	99,474	49,399	50,075	84	44	58	0	107,325	241,148
1980–84	975	458	101	-107	28,845	17,507	11,338	996	1,442	2,661	0	35,371	265,181
1985–89	2,614	586	110	305	20,687	14,927	5,760	461	170	0	0	24,933	284,354
1990–94	1,793	449	119	1,175	21,821	10,398	11,423	1,527	3	4,194	0	31,081	304,012
1995–2000	3,819	584	129	479	3,635	812	2,823	0	0	2,451	0	11,097	312,286
Cumulative through 2000	37,273	7,465	1,647	20,036	271,042	188,230	82,812	6,726	7,230	9,369	34,310	395,098	312,286

The standby/reserve irrigation pumpage was estimated by Vaccaro and Sumioka (2006) as if every year was an extreme drought year (that is, how much would be pumped if there were no surface water available). These estimates were intended to provide information for long-term planning related to how much withdrawals could occur in the basin. Thus, the estimates represent what could have been pumped for an existing crop type but not what was actually pumped; these values are reported in table 5 and indicate the total potential pumpage for irrigation for these rights. The prorating years with standby/reserve pumpage were 1977, 1979, 1987, 1988, 1992, 1993, 1994, 2001, and 2005: the prorating level for 1973 (the first year of prorating) was 80 percent, and standby/reserve pumpage was not implemented to any extent (J. Kirk, Washington State Department of Ecology, oral commun., 2005). Note that only estimates of standby/reserve pumpage through 2001 are being used in other work elements of this study. Based on the methods outlined by Vaccaro and Sumioka (2006), standby/reserve pumpage was estimated to have ranged from about 9,200 acre-ft in 1988 to 52,100 acre-ft in 1994 and 2001, and total standby/reserve pumpage for the 8 prorating years through 2001 was about 250,350 acre-ft.

All pumpage values except standby/reserve reported by Vaccaro and Sumioka (2006) for 2000 and, where available, for 2001, were aggregated by one-quarter township-range in order to provide a map that shows the spatial distribution of most of the groundwater pumpage in the basin. It was assumed that for wells with no estimates for 2001, the pumpage in 2001 was similar to that in 2000. The estimated domestic pumpage for each census block was aggregated to the quarter township-range based on the location of the centroid of a census block. In addition, the standby/reserve pumpage for 2001 (the 52,100 acre-ft described above) also was added. Thus, the map of aggregated pumpage (fig. 22) presents an estimated distribution of pumpage during the major drought year of 2001. The spatial scale of the aggregation did not allow for showing all of the estimated pumpage on the map due to source-water security concerns.

Most of the pumpage in 2001 was in the lower part of the basin (fig. 22), especially in Yakima County, which is also the location of most of the groundwater rights, groundwater claims, and population. The least pumpage occurred in Kittitas County, the least populous county in the study area and the county with the smallest number of groundwater rights. The pumpage occurs from all of the aquifers in the aquifer system. About 48 percent of the pumpage was from the units composed of basin-fill sediments, 50 percent from the CRBG units, and 2 percent from the bedrock units; these percentages include pumpage in the extended study area.

The cumulative pumpage for the 1960–2001 base period is of interest for comparison with historical groundwater levels. The cumulative pumpage from the system was estimated to be about 10 million acre-ft and its spatial distribution (fig. 23) generally is similar to that of 2001 pumpage; differences in the spatial distribution are due to the timing of the growth in pumpage. Areas where the cumulative pumpage was large and groundwater levels were stable indicate large quantities of recharge and (or) a large groundwater source (contributing) area. In contrast, areas where the cumulative pumpage was large and groundwater levels declined indicate limited quantities of recharge and (or) a small source area. These relations will be discussed in more detail in the following section.

Although the information shown in figures 22 and 23 indicates the magnitude of pumpage, it does not show the total allowable pumpage—the total entitlements as of 2001 shown previously. The unused, allowable water-right pumpage (including that from exempt wells) and the requested pumpage in existing (as of 2001) applications for new groundwater rights are nearly twice the estimated pumpage in 2001. Requested instantaneous rates for applications totaled about 613,306 gal/min and the applications were for a total irrigated acreage of 64,308 acres.

In comparison to groundwater use, the surface-water use in the basin, as measured by diversions, is much greater. This aspect of water use is important for improving the understanding of the relation between the effects of the flow and diversion of surface water and groundwater pumpage on the aquifer system. Mean annual discharge and time-series of monthly discharge for surface-water diversions were estimated on the basis of data obtained from several sources; however, the mean annual values from the different sources are not precisely comparable (see appendix D). Note that time series of monthly estimates are needed for other parts of this study, in particular for the groundwater flow models. The estimated mean annual discharge for surface-water diversions, which are principally from the Yakima and Naches Rivers, shows the magnitude of the diversions (fig. 24). Most of the diverted surface water is accounted for by 65 diversions (fig. 24, Reclamation-based information; appendix D) and to a lesser extent by the diversions that were aggregated to 78 points of withdrawal (fig. 24, WaDOE- and YN-based information; appendix D). The distribution of diversions throughout the basin reflects the history of water development in the basin because almost all of the diversions were established before 1920.

Mean annual discharge (for water years 1960–2001) for the 65 diversions was about 5,860 ft^3/s (or about 4.2 million acre-ft), and values for individual diversions ranged from about 1 to 1,031 ft^3/s. About 1,950 ft^3/s of this diverted water was returned as power-generation returns (Roza, Wapatox, and Chandler power plant returns), yielding about 3,910 ft^3/s for other uses (principally irrigation). The total for the other 78 points of withdrawals was 474 ft^3/s (or about 0.3 million acre-ft—about equivalent to the groundwater pumpage in 2001), and the values ranged from about 0.1 to 39 ft^3/s. Together, the non-power quantities total 4,383 ft^3/s, or about 10 times greater than the estimated pumpage in 2000. About 69 percent of the diverted water was for irrigation, 31 percent was for power generation, and less than 1 percent was for municipal supply.

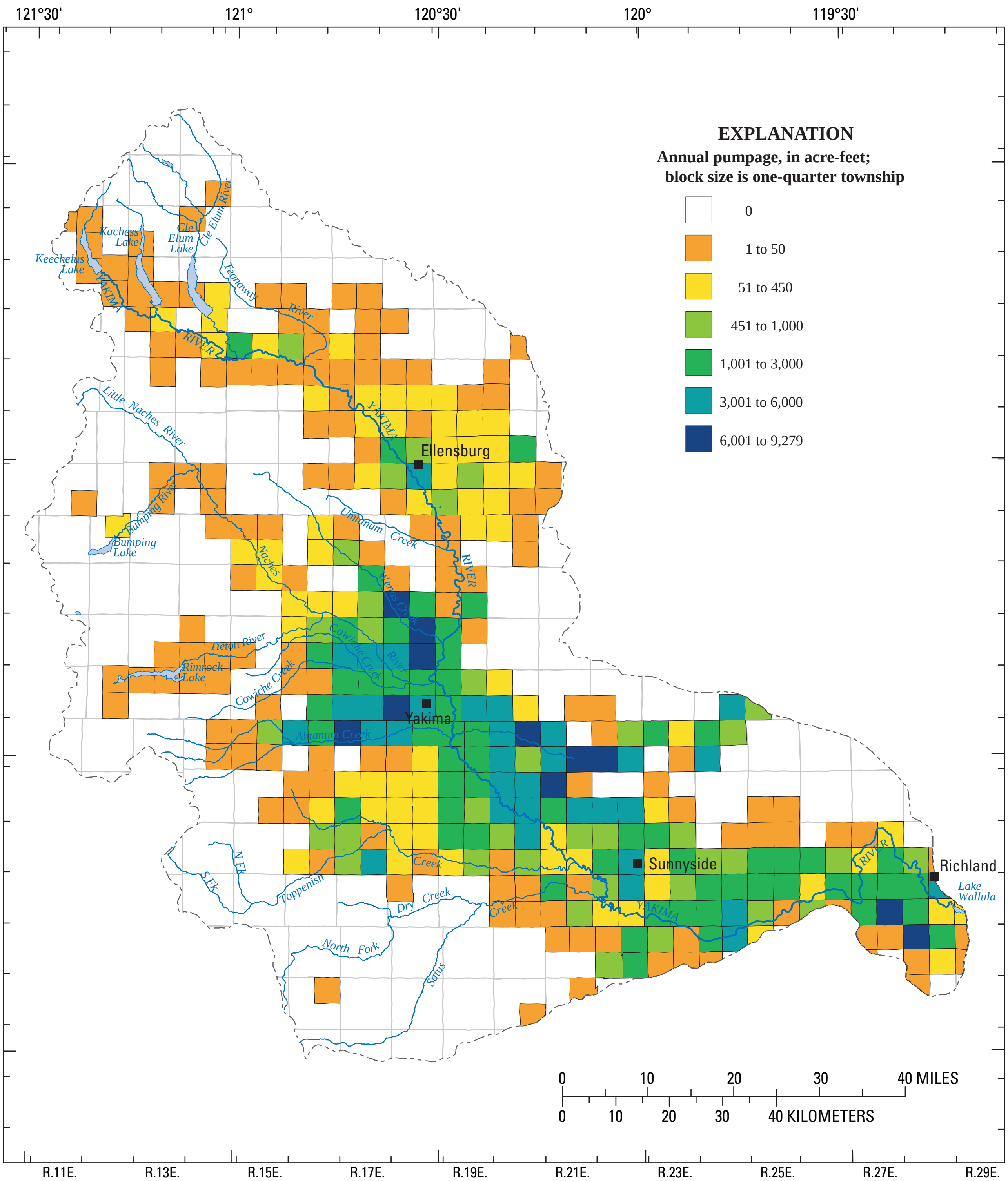


Figure 22. Spatial distribution of pumpage from the Yakima River basin aquifer system, Washington, 2001.

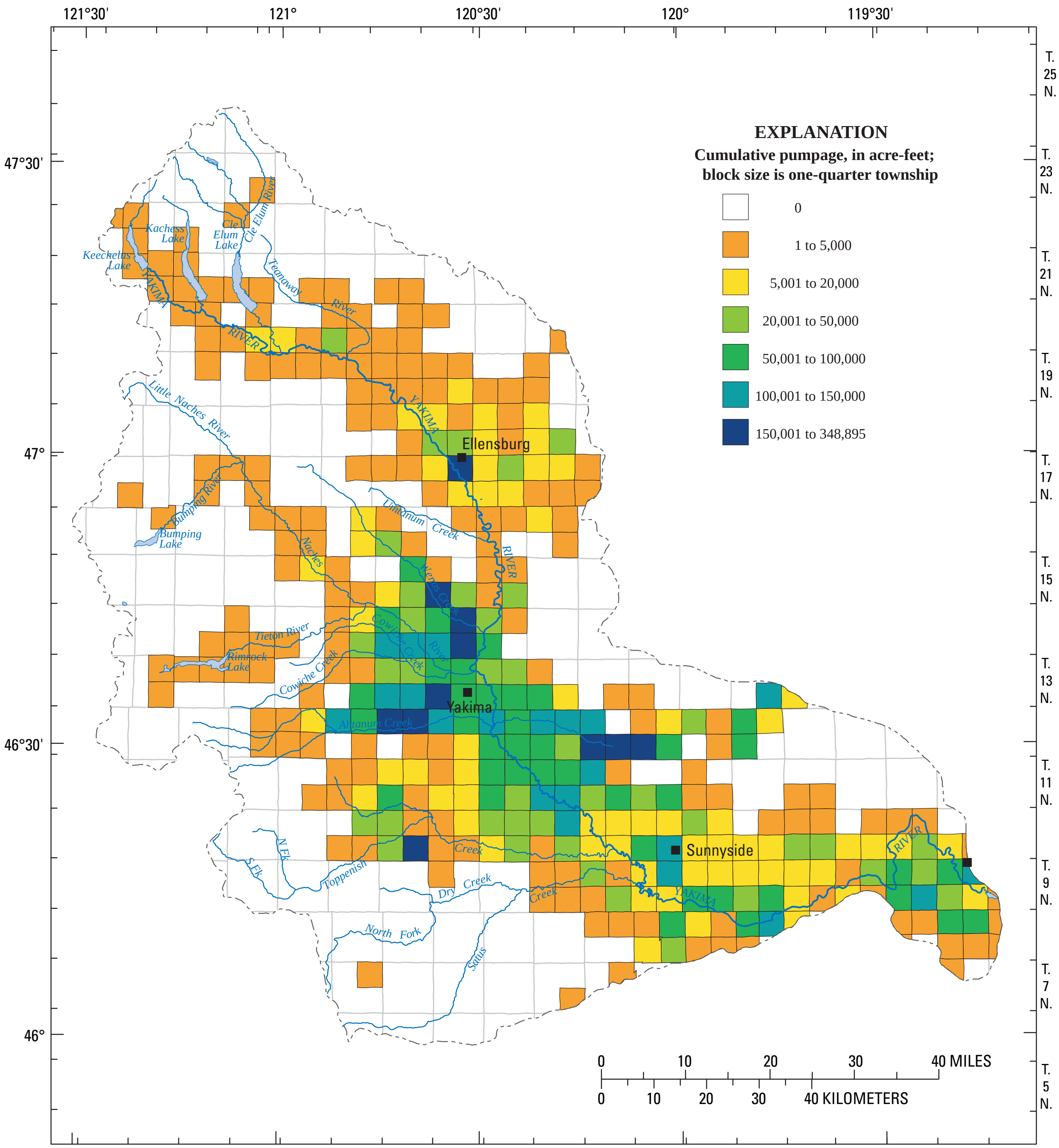


Figure 23. Spatial distribution of cumulative pumpage from the Yakima River basin aquifer system, Washington, 1960–2001.

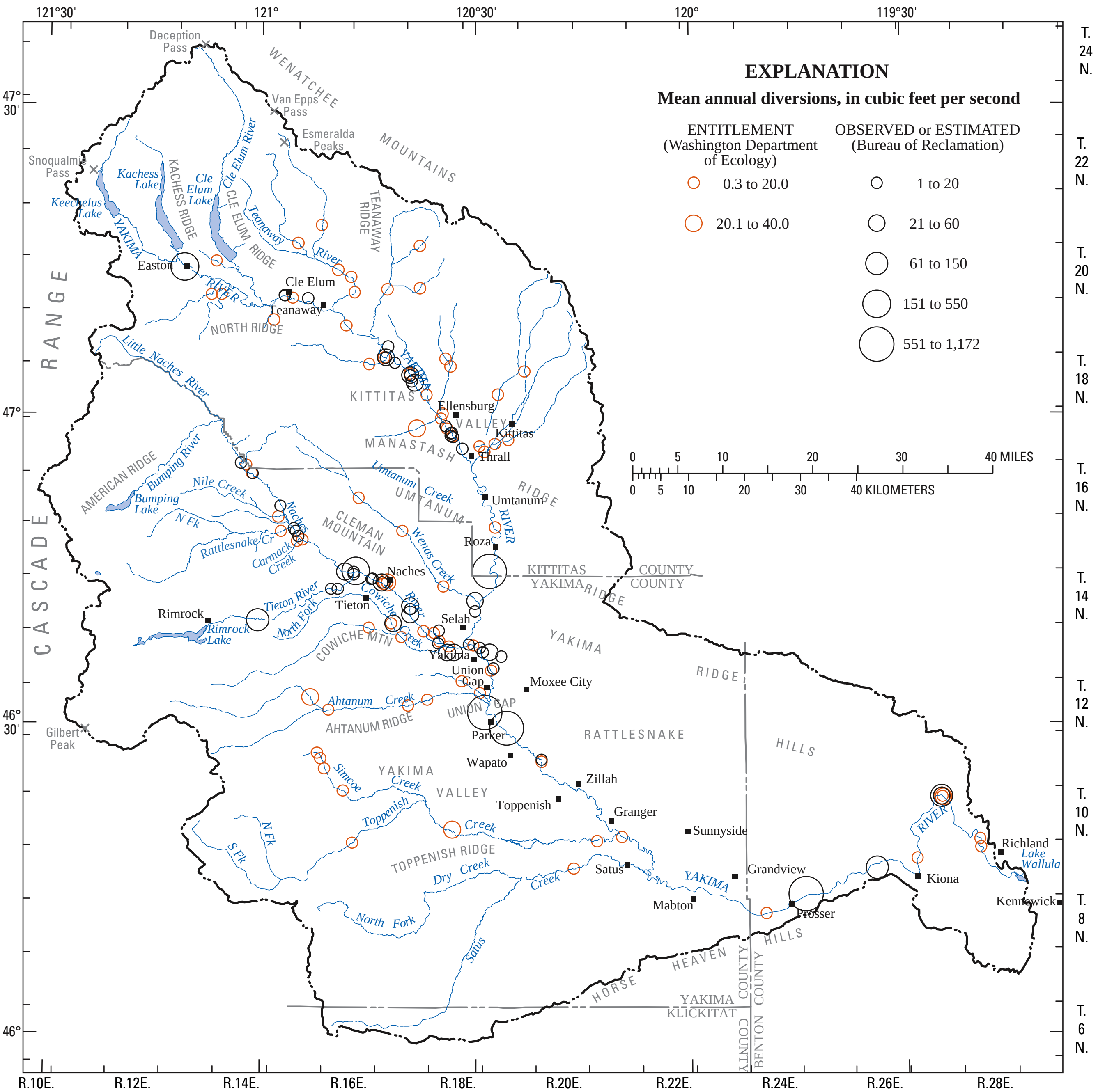


Figure 24. Location and estimated mean annual quantity of surface-water diversions in the Yakima River basin, Washington.

Water-Level Trends

Trends in groundwater levels are one of several factors important to understanding groundwater availability. Trends, and absence of trends, when analyzed in conjunction with groundwater pumpage, streamflow, and recharge, can indicate areas where there have been: (1) changes in groundwater storage, (2) potential capture of recharge, and (3) changes in groundwater discharge to streams, springs, or wetlands. Pumpage (discharge by wells) must be balanced by one or a combination of the above three changes (Theis, 1940).

In order to assess trends, water-level data stored in NWIS were analyzed. These data include water levels entered into the database during this study that were obtained from older reports and field sheets, and measurements made by

(1) the USGS between 2000 and 2002, (2) the WaDOE between 1986 and 2002, (3) the YN between 1993 and 2002, and (4) Reclamation between 2000 and 2002. A total of about 300 wells were identified for which some water-level measurements that spanned at least a 10-year period. Most wells selected for trend analysis were measured in the mid-1990s to 2000–02 (the latter period is when personnel from Reclamation, USGS, WaDOE, and YN measured water levels); this selection yielded more than 220 wells. A few wells with measurements that did not span 10 years also were included in the analysis for areas or units with limited historical data. The lengths of the records vary and, thus, the long-term trends may be more or less than the trend indicated in the measurement period. The selected wells have a reasonable spatial distribution and are a subset of wells that met the criteria. The hydrographs for the final subset of wells were published as an interactive World Wide Web report (Keys and others, 2008).

Groundwater-level trends were categorized as: stable to small water-level declines (0–20 ft), moderate declines (21–75 ft), large declines (76–150 ft), and very large declines (greater than 150 ft). Each category does not include wells completed in all of the water-producing hydrogeologic units because there was either a paucity of long-term data or there were no declines of a particular magnitude for a unit. A few wells exhibited upward trends in water levels and these are discussed separately. Where data permitted, areas and units were identified that had water-level trends for these categories and maps were constructed. The maps were constructed conservatively in that the mapped extent for any particular category of declines was kept to a minimum. The maps show the long-term spatial trends in groundwater levels. Example hydrographs for each category are presented and discussed below.

Stable to small declines in water levels (0–20 ft) are exemplified by four wells (fig. 25*A-D*) in areas with small rates of pumpage, large recharge rates, and (or) large upgradient source areas. Well 9N/21E-16D01 (fig. 25*A*) is completed in Unit 2 in the extended Toppenish basin and nearby well 9N/21E-26M01 (fig. 25*B*) is completed in the Saddle Mountains and Wanapum. The small decline for 9N/21E-16D01 is associated with large quantities of recharge (the well is in WIP). In contrast, the small decline in 9N/21E-26M01 is associated with small rates of pumpage in this area from these units, combined with a large, upgradient groundwater source area. If the declining water level for 26M01 continued through 2008 (assuming the pumpage remained the same), the total decline would still likely be less than 20 ft. Well 11N/19E-10A02 is completed in Units 2 and 3 in the Toppenish basin and the small decline (fig. 25*C*) is associated with low rates of pumpage from these units and a large groundwater source area in this location of upward flow. Finally, well 17N/20E-29R01 (completed in Unit 3 of the Kittitas basin) shows little or no decline with a distinct spring-fall difference in water levels (fig. 25*D*). Pumpage rates are small in this part of Unit 3 in this area.

Moderate declining water levels (21–75 ft) are exemplified by four wells in areas with low to moderate rates of pumpage, moderate to large recharge rates, and (or) large upgradient source areas (fig. 26*A-D*). Well 10N/16E-15N01 (fig. 26*A*) is completed in the Grande Ronde in the Toppenish basin, and declines are moderated because pumpage rates in this unit are not as large as in some other areas/units (for example, pumpage from the Wanapum and Saddle Mountains to the east and north of this area in the western Toppenish basin), the unit receives recharge in its large outcrop area (fig. 8, pl. 2), and its upgradient area receives recharge quantities of more than 10 in/yr. However, nearby large downward trends are noted in both the Wanapum and Saddle Mountains. The moderate decline for well 10N/21E-03H01 (fig. 26*B*; completed in the Saddle Mountains) is due to low rates of nearby pumpage. A moderate decline also is indicated by the hydrograph for Wanapum well 13N/17E-04D02 in the Selah basin (fig. 26*C*). Nearby pumpage irrigates about 400 acres of lands, but the overall small quantity of withdrawals in this area and the size of the upgradient source area moderates the declines. Similarly, declines in well 14N/18E-19L01 (fig. 26*D*; completed in the Wanapum) also are moderated due to its large source area and small rates of nearby pumpage.

Large downward trends in water levels (76-150 ft) (fig. 27*A-E*) have been observed primarily in deep CRBG wells in areas with large rates of pumpage, including standby/reserve pumpage. For example, water levels in well 10N/17E-27Q01 (open to all three CRBG units) display a continual downward trend over a 43-year period—a decline of about 2.9 ft/yr (fig. 27*A*). In a 1,100-ft deep Wanapum-Grande Ronde composite well (11N/16E-25Q01), water levels also show a steady decline that averages about 2.7 ft/yr (fig. 27*B*); a seasonal hydrograph for this well (inset on fig. 27*B*) shows the steady decline after each irrigation season. Water-level declines in Saddle Mountains-Wanapum composite wells 12N/18E-32H01 and -32L01 (fig. 27*C*) indicate a consistent and nearly linear large decline in this area. A Wanapum well (14N/17E-04H02, 1,000 ft deep) experienced a 94-ft decline over a 49-year period—about 1.9 ft/yr (fig. 27*D*). However, most of the decline started in about 1987–88 during a 2-year drought period. Hydrographs from other wells also display an onset of declines during the 1987–88 drought period, especially CRBG wells in areas with standby/reserve irrigation pumpage (see Keys and others [2008] for hydrographs with similar onsets of declines). Effects of droughts on water levels due to both increased pumpage and reduced recharge are well documented. For example, of the 460 wells that were measured during this study in both the spring of 2001 (a major drought year) and the spring of 2002, about 66 percent had water-level declines during this period. In addition to deep CRBG wells, water levels in well 20N/15E-28Q04 (completed in the confined aquifer-Unit 3 in the Roslyn basin) display a large downward trend due to pumpage (fig. 27*E*). This large decline occurred because limited recharge reaches the unit through either overlying

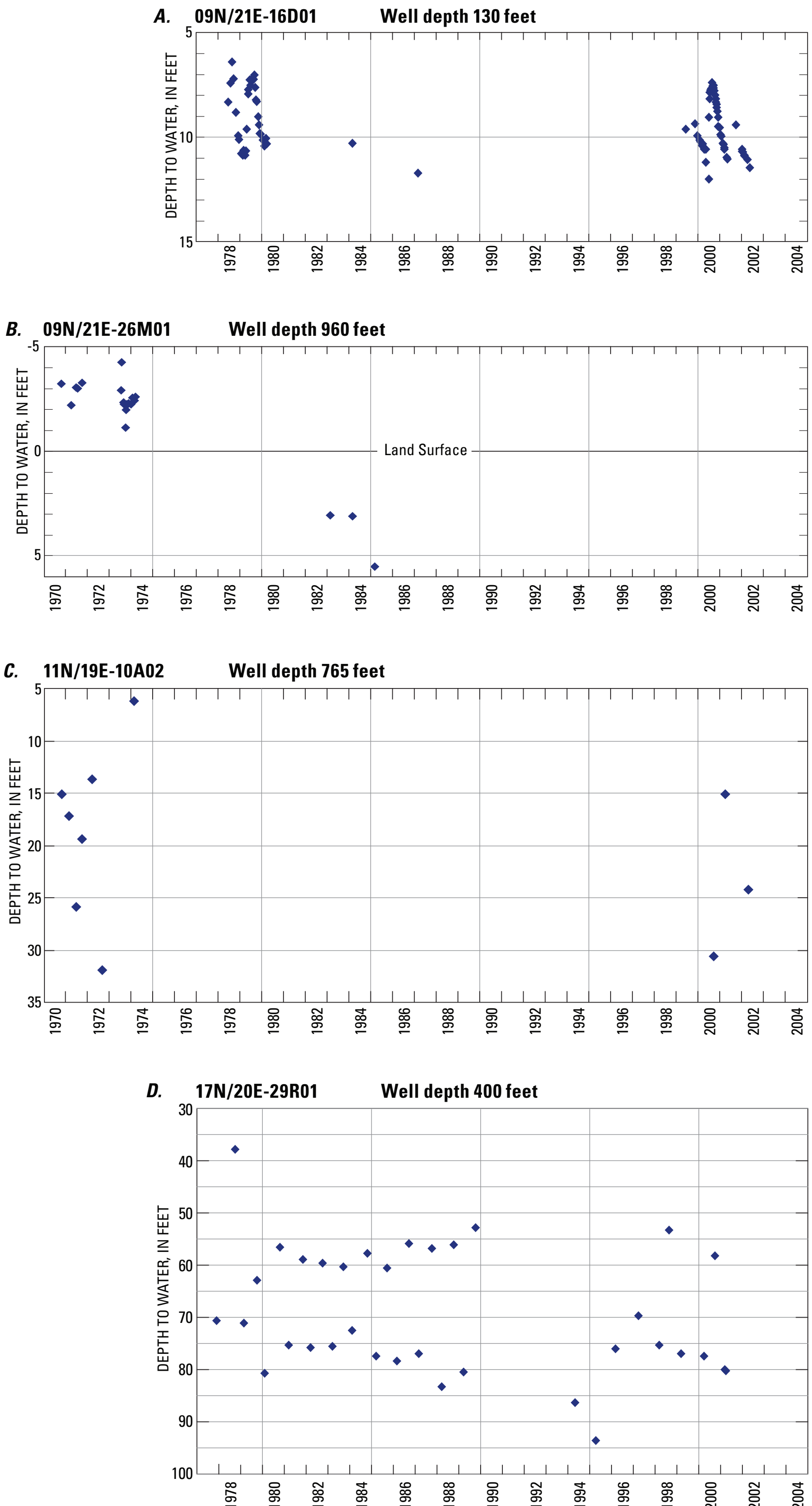


Figure 25. Examples of groundwater level hydrographs with stable or small declines, Yakima River basin, Washington.

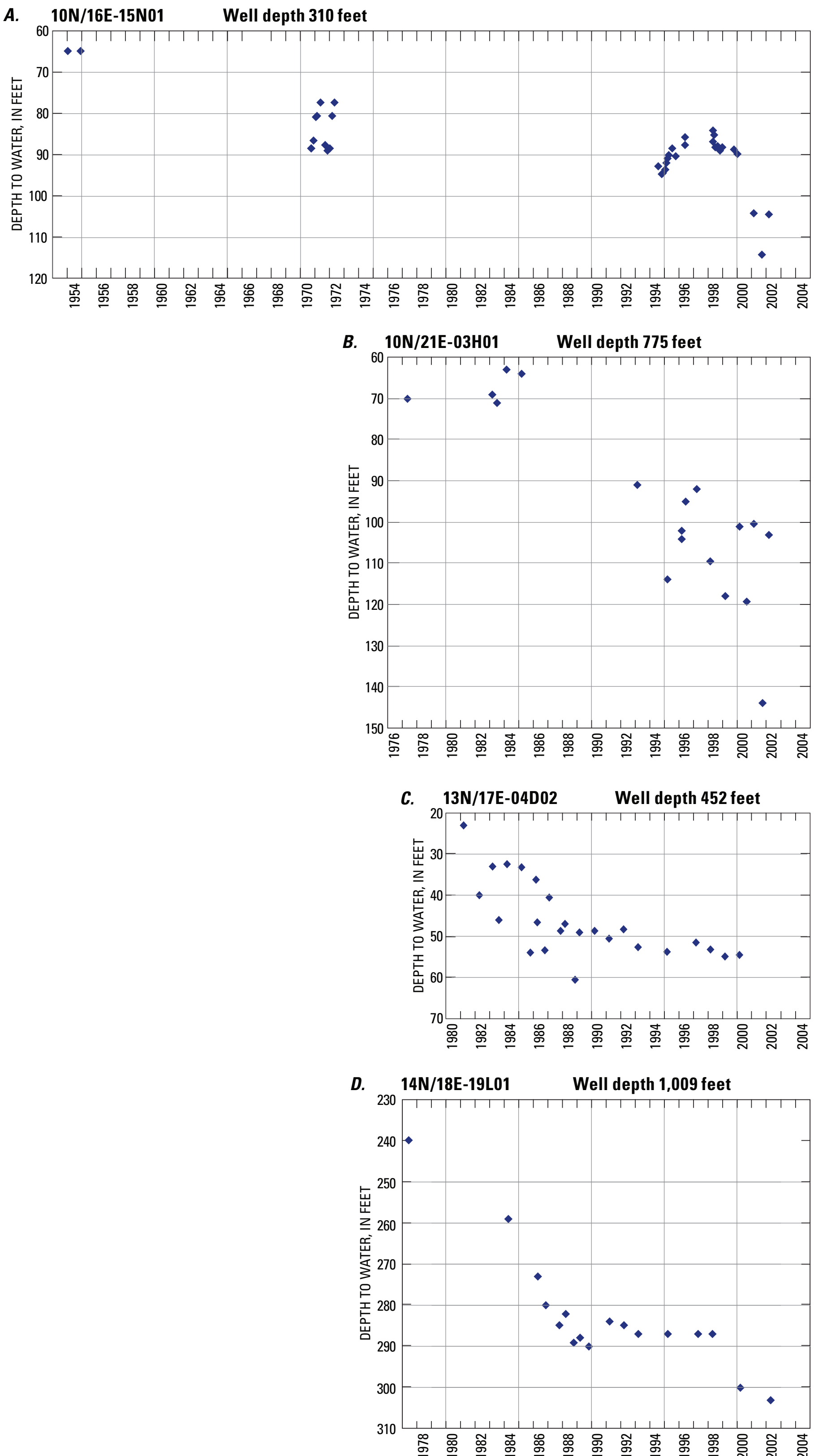


Figure 26. Examples of groundwater level hydrographs with moderate declines, Yakima River basin, Washington.

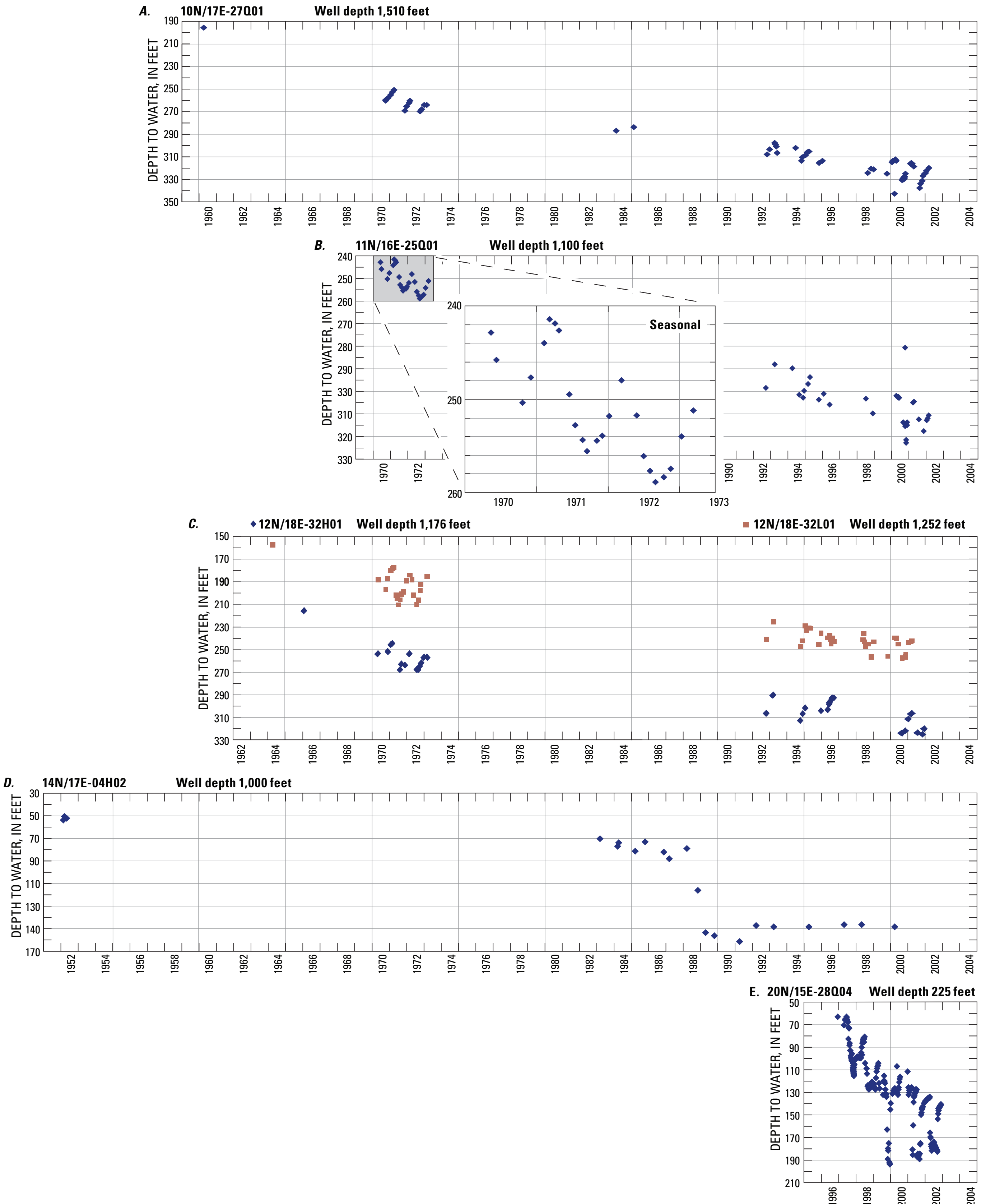


Figure 27. Examples of groundwater level hydrographs with large declines, Yakima River basin, Washington.

units (general flow direction is upward) or underlying low-conductivity bedrock units (the Roslyn Formation)—the pumpage from the wells in this particular area is larger than the amount of water that can replenish the unit. That is, there is no water to capture from rejected recharge and very little water to capture from groundwater discharge, so that pumpage is met by changes in storage.

Very large downward trends in water levels (greater than 150 ft) (fig. 28*A-D*) have mainly been observed in deep wells in areas of intensive pumpage, limited recharge, and (or) small upgradient source areas. Similar to wells with large water-level declines, these deep wells are completed in the CRBG units. The 1,505-ft deep well 11N/20E-01E01 (a Saddle Mountains-Wanapum composite well) displays about a 150-ft decline over a 15-year period or about 10 ft/yr (fig. 28*A*). If the average rate of decline continued through 2008, the long-term decline would have exceeded 300 ft. Very large declines are displayed by the hydrographs for Wanapum wells 12N/22E-21J01 and -21R01 (1,140- and 903-ft deep, respectively) in the upper Moxee subbasin (fig. 28*B*). These wells are in an area of intensive groundwater pumpage and very small rates of recharge. Similarly, water levels for well 12N/24E-03B01 (a 1,239-ft deep Wanapum well) in the upper Cold Creek drainage display very large declines (fig. 28*C*). Water levels in other deep irrigation wells in the upper Cold Creek drainage also have had very large declines. As a last example, water levels for well 13N/24E-27M01 (a 723-ft deep Wanapum well) in the lower Cold Creek drainage basin have declined about 200 ft over about 35 years, or about 5.5 ft/yr (fig. 28*D*).

In the upper Moxee subbasin, and the Black Rock, Dry, and Cold Creek drainages, the Yakima Ridge and Rattlesnake Hills exert structural control on the flow system to limit the area of recharge to the groundwater basin. This area is further separated in an east-west direction by the Hog Ranch structure (fig. 8). Thus, groundwater flow from the north and south into this area is limited (except perhaps in the deeply buried parts of the Grande Ronde), and flow between the Moxee subbasin and the other drainages is limited by the structures (flow barriers that act as a groundwater divide). Historical groundwater levels in the basalts east of the Hog Ranch structure were typically 100–200 ft lower than levels west of the structure; currently, the water levels to the east are about 200–300 ft lower due to the effects of pumpage. In addition, the rate of water-level decline east of the structure (with less total pumpage) is much more than west of the structure. Recharge also is very low in these areas (fig. 16), and it must move downward through thick basin-fill deposits and (or) overlying interbeds and basalt flows. In addition, some of the recharge entering the upper unit(s) moves laterally and is not available to replenish the deeper units. As a result, the pumpage rates are greater than the replenishment rates and there have been large changes in groundwater storage. This also is an area for which Hansen and others (1994) and Bauer and Hansen (2000), using a groundwater flow model, estimated water-level declines on the order of 100–200 ft from predevelopment conditions to 1985 and projected 300–400 ft declines from 1985 to long-term equilibrium conditions under 1985 pumpage (observed declines since 1985 have ranged from 100 ft to more than 200 ft).

Rising water levels were observed in very few of the wells. The levels in both Saddle Mountains and Wanapum wells in Badger Coulee in the Benton basin had long-term rises. In addition, water levels in two of the selected wells completed in the Pasco gravels (toward the eastern end of Badger Coulee) also had long-term rises. Rises throughout Badger Coulee due to surface-water irrigation have been described by Drost and others (1997). Unlike other surface-water irrigated areas in the basin, rises are still occurring because irrigation started as late as 1957 in the Kennewick Irrigation District and 1978 in the Badger Mountain Irrigation District. In comparison, irrigation started in 1893 in the Columbia Irrigation District. Water-level rises in the other surface-water irrigated areas have reached equilibrium and the total rise for those areas will be assessed in a subsequent part of this study. Water levels for well 13N/18E-24K03 (a 1,683-ft artesian well drilled in about 1906 in an area of upward vertical gradient—a Saddle Mountains or Unit 3/Saddle Mountains composite well) in the City of Yakima showed a rise of about 10 ft from about 1981 through March 1993. However, an August 2000 water level was about 26 ft lower than the March 1993 reading, and a measured shut-in pressure of 25 lb/in^2 made while conducting a field test for a temperature profile for the well (the actual date is unknown but thought to be about 1982) indicated a water level that was about 16 ft lower than the March 1993 reading. These differences may be due to seasonal variations, use of the water, location of the measurement line, and variations in pressure-gage accuracies—an upward trend would not be expected for a well completed at this depth and location. Moderate rises in water levels (about 65 ft) were observed from 1983 through 1991 in a 543-ft deep basalt well (16N/20E-07Q01). This well is just upgradient of an irrigation canal and the rises may be due to leakage around the borehole. Water levels from 1991 through 2001 in this well declined during the drought years and rose to the 1991 level during other years. Another well (19N/18E-21B01, a 700-ft deep Grande Ronde well) had a small water-level rise of about 7 ft from 1983 through 2001; the reason for the rise is unknown.

Trends in groundwater levels show distinct spatial patterns. Rising water levels have been observed primarily in the Badger Coulee area in the Benton basin. Areas with declining water levels generally are limited to Yakima and Benton Counties. There also are distinct differences in water-level trends in the basin-fill units compared to those in the basalt units (no long-term data are available for the bedrock units). Based on available long-term water-level measurements for 55 wells completed in basin-fill units, only 2 areas were not categorized as stable (less than 21 ft of decline). Although continuous long-term data are not available, water-level declines have also occurred in the Ellensburg Formation in the upper Moxee subbasin; these declines are discussed below.

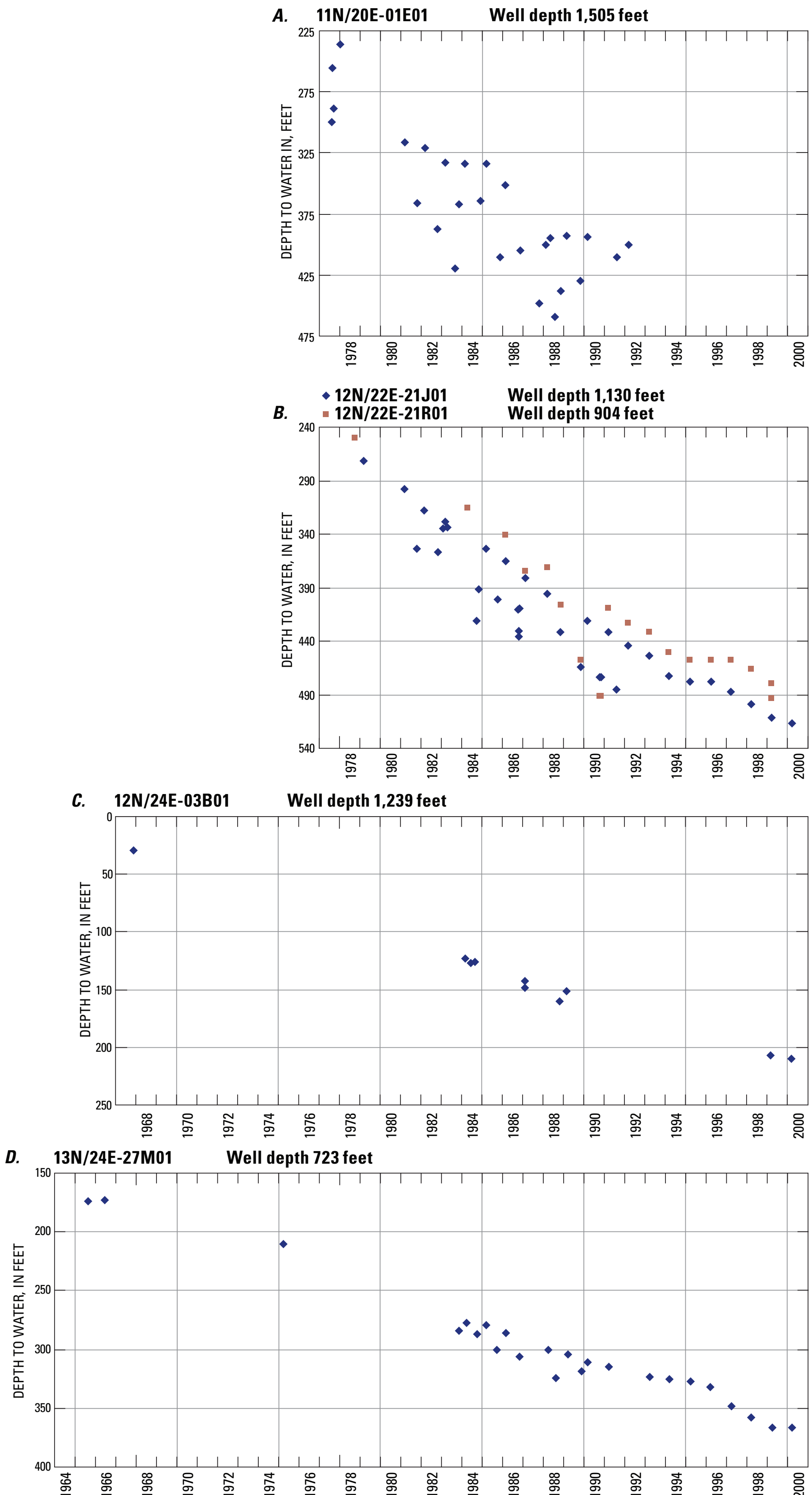


Figure 28. Examples of groundwater level hydrographs with very large declines, Yakima River basin, Washington.

One of the areas with declines in water levels in the basin-fill units is the Roslyn basin, where moderate to large declines have occurred in parts of the confined Unit 3. Unit 3 in this area receives water principally from the underlying Roslyn Formation (an Eocene-age coal-rich formation—part of the Tertiary sediment bedrock unit) that is relatively 'tight,' and current pumpage in this part of Unit 3 is apparently larger than inflow from the Roslyn Formation and any capture of discharge. The only other basin-fill well with long-term data that had a moderate decline in water level was a 1,044-ft-deep well (13N/18E-35K01) finished in the lower part of the consolidated unit in the Yakima basin. The water level in this flowing well declined about 40 ft from 1961 to 2001. The water level in a nearby deep well (13N/18E-35D04 – 1,171 ft) was categorized as stable, with a decline of less than 10 ft from 1995 through 2001. The total decline for this well may have been more than 10 ft, but measurements were not available from 1961 to 1995 to compare with those for the 1,044-ft well. This example highlights the differences in water-level trends based on the length of water-level records. The lower part of the consolidated unit in this area (Yakima basin) receives most of its water from the underlying Saddle Mountains.

In the upper Moxee subbasin, more than 30 deep wells were drilled from the 1890s through 1901 (Smith, 1901). All of these wells tapped a flowing artesian zone in the Ellensburg Formation (the consolidated unit in the Yakima basin, see table 2). Smith reported that several of the most eastward of the wells had stopped flowing by 1901. A 1982 WaDOE log for the well that Smith called Clark No. 2 (13N/20E-31L01, 1,026-ft deep) reported the depth to water as 30 ft and the water level in 1897 was more than 100 ft above land surface. This indicates that there have been large declines in water levels in the lower part of the consolidated unit within the upper Moxee subbasin. A few of these early wells are still reported to be flowing, but they are the more downgradient wells.

Total irrigation pumpage (the largest use) from the basin-fill deposits in 2001 was about 68,000 acre-ft (94 ft^3/s). A map of irrigation wells completed in these deposits (fig. 29) indicates they are widely distributed in Yakima and Benton Counties with only a few in Kittitas County. Predevelopment recharge in the structural basins in Yakima and Benton Counties was estimated to be generally less than 3 in/yr, with most of the basins receiving less than 0.5 in/yr of recharge; the estimate of total mean annual predevelopment recharge in these structural basins was about 63 ft^3/s. However, current-condition recharge was estimated to be on the order of 20 in/yr in the surface-water irrigated areas that overlie most of the deposits in the structural basins (fig. 16). The predevelopment recharge would not have been sufficient to meet about a third of the historical 2001 pumpage demand without changes in either storage and (or) groundwater discharge.

Groundwater levels in the basin-fill deposits generally are stable, indicating that the current-condition recharge is sufficient to prevent storage changes. However, part of the pumpage demand is met by capture of some of the current-condition recharge, which suggests a reduction in groundwater discharge. For example, in the Ahtanum subbasin, Vaccaro and Olsen (2007a) estimated mean annual predevelopment and current-condition recharge to be 2.7 in. (31 ft^3/s) and 11.6 in. (134 ft^3/s), respectively. Irrigation pumpage from the basin-fill deposits in this subbasin was estimated to be about 20,374 acre-ft (28 ft^3/s) in 2001 (about 58 percent of the total irrigation pumpage in the subbasin in 2001), and other basin-fill pumpage accounts for about 6,320 acre-ft (8.7 ft^3/s). Pumpage was thus slightly greater (5.7 ft^3/s) than predevelopment recharge, and adjustments would have occurred in the basin-fill flow system (such as decreased storage) under existing pumpage and predevelopment recharge quantities. Thus, pumpage demand is being partly met by the increased recharge under current conditions because no changes in storage have occurred. Part of the demand appears to be met by decreasing groundwater discharge to streams, as evidenced by records of the flow of Ahtanum Creek at Union Gap (USGS stream-gaging station 12502500). This station has a statistically significant (significance level of 0.003) downward trend (slope of -1.2 ft^3/s per year) in discharge during the August–November period for calendar years 1962–2002. This trend is not related to climate because the rivers without human impacts, such as the American River and the Teanaway River below the Forks, do not show this trend. This suggests that the pumpage demand that is partly met by additional recharge is decreasing the amount of groundwater discharge to Ahtanum Creek during its baseflow period.

Most of the downward trends in groundwater levels in the Yakima River basin are in the CRBG units. Maps showing the generalized trends in water levels (location and magnitude of declines) in these units indicate that declines have occurred in distinct areas (figs. 30–32). Declines greater than 20 ft have the largest areal extent for the Saddle Mountains, about 177 mi^2—about 8 percent of the unit (fig. 30). Assuming declines of 21 to 42 ft over this area based on the distribution of declines and a storage coefficient of 0.04, the loss in groundwater storage due to the water-level declines in this unit is conservatively estimated to be from 24,000 to 48,000 acre-ft. The largest area of decline is on the south flank of Rattlesnake Hills, followed by that in the Moxee subbasin. Other declines also have occurred in the west-northwest part of the Toppenish basin, and near Red Mountain and Richland in the Benton basin. The largest declines (greater than 150 ft) have principally occurred in parts of the Rattlesnake Hills, Moxee, and Cold Creek areas. Declines greater than 20 ft also have occurred in the Wanapum (fig. 31) but their areal extent (83 mi^2—about 2 percent of the unit) is not as large as those in the Saddle Mountains; probable declines of more than 20 ft have occurred in an additional area of about 74 mi^2.

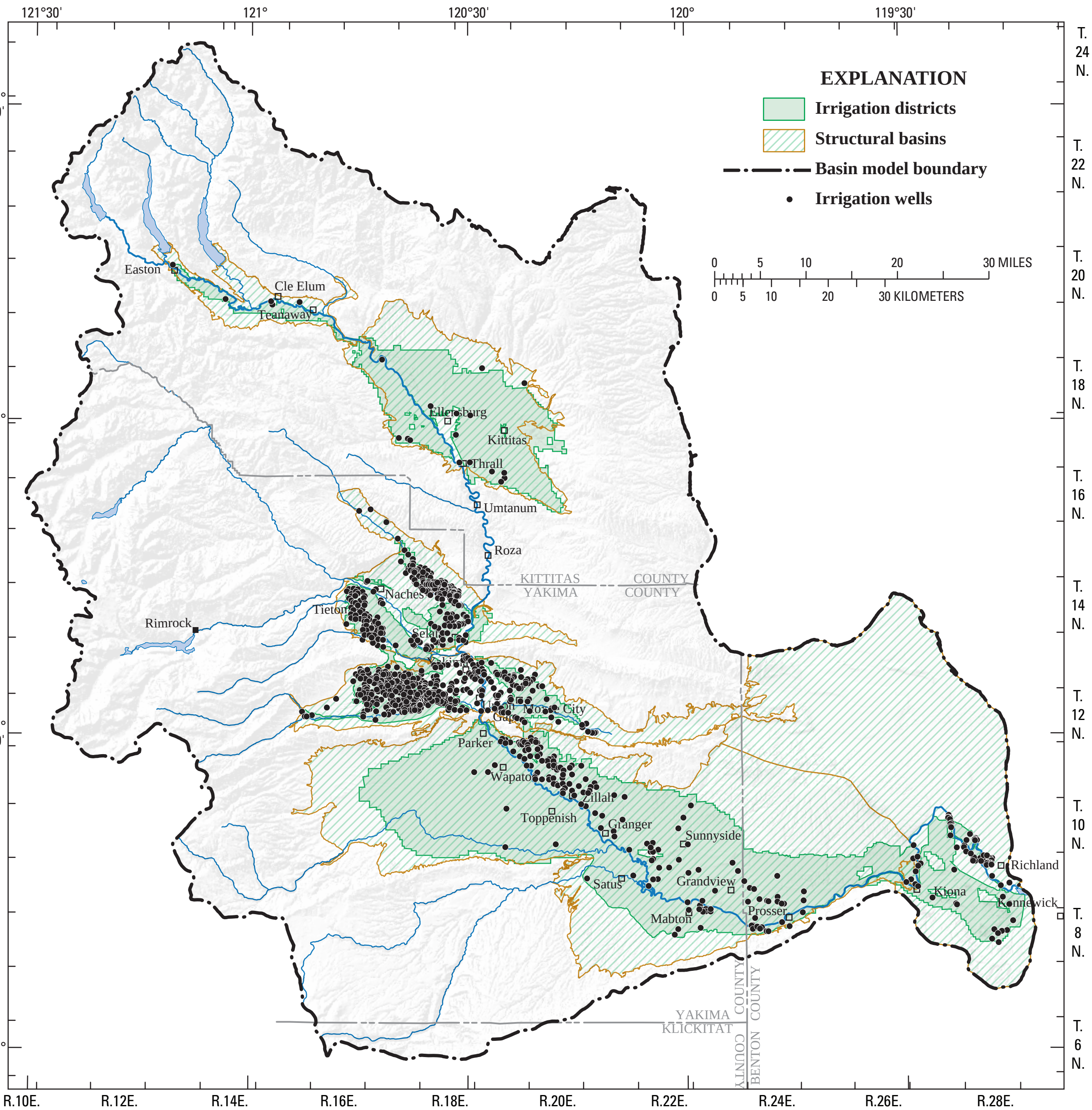


Figure 29. Location of irrigation wells completed in the basin-fill deposits, Yakima River basin aquifer system, Washington.

The areas in which water-level declines have occurred are more widespread in the Wanapum (in the Toppenish basin, the Ahtanum and Moxee subbasins, the Selah basin, and in the Black Rock, Dry Creek, and Cold Creek valleys) than in the Saddle Mountains, and areas of water-level decline greater than 75 ft are larger in the Wanapum. Water-level declines in the Grande Ronde are limited to the western Toppenish basin and the Ahtanum subbasin, with two small areas of decline in the Moxee and Black Rock valleys (fig. 32). Declines of more than 20 ft occur in the Grande Ronde over only about 29 mi^2, less than one-half percent of the areal extent of the unit.

The areas categorized as having water levels that are stable or showing only small declines generally are consistent with areas of surface-water irrigation, basin-fill deposits, small quantities of groundwater pumpage, adequate quantities of recharge, and (or) low population density; basin-fill units with stable water levels also overlie the area of the Saddle Mountains and Wanapum on Rattlesnake Hills in which water levels declined. The surface-water irrigated areas are associated with high recharge values and the low population density areas have widely ranging recharge values. The areas with moderate declines generally do not lie within most surface-water irrigation districts (or if they do, the areas are in deeply buried units), have either medium to large population density and (or) a reasonable amount of croplands irrigated with groundwater. Areas with large declines in water levels are in locations similar to those with moderate declines, but are closely associated with small rates of groundwater recharge, deeply buried units, and large quantities of pumpage for irrigation. Lastly, the areas with very large declines are generally associated with pumpage from units receiving small rates of recharge and (or) large quantities of pumpage from more deeply buried units. For the latter case, the units may outcrop in an area with small rates of recharge or the unit may receive water only from overlying or underlying hydrogeologic units (such as the Grande Ronde in the Moxee subbasin and Black Rock valley). For the areas of moderate to very large water-level declines, the quantity of recharge is important. For example, as described previously, mean annual recharge for current conditions was estimated to be less than 2 in. (about 86 ft^3/s) over about 2,553 mi^2, and during 2000, about 30 percent of the groundwater pumpage (about 130 ft^3/s) was estimated to occur in this area of less than 2 in/yr of recharge. Thus, pumpage is much larger than the recharge in this drier area.

The largest exception to the general relations described above occurs on the south slope of Rattlesnake Hills in a surface-water irrigated area. Moderate to very large declines in water levels have occurred in the Saddle Mountains and Wanapum in this area. The generally fine-grained surficial deposits (Touchet Beds and loess), the slope of the land surface, drains, and the presence of fine-grained layers in Unit 3 (the upper Ellensburg Formation is as much as 600 ft thick) probably act together to limit deeper downward movement of excess surface water (especially through parts of Unit 3). As a result, recharge to the Saddle Mountains and Wanapum units is limited (mean annual recharge in their outcrop areas or where they are overlain by a thin veneer of sediments along Rattlesnake Hills is generally less than 0.5 in.). Locally, pumpage exceeds the sum of (1) recharge, (2) lateral inflow from the upgradient extent of the units, and (3) natural and induced vertical leakage from either Unit 3 to the Saddle Mountains or from the Saddle Mountains to the Wanapum. These factors do not apply to pumpage by wells tapping the basin-fill deposits on Rattlesnake Hills, as demonstrated by the fact that no water-level declines greater than 20 ft have been documented. In this area of naturally small recharge, pumpage demand is likely met by additional recharge from surface-water irrigation. Similar situations occur in or near other surface-water irrigation areas in Yakima County for pumpage by wells tapping basin-fill deposits, and in Benton County for both basin-fill and Saddle Mountains wells.

The factors that influence the degree to which water levels are affected by pumping vary spatially and they exert a wide range of controls on the groundwater flow system. These factors include (1) hydraulic characteristics of the aquifer materials, (2) depth in flow system from which withdrawals are made, (3) extent and location of outcrop area—including the dip and exposure of interflow zones for the CRBG units, (4) structural controls, (5) rate of natural recharge and of recharge from surface-water delivery and irrigation application, and (6) quantity of pumpage. Of these factors, the latter two—rates of recharge and amount of water pumped—probably exert the most control on groundwater levels. Estimated total pumpage (fig. 23) and recharge (fig. 33), accumulated over the base period of 1960–2001, were about 33,000 and 291,000 ft^3/s, respectively. Where pumping occurs, however, the difference between recharge and pumpage (fig. 34) indicates that pumpage was much greater than recharge in a few areas, and in other areas, recharge was much greater than pumpage. The latter case is generally true in the structural basins (where surface water is used for irrigation), in the wetter parts of the basin, and (or) in areas of low population density. Where pumpage is greater than recharge and the upgradient source area is small, such as in the upper Moxee subbasin and Cold Creek drainage, downward trends in groundwater levels are noted (figs. 30–32). Thus, pumpage declines may continue in some areas where pumpage is either greater than recharge or is only slightly less than recharge and the upgradient source area is small. In some areas there appear to be large upgradient recharge areas (such as the western part of the Toppenish basin), yet water levels are declining. Thus, recharge in the upgradient areas cannot meet the groundwater withdrawals —likely due to the depth of the withdrawals, possible further isolation from overlying, tighter basalt flows or fine-grained interbeds, and local discharge of recharge in the dissected uplands. In contrast, where pumpage is much less than recharge, groundwater levels were categorized as stable, such as in the basin-fill deposits in the structural basins. Therefore, total cumulative recharge may be an order of magnitude larger than cumulative pumpage; nonetheless, the spatial distribution of both water-budget components is important for determining the effects of pumpage on the flow system. Indeed, the spatial distribution of recharge is important because mean annual recharge in the Saddle Mountains, Wanapum, and Grande Ronde outcrop areas ranges widely—1.37, 1.47, and 7.69 in., respectively.

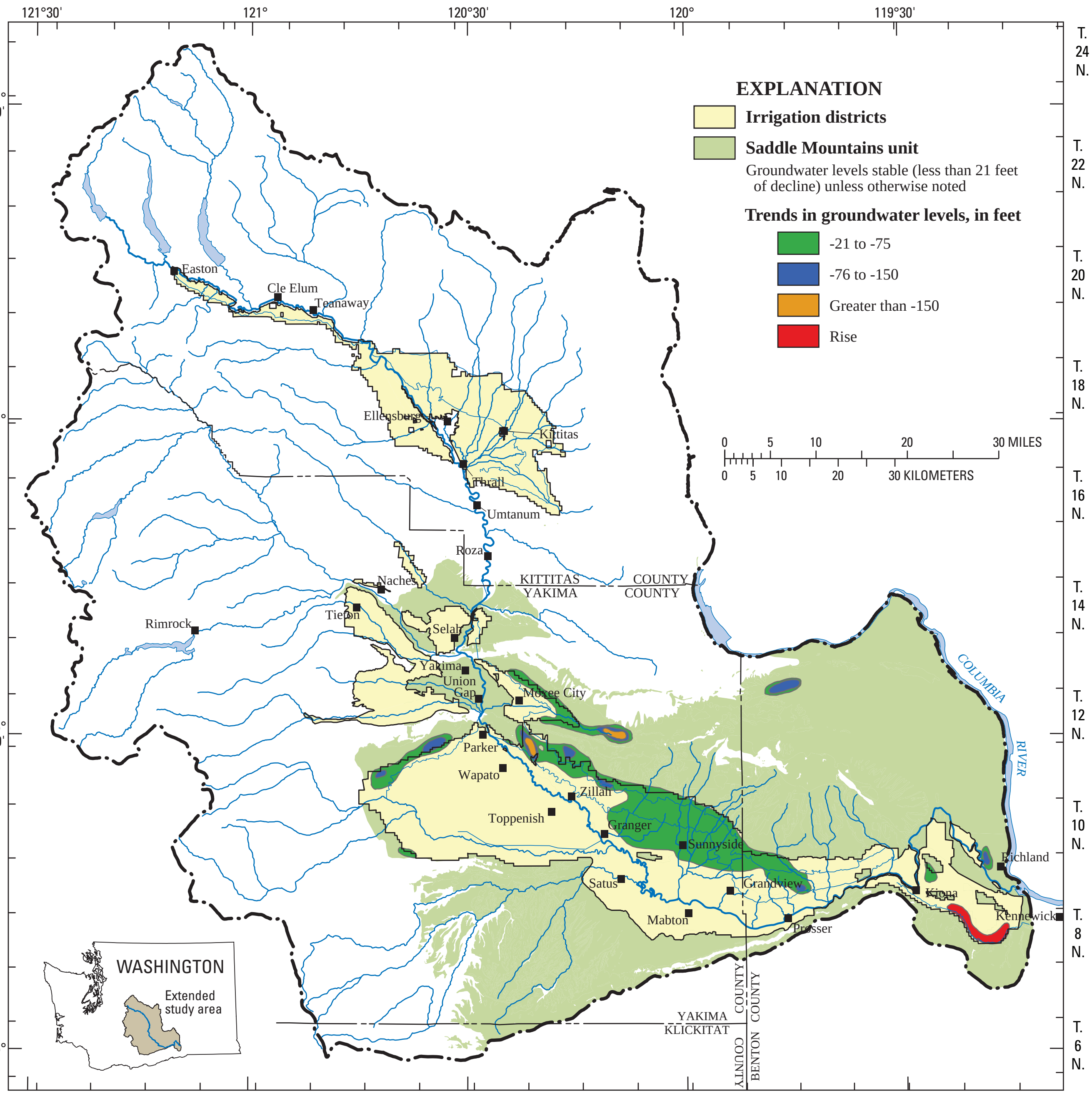


Figure 30. Generalized trends in groundwater levels, Saddle Mountains unit, Yakima River basin aquifer system, Washington.

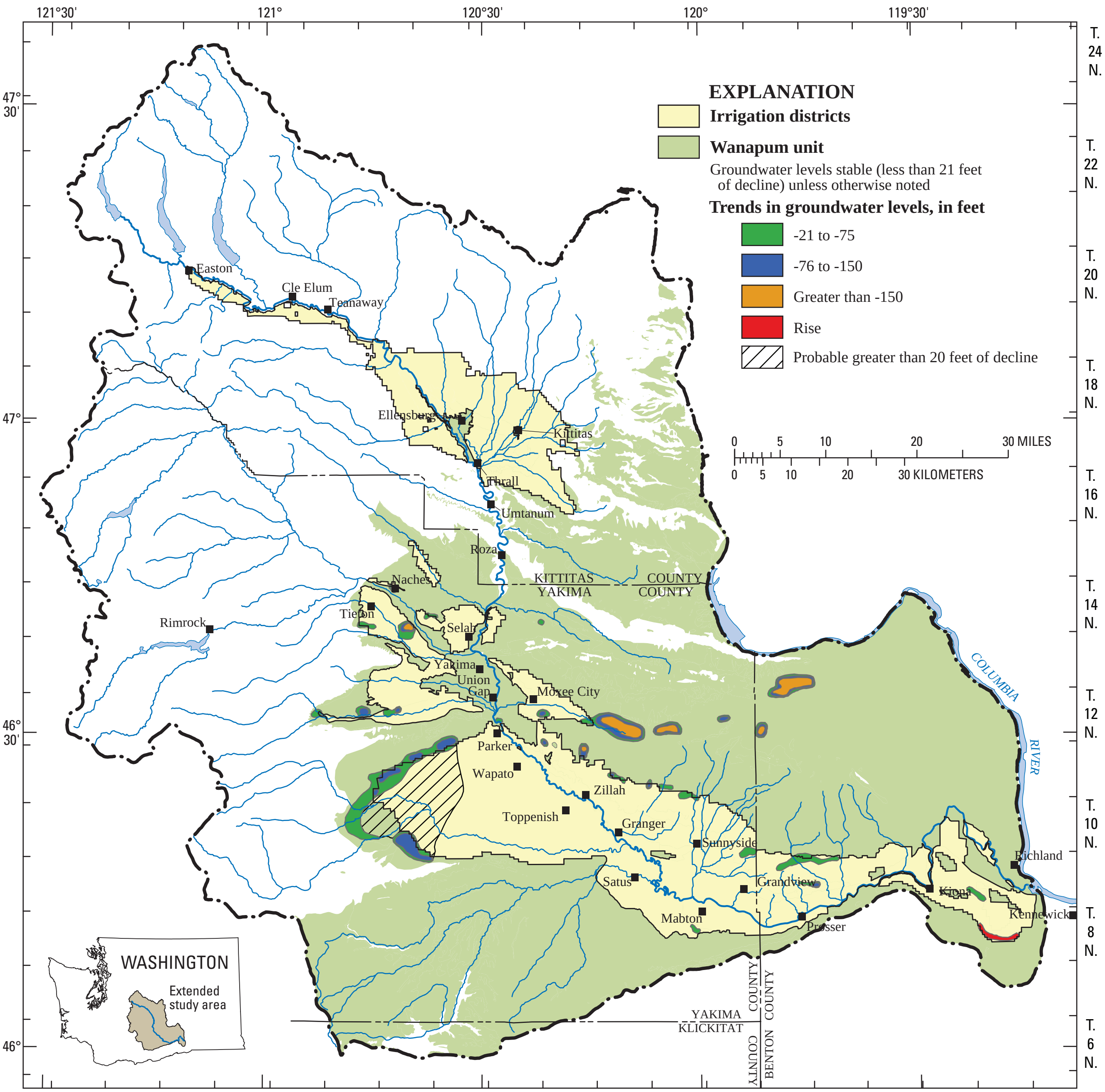


Figure 31. Generalized trends in groundwater levels, Wanapum unit, Yakima River basin aquifer system, Washington.

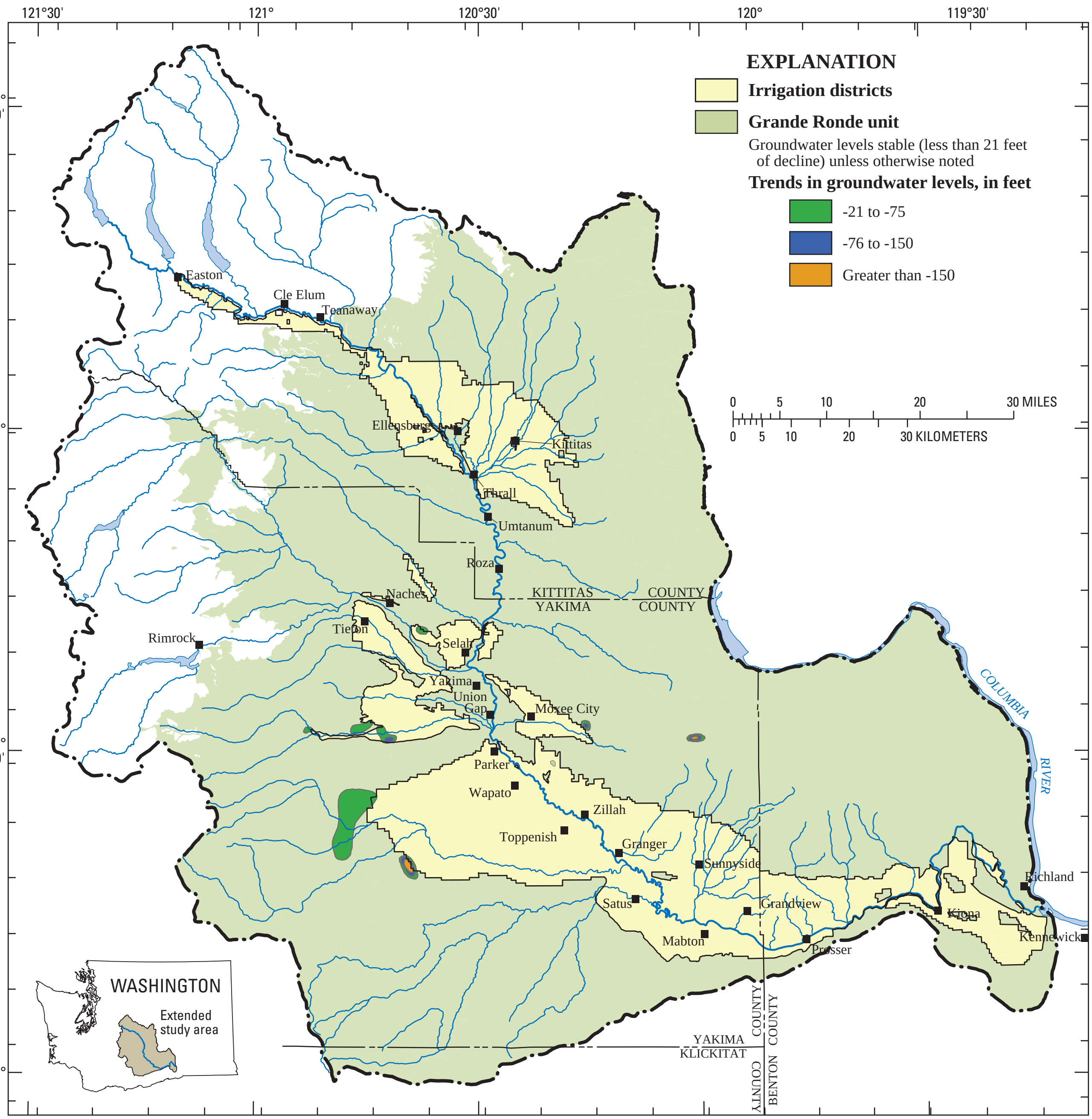


Figure 32. Generalized trends in groundwater levels, Grande Ronde unit, Yakima River basin aquifer system, Washington.

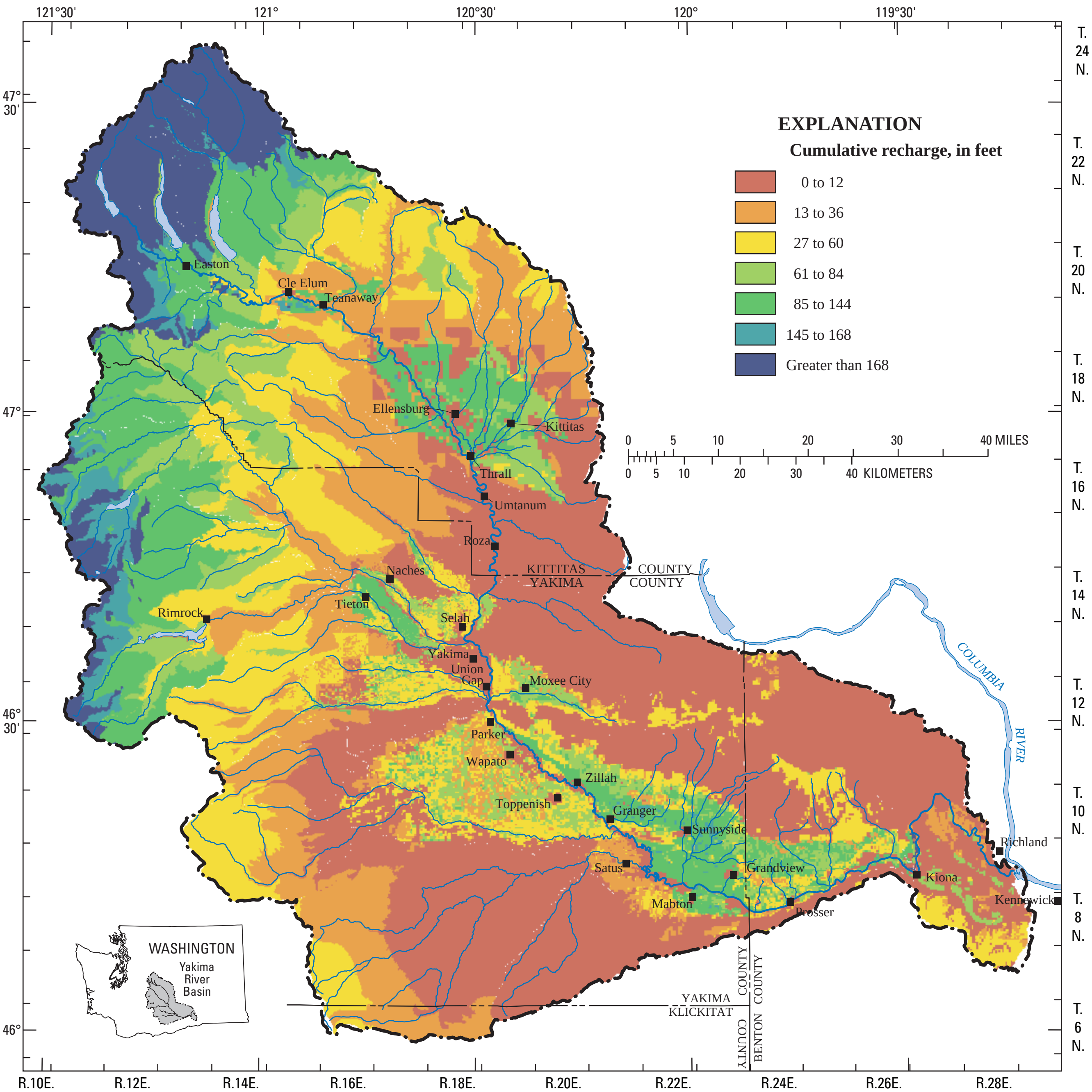


Figure 33. Spatial distribution of cumulative groundwater recharge for 1960–2001, Yakima River basin aquifer system, Washington.

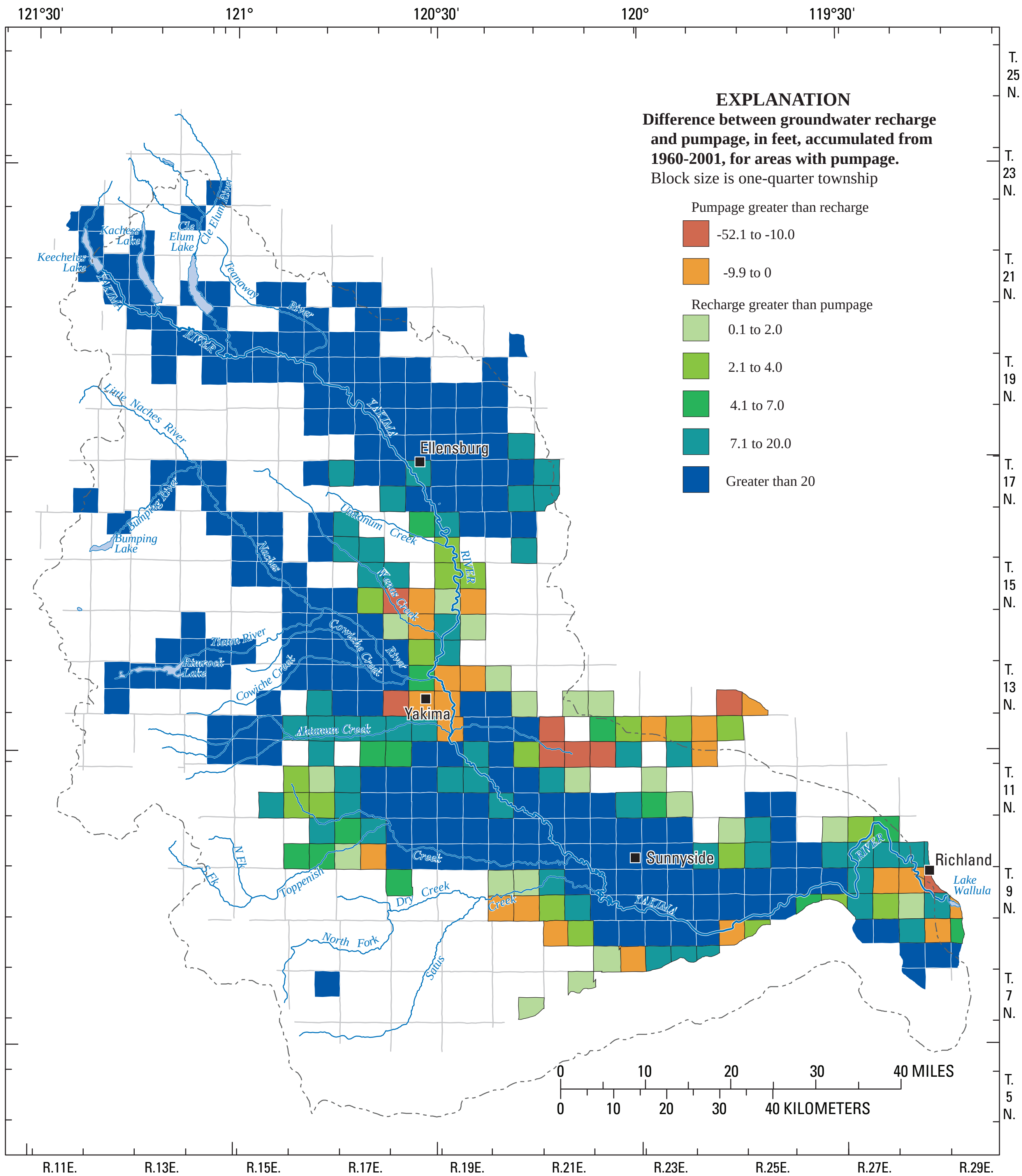


Figure 34. Spatial distribution of the difference between groundwater recharge and pumpage acummulated for 1960–2001 for areas with pumpage, Yakima River basin aquifer system, Washington.

Although data on recharge and pumpage are important for understanding downward trends in water levels, the magnitude of the declines also is controlled by geologic structure, topography, and the extent and hydraulic characteristics of a given hydrogeologic unit. For example, geologic structure can compartmentalize the flow system and thus, limit the upgradient groundwater source area. Similarly, topography can affect the flow system and also limit the source area. The extent and hydraulic characteristics of a unit act in concert to control water-level changes. For example, water levels in a buried unit have less potential to be stable under pumping stresses unless the flow rate into the unit is large. Flow into a buried unit can be large in areas of regional groundwater discharge. For example, the Saddle Mountains unit below the Yakima River would receive not only lateral inflow but also upward flow from the Wanapum. Where the rate of replenishment is small, such as in parts of the Grande Ronde and Unit 3 in the Roslyn basin, water-level declines occur due to pumping; where the rate of replenishment is large, such as in Unit 2 in the Toppenish basin and most of Unit 3 in the Ahtanum subbasin, water levels tend to be stable. The more permeable hydrogeologic units, such as the alluvial units, can be readily recharged and replenished, whereas the less permeable units, such as the Mesozoic unit, are less readily recharged, and lateral flow (replenishment) into a pumping area is limited.

Water Budget for the Yakima River Basin

A generalized mean annual water budget for the Yakima River basin was estimated based on information developed during this study. The budget provides an overview of the magnitude of the various components and includes information important for understanding the framework of the groundwater system in the basin. The water-budget components are: precipitation, streamflow, evapotranspiration, potential recharge, groundwater pumpage, reservoir storage, surface-water diversions, groundwater storage, and changes in storage. The components are briefly described in table 6. These components are, where applicable, shown for both predevelopment (no human activities) and current conditions, and are referenced to the base period, water years 1960–2001. Values presented also are generally rounded values. A more detailed water budget for the aquifer system will be estimated on the basis of the regional groundwater flow model simulations.

The mean annual precipitation over the basin is about 27 in. (about 12,300 ft^3/s or 8.9 million acre-ft). The mean annual precipitation is referenced to the climate normal of 1971–2000, which, based on mean annual streamflow, would be about 4 percent larger than the 1960–2001 base period. This change is similar to the previously described percent decrease in mean annual recharge for the base period in comparison to the 1950–2001 period. Thus, referenced to the base period, mean annual precipitation is about 26 in. (11,800 ft^3/s or 8.6 million acre-ft), and is applicable to both predevelopment and current conditions.

The Yakima River basin produces a mean annual unregulated streamflow (adjusted for regulation and without diversions or returns, and referenced to the base period) of about 5,300 ft^3/s (about 4.1 million acre-ft) and a regulated streamflow, referenced to the base period, of about 3,500 ft^3/s (about 2.5 million acre-ft). The regulated streamflow is based on discharge records collected at the USGS stream-gaging station for the Yakima River at Kiona (station number 12510500) located in the lower part of the basin, which has a long-term mean annual discharge of about 3,500 ft^3/s for its period of record (1934–2007).

The difference between the mean annual precipitation and mean annual unregulated streamflow is about 7,000 ft^3/s (5.1 million acre-ft). Assuming no change in groundwater storage over the base period, this difference indicates that about 57 percent of the precipitation is lost to evapotranspiration under natural conditions. The difference between the mean annual precipitation and mean annual regulated streamflow is about 8,300 ft^3/s (6.1 million acre-ft), indicating that about 70 percent of the precipitation is currently being lost to evapotranspiration. Thus, human activities, in particular irrigated agriculture, have increased evapotranspiration by about 1,300 ft^3/s (0.9 million acre-ft), which is about equivalent to the total storage capacity in the basin.

Table 6. Generalized mean annual water budget for the Yakima River basin, Washington.

[**Predevelopment conditions** assumes 1960–2001 climate without human activities. **Current conditions** represent a composite 1995–2004 land use/land cover; mean annual values referenced to water years 1960–2001. All values in million acre-feet, rounded]

Water budget component	Predevelopment conditions	Current conditions
Precipitation	8.6	8.6
Streamflow	4.1	2.5
Evapotranspiration	5.1	6.1
Recharge[1]	3.8	5.0
Pumpage	.0	.24
Reservoir storage	.0	1.1
Diversions[2]	.0	3.1

[1]For predevelopment conditions, a large part of estimated recharge becomes streamflow; for current conditions, part of the recharge also becomes evapotranspiration.

[2]Does not include power generation diversions.

The potential mean annual recharge for the base period was about 15.2 in. or 6,860 ft^3/s (4.97 million acre-ft). Predevelopment mean annual recharge is about 11.5 in. or 5,250 ft^3/s (3.8 million acre-ft). The predevelopment estimate is also referenced to the base period, which is about 3 percent less than the estimate for the 1950–2003 recharge-estimation period. Most of the recharge supports local streamflow along short flow paths. The estimated mean annual groundwater pumpage for the base period was about 244,000 acre-ft (340 ft^3/s).

Six dams were constructed as part of Reclamation's Yakima Project: Bumping Dam in 1910, Kachess Dam in 1912, Clear Creek Dam in 1914, Keechelus Dam in 1917, Tieton Dam (Rimrock Lake) in 1925, and Cle Elum Dam in 1933. The reservoirs have a total storage capacity of about 1.07 million acre-ft, of which about 78 percent is in the upper arm of the Yakima River and 22 percent is in the Naches River arm.

Excluding diversions for power generation, mean annual discharge (water years 1960–2001) for the 65 diversions was about 3,910 ft^3/s (2.8 million acre-ft). The total for the other 78 points of withdrawals was 474 ft^3/s (0.3 million acre-ft). Together, the non-power quantities total about 4,384 ft^3/s or 3.1 million acre-ft.

Summary and Conclusions

The Yakima River basin aquifer system underlies about 6,200 mi^2 in south-central Washington. The study-area boundary was extended eastward to the Columbia River (incorporating an additional 700 mi^2) to provide a more definitive hydrologic boundary for construction of a regional groundwater flow model.

The aquifers in the Yakima River basin consist of sedimentary deposits in six structural basins, volcanic rocks of the Columbia River Basalt Group (CRBG), and generally older bedrock. The basin-fill deposits were divided into 19 hydrogeologic units, the CRBG was divided into three units separated by two interbed units, and the bedrock was divided into four units (based on geologic age) and several subunits. The thickness of the basin-fill units and the depth to the top of each unit and interbed of the CRBG were mapped. Only the surficial extent of the bedrock units was mapped due to insufficient data.

The lithology of the basin-fill units ranges from coarse-grained alluvial-valley deposits to fine-grained clay-dominated deposits. In three of the structural basins, an alluvial, unconsolidated, and a consolidated unit were mapped, although five units were mapped in the Toppenish basin. Two units, one unconsolidated and the other consolidated, were mapped in the Benton basin; in the Eastern Benton basin, the unconsolidated unit was further divided into six subunits which occur together only in this part of the study area. Average mapped thickness of the different units ranged from 10 to 600 ft and the median thickness ranged from 10 to 450 ft.

The CRBG units consist of three formations and their interbedded sediments. The formations were called units to account for the sediments and they are named, from oldest to youngest, the Grande Ronde, Saddle Mountains, and Wanapum units. The Grande Ronde unit is the most extensive and the Saddle Mountains the least extensive. Average mapped thicknesses of the Saddle Mountains and Wanapum units were 550 and 600 ft, respectively. The thickness of the Grande Ronde unit was not mapped due to insufficient data. The Mabton Interbed lies between the Saddle Mountains and Wanapum units, and the Vantage Interbed lies between the Wanapum and Grande Ronde units. The Mabton Interbed averages 70 ft in thickness and the Vantage Interbed averages 30 ft.

The four bedrock units, consisting of the Paleozoic, the Mesozoic, the Tertiary, and the Quaternary bedrock, underlie the western part of the basin. The combined areas of these units where they abut the CRBG are about 1,520 mi^2. The Tertiary unit has the largest extent (1,300 mi^2) and the Paleozoic unit the smallest (2 mi^2). The Paleozoic unit consists of metamorphic rocks and the Mesozoic unit has three subdivisions consisting of metamorphic, volcanic, and plutonic rocks. The Tertiary unit has three subdivisions consisting of sediments, and volcanic and plutonic rocks. The Quaternary bedrock unit consists principally of volcanics but in a few small areas it consists of sediments.

Lateral hydraulic conductivity (K_h) of the hydrogeologic units ranges widely, indicating the heterogeneity of the geologic materials making up the aquifer system. Average or effective K_h values of the water-producing zones of the basin-fill units are on the order of 1 to 800 ft/d and are about 1 to 5 ft/d for the CRBG units. Effective K_h for the different rock types of the Paleozoic, Mesozoic, and Tertiary bedrock units appear to be about 0.0001 to 3 ft/d. The more permeable Quaternary bedrock unit may have K_h values that range from 1 to 7,000 ft/d. Based on specific-capacity data, mean and median values for basin-fill units were 182 and 6 ft/d, respectively. The specific-capacity derived mean and median values for CRBG units were 182 and 3 ft/d, respectively, and for the bedrock units they were 13 and 3 ft/d, respectively. Vertical hydraulic conductivity (K_v) of the units is largely unknown. K_v values have been estimated to range from about 0.009 to 2 ft/d for the basin-fill units; values for the clay-to-shale parts of these units may be as small as 10^{-10} to 10^{-7} ft/d. Reported K_v values for the CRBG units range from 4×10^{-7} to 4 ft/d.

Variations in the concentrations of geochemical solutes and the concentrations and ratios of the isotopes of hydrogen, oxygen, and carbon in groundwater provide information on the hydrogeologic framework of the groundwater system in the Yakima basin. The concentrations of dissolved constituents in groundwater vary in response to the interactions of groundwater with the aquifer matrix material. Dissolved solutes in basaltic groundwater of the CRBG are derived principally from the dissolution of basaltic glass and to a lesser extent plagioclase feldspar. Ratios of the stable isotopes of hydrogen and oxygen (deuterium and oxygen-18) indicated dispersed sources of groundwater recharge to the CRBG and Ellensburg aquifers. Carbon and tritium isotope data and the concentrations of dissolved constituents indicate a complex groundwater flow system with multiple contributing zones to groundwater wells and relative groundwater residence time on the order of a few tens to many thousands of years. The concentrations of dissolved methane were larger than could be attributed to atmospheric sources in more than 80 percent of wells with measured methane concentrations. Concentrations of methane as large as 50 mg/L were measured even though dissolved concentrations of noble gases argon and neon showed substantial alteration from atmospheric concentrations, likely due to partial stripping of the gases during sampling. The stable carbon isotopic composition ($\delta^{13}C_{methane}$; $^{13}C/^{12}C$) of dissolved methane varied from -32.7 to -53.0 ‰—which is indicative of a thermogenic source for the methane. Most of the detections of methane were in samples from irrigation and municipal wells at locations several miles distant from mapped structural fault features, suggesting the upward vertical movement of thermogenic methane from the underlying bedrock may be more widespread than assumed or the presence of unmapped (buried) fault structures. The spatial distribution of methane suggests both diffuse and point sources (faults).

Potential mean annual recharge for current conditions for water years 1950–2003 was estimated to be about 15.6 in. or 7,149 ft^3/s (5.2 million acre-ft) and if there had been no human activities (predevelopment conditions) during that period, estimated recharge would have been about 11.9 in. or 5,450 ft^3/s (3.9 million acre-ft). Recharge amounts range widely with location within the basin. Estimated mean annual recharge ranges from virtually zero in the dry parts of the lower basin to more than 100 in. in the humid uplands that receive more than 120 in/yr of precipitation. Within the Yakima River basin, mean annual recharge for the base period of interest (water years 1960–2001) was about 15.2 in. or 6,860 ft^3/s (4.97 million acre-ft). These values account for the refinements in selected irrigation districts for the years with prorating when junior surface-water users do not obtain their full entitlement, and for the different base period (1960–2001 in contrast to 1950–2003). For the extended study area, mean annual recharge was estimated to be about 13.6 in. or 6,950 ft^3/s.

Groundwater in the study area occurs under perched, unconfined, semiconfined, and confined conditions. The water moves from topographic highs in the uplands to topographic low areas along the streams. The flow system in the basin-fill units is compartmentalized due to topography and geologic structure. The system also is compartmentalized for the CRBG units but not to as large an extent as the basin-fill units. Regional groundwater flow discharges to surface drainage features in the lowlands in the structural basins and to the Columbia River. The gradient of the water table ranges from about 7 feet per mile to more than 400 feet per mile, with the more “gentle” gradients in the topographically smooth parts of the structural basins. Typically, hydraulic gradients are similar to topographic gradients. The lateral hydraulic gradient in the CRBG units also is highly variable and ranges from about 3 to 14 feet per mile in the flatter parts of the structural basins to as much as 800 feet per mile in areas of steep terrain. In the CRBG units, the hydraulic gradient generally is within 5 degrees of the topographic gradient.

There are nearly 2,874 groundwater rights in the basin, with an annual total of 530,000 acre-ft (730 ft^3/s). Allowable acreage of groundwater irrigation rights is about 130,000 acres. A total of 312,284 acre-ft of water (about 430 ft^3/s) was estimated to have been pumped in 2000 for multiple uses, about 60 percent of which was for irrigation. Standby/reserve pumpage was estimated to have ranged from about 9,200 acre-ft in 1988 to 52,100 acre-ft in 1994 and 2001, and total standby/reserve pumpage for the 8 prorating years through 2001 was about 250,350 acre-ft. Surface-water diversions (minus power returns) are about 4,383 ft^3/s, of which about 66 percent was delivered for irrigation.

Excluding the initial water-level rises due to surface-water irrigation that started in the late 1800s, long-term measurements of the levels indicate they have been stable or have declined less than 20 ft in most of the aquifer system. Declines from 21 to 300 ft have occurred in some areas due to pumpage. The largest declines have been in the CRBG units. The pumpage from the basin-fill units, from which about 50 percent of the pumpage occurs, appears to have caused water-level declines in only two small areas, but the withdrawals may be reducing groundwater discharge.

For current conditions, the largest mean annual water-budget component was precipitation (8.6 million acre-ft), followed by evapotranspiration (6.1 million acre-ft). Potential groundwater recharge was estimated to be 5.0 million acre-ft, and streamflow was about one-half of the recharge (2.5 million acre-ft). Diversions amounted to about 3.1 million acre-ft, which is about three times larger than the reservoir capacity in the basin (1.1 million acre-ft). Pumpage was the smallest water budget component at 0.25 million acre-ft.

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Appendix A. Methods and Assumptions Used to Estimate Lateral Hydraulic Conductivity from Specific-Capacity Data

The modified Theis equation (Ferris and others, 1962, p. 99) was used to estimate transmissivity of the pumped interval for wells with a screened or perforated interval. Transmissivity is the product of hydraulic conductivity and thickness of the part of the hydrogeologic unit supplying water to the well. The modified equation is

$$s = \frac{Q}{4\pi T} \ln \frac{2.25Tt}{r^2 S} , \qquad (1)$$

where

- s is drawdown in the well, in feet;
- Q is discharge, or pumping rate, of the well, in cubic feet per day;
- T is transmissivity of the hydrogeologic unit, in foot squared per day;
- t is length of time the well was pumped, in days;
- r is radius of the well, in feet; and
- S is storage coefficient, a dimensionless number, assumed to be 0.0001 for basalt and bedrock units and 0.1 for overburden units.

Assumptions for using equation 1 are that aquifers are homogeneous, isotropic, and infinite in extent; wells are fully penetrating; flow to the well is lateral; steady-state drawdown has been achieved; and water is released from storage instantaneously. Additionally, for unconfined aquifers, drawdown is assumed to be small in relation to the saturated thickness of the aquifer. Although many of the assumptions are not precisely met, the field conditions in the study area approximate most of the assumptions and the calculated hydraulic conductivities are reasonable estimates.

A computer program was used to solve equation 1 for transmissivity (T) using Newton's iterative method (Carnahan and others, 1969). The iterative approach is necessary because T cannot be easily solved by a direct solution method. Next, the following equation was used to calculate hydraulic conductivity:

$$K_h = \frac{T}{b} , \qquad (2)$$

where

- K_h is lateral hydraulic conductivity of the geologic material in the vicinity of the well opening, in feet per day; and
- b is thickness, in feet, approximated using the length of the open interval as reported in the driller's report.

The use of the length of a well's open interval for b may overestimate values of K_h because the equations assume that all the water flows laterally within a layer of this thickness. Although some of the flow will be outside this region, the amount can be expected to be small because in most deposits, vertical flow is inhibited by layering. The difference in computed transmissivity between using 0.1 and 0.0001 for the storage coefficient is a factor of only about 2. For wells for which no data were available for the screen interval or time of pumping, a value of 100 ft and 1 hour, respectively, were used.

Appendix B. Hydrochemical Data from Groundwater Samples, Yakima River Basin Aquifer System, Washington

Table B1. Hydrochemical data from groundwater samples, September 1990 through August 2005, Yakima River basin aquifer system, Washington.

[**Hydrogeologic unit:** CN,consolidated deposits (Ellensburg Formation); GR, Grande Ronde; SM, Saddle Mountains; UNCON, unconsolidated deposits-mainly alluvium; WN, Wanapum. **Abbreviations:** NGVD 29, National Geodetic Vertical Datum of 1929; E, estimated value below the long term detection limit; M, presence of material verified but not quantified; DIC, dissolved inorganic carbon; PMC, percent modern carbon; NTU, Nephalometric Turbidity Unit; TU, tritium units; ccSTP/g, cubic centimeter at standard temperature and pressure per gram water; °C, degrees Celsius; mg/L, milligram per liter; µS/cm, microseimens per centimeter; µg/L, microgram per liter; δ, delta stable isotpe abundance; <, less than; –, no data available]

Well No.	Collection agency	Date	Hydrogeo-logic unit	Land-surface altitude (feet above NGVD 29)	Hole depth (ft)	Well depth (ft)	Tempe-rature (°C)	Specific conduc-tance (µS/cm)	pH	Dissolved oxygen (mg/L)	Total dissolved solids	Turbidity (NTU)	Carbonate (mg/L)	Bicar-bonate (mg/L)
08N-24E-01J01	USGS	09-27-00	WN	734	1,264	1,033	21.1	403	8.6	M	–	–	–	–
08N-28E-09F01	USGS	09-05-01	WN	935	1,262	920	25.3	438	8.1	<0.1	–	–	–	–
09N-23E-22J01	USGS	09-04-01	SM-WN	827	1,632	308	18.5	410	7.7	4.0	–	–	–	–
09N-23E-22L01	USGS	09-05-01	WN	770	1,294	739	25.6	405	9.2	<.1	–	–	–	–
09N-27E-24N01	USGS	09-05-01	WN	830	1,378	922	23.7	410	8.4	<.1	–	–	–	–
10N-16E-19H01D1	USGS	09-11-01	GR	1,540	425	30	20.0	265	7.5	8.0	–	–	–	–
10N-16E-21D01	USGS	09-11-01	CN	1,390	158	111	14.2	207	7.2	<.1	–	–	–	–
10N-17E-27Q01	USGS	09-07-01	SM-WN-GR	1,120	1,510	134	–	425	7.2	–	–	–	–	–
10N-20E-04L01	USGS	09-07-01	CN	760	1,020	810	22.4	209	7.9	.1	–	–	–	–
10N-20E-09A01	USGS	09-07-01	CN	757	863	783	20.1	162	7.9	<.1	–	–	–	–
10N-20E-09A02	USGS	09-07-01	CN	760	250	138	15.2	119	8.2	4.5	–	–	–	–
10N-22E-25F03	USGS	02-27-00	WN	748	1,701	1,283	25.3	310	8.5	.0	–	–	–	–
11N-18E-26M03	USGS	09-06-01	WN	835	1,471	1,385	30.5	335	7.7	M	–	–	–	–
11N-18E-27H01	USGS	02-12-01	CN	840	111	111	–	376	7.6	8.8	–	–	–	–
11N-21E-34F01	USGS	09-12-01	CN	970	198	195	15.5	997	7.6	8.6	–	–	–	–
11N-22E-30C01	USGS	09-06-01	GR	1,220	2,715	2,049	28.8	333	9.1	M	–	–	–	–
11N-22E-33M01	USGS	09-12-01	SM	1,130	1,105	702	22.0	281	8.1	<.1	–	–	–	–
12N-20E-12K01	USGS	08-24-05	GR	1,435	2,700	1,550	31.6	319	8.7	<.1	–	–	–	–
13N-20E-34Q01D1	USGS	08-24-05	WN	1,548	540	–	23.6	347	7.9	<.1	–	–	–	–
13N-27E-26H01	USGS	09-12-01	GR	402	4,340	2,260	20.2	1,230	10.0	<.1	–	–	–	–
13N-20E-33N01	USGS	10-01-04	WN	1,285	1,628	1,000	–	273	8.0	.1	–	–	–	–
11N-19E-06N	Yakama Nation	02-28-91	UNCON	885	53	50	14.3	316	7.0	7.6	114	1.2	<1.0	163
11N-19E-34A01	Yakama Nation	02-28-91	UNCON	815	62	57	13.5	269	6.9	8.2	82.0	.2	<1.0	28.0
11N-19E-34B01A	Yakama Nation	04-10-91	UNCON	815	226	216	15.7	132	7.4	6.0	120	32	<1.0	74
11N-18E-03E	Yakama Nation	03-06-91	UNCON	940	195	190	12.6	860	7.3	5.2	573	.5	<1.0	439
11N-18E-20R02A	Yakama Nation	04-15-91	UNCON	860	45	35	13.9	639	7.2	1.9	532	29	<1.0	340
11N-18E-20R02B	Yakama Nation	04-15-91	UNCON	860	64	58	13.6	621	7.3	2.0	484	4.5	<1.0	33.0
11N-17E-23D02	Yakama Nation	03-06-91	UNCON	890	95	90	13.4	366	7.0	6.6	344	.6	<1.0	153
11N-16E-23B02	Yakama Nation	03-06-91	UNCON	1,185	90	85	11.2	345	6.9	3.8	324	.2	<1.0	209

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

[**Hydrogeologic unit:** CN,consolidated deposits (Ellensburg Formation); GR, Grande Ronde; SM, Saddle Mountains; UNCON, unconsolidated deposits-mainly alluvium; WN, Wanapum. **Abbreviations:** NGVD 29, National Geodetic Vertical Datum of 1929; E, estimated value below the long term detection limit; M, presence of material verified but not quantified; DIC, dissolved inorganic carbon; PMC, percent modern carbon; NTU, Nephalometric Turbidity Unit; TU, tritium units; ccSTP/g, cubic centimeter at standard temperature and pressure per gram water; ºC, degrees Celsius; mg/L, milligram per liter; µS/cm, microseimens per centimeter; µg/L, microgram per liter; δ, delta stable isotpe abundance; <, less than; –, no data available]

Well No.	Collection agency	Date	Alkalinity (mg/L)	Potassium (mg/L)	Manga- nese (mg/L)	Magne- sium (mg/L)	Calcium (mg/L)	Boron (mg/L)	Phosphate (dissolved) (mg/L)	Total phos- phorus (mg/L)	Nitrite (mg/L)	Nitrite + nitrate as N (mg/L)	Ammonia as N (mg/L)	Sulphate (mg/L)
08N-24E-01J01	USGS	09-27-00	239	15.8	–	3.94	10.9	–	–	–	–	–	–	0.59
08N-28E-09F01	USGS	09-05-01	220	13.5	–	6.92	15.9	–	–	–	–	<0.05	–	30.1
09N-23E-22J01	USGS	09-04-01	163	5.02	–	14.1	42.3	–	–	–	–	4.48	–	41.5
09N-23E-22L01	USGS	09-05-01	212	9.98	–	.04	.86	–	–	–	–	<.05	–	.64
09N-27E-24N01	USGS	09-05-01	212	14.1	–	2.13	8.61	–	–	–	–	<.05	–	10.7
10N-16E-19H01D1	USGS	09-11-01	161	4.23	–	13.5	25.0	–	–	–	–	.3	–	3.22
10N-16E-21D01	USGS	09-11-01	116	2.95	–	10.3	19.7	–	–	–	–	.4	–	2.77
10N-17E-27Q01	USGS	09-07-01	–	6.35	–	17.6	26.8	–	–	–	–	<.05	–	.25
10N-20E-04L01	USGS	09-07-01	131	5.22	–	5.07	12.6	–	–	–	–	<.05	–	E.06
10N-20E-09A01	USGS	09-07-01	107	3.38	–	3.05	11.1	–	–	–	–	E.03	–	E.09
10N-20E-09A02	USGS	09-07-01	65.0	1.66	–	3.52	12.6	–	–	–	–	.39	–	2.74
10N-22E-25F03	USGS	02-27-00	171	14.0	–	.62	3.17	–	–	–	–	–	–	2.62
11N-18E-26M03	USGS	09-06-01	–	7.90	–	11.8	21.8	–	–	–	–	–	–	.16
11N-18E-27H01	USGS	02-12-01	144	4.75	–	12.9	34.9	–	–	–	–	1.62	–	28.9
11N-21E-34F01	USGS	09-12-01	–	5.17	–	33.2	107	–	–	–	–	17.4	–	139
11N-22E-30C01	USGS	09-06-01	–	10.3	–	.17	2.52	–	–	–	–	<.05	–	.38
11N-22E-33M01	USGS	09-12-01	–	6.56	–	9.34	25.4	–	–	–	–	E.03	–	29.9
12N-20E-12K01	USGS	08-24-05	160	7.79	–	.60	3.31	–	–	–	–	<.06	–	.33
13N-20E-34Q01D1	USGS	08-24-05	157	6.63	–	11.1	20.7	–	–	–	–	<.06	–	14.4
13N-27E-26H01	USGS	09-12-01	–	6.43	–	E.004	1.87	–	–	–	–	<.05	–	123
13N-20E-33N01	USGS	10-01-04	142	8.61	–	4.46	11.5	–	–	–	–	<.06	–	E.14
11N-19E-06N	Yakama Nation	02-28-91	134	3.00	–	11.0	32.0	<0.1	0.06	0.06	<0.002	2.40	<0.07	15.8
11N-19E-34A01	Yakama Nation	02-28-91	23.0	3.00	–	10.0	27.0	<.1	.09	.09	<.002	3.20	<.07	8.70
11N-19E-34B01A	Yakama Nation	04-10-91	61.0	2.00	–	2.70	11.0	.1	<.05	.05	<.002	.60	.56	5.80
11N-18E-03E	Yakama Nation	03-06-91	360	3.00	–	35.0	76.0	<.1	<.05	<.05	<.002	1.70	<.07	87.4
11N-18E-20R02A	Yakama Nation	04-15-91	280	5.00	0.13	21.0	64.0	<.1	.12	.09	<.002	6.50	<.07	37.0
11N-18E-20R02B	Yakama Nation	04-15-91	270	3.00	.02	24.0	58.0	<.1	.08	.08	<.002	4.60	<.07	29.8
11N-17E-23D02	Yakama Nation	03-06-91	125	2.00	<.01	16.0	31.0	<.1	.09	.08	<.002	1.10	<.07	41.5
11N-16E-23B02	Yakama Nation	03-06-91	171	4.00	<.01	14.0	23.0	<.1	.15	.17	<.002	.60	<.07	6.80

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

[**Hydrogeologic unit:** CN,consolidated deposits (Ellensburg Formation); GR, Grande Ronde; SM, Saddle Mountains; UNCON, unconsolidated deposits-mainly alluvium; WN, Wanapum. **Abbreviations:** NGVD 29, National Geodetic Vertical Datum of 1929; E, estimated value below the long term detection limit; M, presence of material verified but not quantified; DIC, dissolved inorganic carbon; PMC, percent modern carbon; NTU, Nephalometric Turbidity Unit; TU, tritium units; ccSTP/g, cubic centimeter at standard temperature and pressure per gram water; °C, degrees Celsius; mg/L, milligram per liter; µS/cm, microseimens per centimeter; µg/L, microgram per liter; δ, delta stable isotpe abundance; <, less than; –, no data available]

Well No.	Collection agency	Date	Chloride (mg/L)	Sodium (mg/L)	SiO_2 (mg/L)	Fluoride (mg/L)	Dissolved organic carbon (mg/L)	$\delta^{18}O$ (per mil)	δD (per mil)	Tritium (TU)	^{14}C DIC (PMC)	$\delta^{13}C$ DIC (per mil)	Hydrogen sulfide (mg/L)	Oxidation reduction potential (millivolts)
08N-24E-01J01	USGS	09-27-00	8.88	75.5	64.6	0.91	0.40	-17.2	-133	1.28	19.2	-15.5	–	-287
08N-28E-09F01	USGS	09-05-01	9.48	62.3	68.0	.93	.90	-17.9	-141	.19	5.57	-13.4	0.19	-324
09N-23E-22J01	USGS	09-04-01	13.6	15.7	60.3	.4	1.0	-15.7	-129	.41	28.0	-12.3	--	-324
09N-23E-22L01	USGS	09-05-01	20.5	84.9	80.5	2.75	.70	-18.4	-144	.06	2.03	-1.37	.37	–
09N-27E-24N01	USGS	09-05-01	17.4	71.1	68.0	1.44	1.0	-17.9	-139	.50	3.87	-5.99	.01	-388
10N-16E-19H01D1	USGS	09-11-01	1.64	8.18	61.3	E.15	.50	-15.2	-116	.13	85.2	-16.9	–	-286
10N-16E-21D01	USGS	09-11-01	2.28	6.23	53.9	E.14	.50	-15.1	-114	1.53	89.2	-16.9	–	–
10N-17E-27Q01	USGS	09-07-01	4.74	39.7	89.5	.90	1.3	-16.7	-130	.00	3.22	-8.52	–	-220
10N-20E-04L01	USGS	09-07-01	3.02	22.4	71.9	.65	.30	-16.3	-124	.13	11.4	-12.1	.08	-203
10N-20E-09A01	USGS	09-07-01	2.2	17.5	65.0	.55	1.0	-15.8	-118	M	21.8	-13.3	.01	-205
10N-20E-09A02	USGS	09-07-01	1.26	5.69	23.7	<.20	.80	-14.4	-106	1.38	84.5	-13.5	–	45.0
10N-22E-25F03	USGS	02-27-00	8.85	64.5	69.6	1.05	.20	-18.2	-144	.00	7.20	-11.1	.62	-307
11N-18E-26M03	USGS	09-06-01	4.77	34.6	83.4	.80	1.8	-17.7	-138	.00	5.30	-11.5	.01	-234
11N-18E-27H01	USGS	02-12-01	10.2	21.7	55.2	.17	.60	-16.9	-132	.00	38.8	-11.6	<.01	24.1
11N-21E-34F01	USGS	09-12-01	46.1	38.2	43.8	.26	2.7	-15.6	-129	12.0	91.8	-13.7	–	–
11N-22E-30C01	USGS	09-06-01	12.1	68.3	46.6	6.08	7.0	-18.2	-143	.06	1.32	2.80	.07	130
11N-22E-33M01	USGS	09-12-01	5.46	15.1	54.9	.48	.30	-15.6	-128	.16	20.1	-11.8	<.01	--
12N-20E-12K01	USGS	08-24-05	4.83	63.2	84.2	1.48	E.30	-17.8	-139	.00	1.99	-5.36	.07	-160
13N-20E-34Q01D1	USGS	08-24-05	7.11	36.1	68.7	.85	.44	-17.1	-135	.16	12.3	-10.6	.03	-20.7
13N-27E-26H01	USGS	09-12-01	146	255	119	38.9	3.8	-14.5	-120	.00	–	-20.0	1.60	-350
13N-20E-33N01	USGS	10-01-04	4.79	40.4	66.2	1.06	.40	-17.4	-137	.09	9.83	-8.71	–	–
11N-19E-06N	Yakama Nation	02-28-91	5.80	13.0	31.0	<.20	.91	-14.5	-105	10.1	–	–	–	–
11N-19E-34A01	Yakama Nation	02-28-91	3.90	10.0	31.0	<.20	.74	-13.9	-107	9.50	–	–	–	–
11N-19E-34B01A	Yakama Nation	04-10-91	1.80	14.0	22.0	<.20	.30	-15.0	-105	.00	–	–	–	–
11N-18E-03E	Yakama Nation	03-06-91	23.6	62.0	54.0	.50	3.2	-15.3	-114	9.95	98.7	-13.9	–	–
11N-18E-20R02A	Yakama Nation	04-15-91	13.9	34.0	34.0	.30	1.3	-14.5	-106	19.9	–	–	–	–
11N-18E-20R02B	Yakama Nation	04-15-91	11.0	43.0	43.0	.30	1.1	-14.6	-108	26.9	–	–	–	–
11N-17E-23D02	Yakama Nation	03-06-91	20.0	20.0	55.0	.40	1.0	-15.8	-122	1.10	–	–	–	–
11N-16E-23B02	Yakama Nation	03-06-91	3.30	25.0	48.0	.40	.60	-15.6	-115	9.00	–	–	–	–

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

[**Hydrogeologic unit:** CN,consolidated deposits (Ellensburg Formation); GR, Grande Ronde; SM, Saddle Mountains; UNCON, unconsolidated deposits-mainly alluvium; WN, Wanapum. **Abbreviations:** NGVD 29, National Geodetic Vertical Datum of 1929; E, estimated value below the long term detection limit; M, presence of material verified but not quantified; DIC, dissolved inorganic carbon; PMC, percent modern carbon; NTU, Nephalometric Turbidity Unit; TU, tritium units; ccSTP/g, cubic centimeter at standard temperature and pressure per gram water; °C, degrees Celsius; mg/L, milligram per liter; µS/cm, microseimens per centimeter; µg/L, microgram per liter; δ, delta stable isotpe abundance; <, less than; –, no data available]

Well No.	Collection agency	Date	Nitrogen (mg/L)	Argon (ccSTP/g)	Neon (ccSTP/g)	Hellium (ccSTP/g)	Krypton (ccSTP/g)	Xenon (ccSTP/g)	Methane (mg/L)	$\delta^{13}C$-CH4 (per mil)
08N-24E-01J01	USGS	09-27-00	–	–	–	–	–	–	–	–
08N-28E-09F01	USGS	09-05-01	18.8	–	–	–	–	–	2.38	-45.4
09N-23E-22J01	USGS	09-04-01	19.1	–	–	–	–	–	.88	-32.7
09N-23E-22L01	USGS	09-05-01	1.81	–	–	–	–	–	42.7	-35.5
09N-27E-24N01	USGS	09-05-01	11.0	–	–	–	–	–	49.8	-35.1
10N-16E-19H01D1	USGS	09-11-01	–	–	–	–	–	–	–	–
10N-16E-21D01	USGS	09-11-01	20.0	4.09E-04	2.47E-07	6.34E-08	8.92E-08	1.29E-08	<.01	–
10N-17E-27Q01	USGS	09-07-01	12.6	2.99E-04	1.82E-07	8.58E-07	6.99E-08	1.06E-08	10.3	-43.1
10N-20E-04L01	USGS	09-07-01	17.3	3.69E-04	2.06E-07	5.29E-07	8.29E-08	1.22E-08	4.49	-9.87
10N-20E-09A01	USGS	09-07-01	17.9	3.76E-04	2.11E-07	2.90E-07	8.30E-08	1.24E-08	<.01	-53.0
10N-20E-09A02	USGS	09-07-01	18.2	3.90E-04	2.19E-07	5.15E-08	8.63E-08	1.27E-08	2.1	–
10N-22E-25F03	USGS	02-27-00	8.70	–	–	–	–	–	16.0	–
11N-18E-26M03	USGS	09-06-01	16.9	3.57E-04	2.03E-07	1.45E-06	8.42E-08	1.20E-08	8.17	-52.3
11N-18E-27H01	USGS	02-12-01	21.5	–	–	–	–	–	–	–
11N-21E-34F01	USGS	09-12-01	20.5	3.66E-04	3.04E-07	7.47E-08	7.74E-08	1.12E-08	–	–
11N-22E-30C01	USGS	09-06-01	2.10	–	–	–	–	–	38.3	-35.6
11N-22E-33M01	USGS	09-12-01	18.4	3.47E-04	2.09E-07	1.11E-07	7.70E-08	1.17E-08	.04	–
12N-20E-12K01	USGS	08-24-05	8.20	–	–	–	–	–	25.7	–
13N-20E-34Q01D1	USGS	08-24-05	11.1	–	–	–	–	–	16.5	–
13N-27E-26H01	USGS	09-12-01	20.5	3.77E-04	2.22E-07	1.60E-05	8.08E-08	1.16E-08	.10	–
13N-20E-33N01	USGS	10-01-04	15.6	–	–	–	–	–	8.70	–
11N-19E-06N	Yakama Nation	02-28-91	–	–	–	–	–	–	–	–
11N-19E-34A01	Yakama Nation	02-28-91	–	–	–	–	–	–	–	–
11N-19E-34B01A	Yakama Nation	04-10-91	–	–	–	–	–	–	–	–
11N-18E-03E	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
11N-18E-20R02A	Yakama Nation	04-15-91	–	–	–	–	–	–	–	–
11N-18E-20R02B	Yakama Nation	04-15-91	–	–	–	–	–	–	–	–
11N-17E-23D02	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
11N-16E-23B02	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

Well No.	Collection agency	Date	Hydrogeo-logic unit	Land-surface altitude (feet above NGVD 29)	Hole depth (ft)	Well depth (ft)	Tempe-rature (°C)	Specific conduc-tance (µS/cm)	pH	Dissolved oxygen (mg/L)	Total dissolved solids	Turbidity (NTU)	Carbonate (mg/L)	Bicar-bonate (mg/L)
10N-21E-19P01	Yakama Nation	02-28-91	UNCON	705	52	47	16.1	258	7.4	0.8	100	0.2	<1.0	32.0
10N-21E-28F01	Yakama Nation	02-28-91	UNCON	735	57	52	13.8	349	7.0	5.4	106	.2	<1.0	34.0
10N-18E-09N04	Yakama Nation	04-16-91	UNCON	830	94	84	14.7	536	7.3	2.2	388	2.5	<1.0	351
10N-17E-05L	Yakama Nation	03-06-91	UNCON	970	59	54.5	12.4	170	6.6	8.7	202	.3	<1.0	94.0
10N-17E-17N01	Yakama Nation	03-06-91	UNCON	1,055	48	46.5	12.1	124	6.8	9.1	150	.2	<1.0	70.0
12N-19E-30P01	Yakama Nation	04-16-91	CN	935	160	156	14.0	218	7.4	2.6	216	3.6	<1.0	155
11N-19E-10Q01	Yakama Nation	12-06-90	CN	855	765	635	18.2	119	8.1	2.4	190	.3	<1.0	63.0
11N-19E-34B-9B	Yakama Nation	04-10-91	CN	815	404	389	16.1	98	8.4	4.4	93.0	1.5	<1.0	59.0
11N-18E-03D	Yakama Nation	03-26-91	CN	975	158	151	16.4	205	7.5	<.1	138	.3	<1.0	107
11N-18E-13N01	Yakama Nation	04-15-91	CN	860	127	96	13.3	390	7.3	6.0	360	7.8	<1.0	193
11N-17E-02L2E	Yakama Nation	11-08-90	CN	1,060	270	200	17.9	268	7.3	6.3	152	.2	<1.0	182
11N-17E-16R02	Yakama Nation	09-18-90	CN	963	200	195	14.1	431	7.4	6.8	362	.2	<1.0	201
11N-17E-30D01	Yakama Nation	04-10-91	CN	1,100	302	290	16.9	218	7.1	4.8	178	19	<1.0	122
11N-16E-21J	Yakama Nation	04-16-91	CN	1,320	345	264	17.4	175	7.9	1.1	172	7.8	<1.0	102
10N-21E-32K02	Yakama Nation	12-06-90	CN	700	174	166	16.5	300	8.1	<.1	280	.3	<1.0	183
10N-20E-04L01	Yakama Nation	04-15-91	CN	764	1,022	809	22.2	214	7.6	.2	247	.3	<1.0	143
10N-19E-30Q01	Yakama Nation	09-12-90	CN	803	715	535	23.0	377	7.2	2.1	286	.5	<1.0	237
10N-19E-34K01	Yakama Nation	12-06-90	CN	850	219	212	16.2	427	7.4	3.9	233	.5	<1.0	287
10N-18E-04A03	Yakama Nation	04-10-91	CN	810	150	135	15.3	441	7.4	<.1	298	3.8	<1.0	281
10N-18E-04A07	Yakama Nation	04-12-91	CN	810	267	257	16.2	475	7.4	<.1	289	5.6	<1.0	310
10N-18E-09N03	Yakama Nation	04-16-91	CN	830	310	297	16.9	427	7.4	.6	334	.4	<1.0	268
10N-18E-25E01	Yakama Nation	04-16-91	CN	780	150	102	12.5	341	7.8	1.3	230	.4	<1.0	210
10N-17E-05E04	Yakama Nation	03-06-91	CN	960	395	300	12.9	165	7.6	<.1	220	1.0	<1.0	100
10N-17E-21A01	Yakama Nation	03-26-91	CN-SM	985	220	198	12.5	163	7.5	<.1	110	.3	<1.0	99.0
12N-18E-26C01	Yakama Nation	04-16-91	SM	1,140	425	213	25.0	170	7.9	< .1	210	.2	<1.0	178
12N-18E-27N02D1	Yakama Nation	09-30-90	SM-WN	1,120	1,009	690	27.1	287	7.8	–	213	2.2	<1.0	162
11N-21E-16P01	Yakama Nation	04-10-91	SM-WN	1,300	1,402	700	29.4	294	8.7	.1	244	1.3	8.4	141
11N-18E-26M03	Yakama Nation	04-15-91	SM-WN	835	1,471	1,270	28.6	348	7.4	< .1	378	.3	<1.0	223
11N-17E-02P01	Yakama Nation	03-06-91	WN	1,100	1,000	870	26.2	269	7.8	-1.3	229	1.0	<1.0	168
11N-17E-21C01	Yakama Nation	03-06-91	SM	980	640	571	19.4	156	7.8	5.2	246	1.4	<1.0	211
11N-16E-09P01	Yakama Nation	03-06-91	WN	1,480	185	143	15.4	217	7.0	5.2	209	1.4	<1.0	121

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

Well No.	Collection agency	Date	Alkalinity (mg/L)	Potassium (mg/L)	Manga-nese (mg/L)	Magne-sium (mg/L)	Calcium (mg/L)	Boron (mg/L)	Phosphate (dissolved) (mg/L)	Total phos-phorus (mg/L)	Nitrite (mg/L)	Nitrite + nitrate as N (mg/L)	Ammonia as N (mg/L)	Sulphate (mg/L)
10N-21E-19P01	Yakama Nation	02-28-91	26.0	6.00	0.11	5.60	17.0	<0.1	<0.05	<0.05	<0.002	<0.05	<0.07	0.33
10N-21E-28F01	Yakama Nation	02-28-91	28.0	3.00	<.01	14.0	35.0	<.1	<.05	.06	<.002	6.10	<.07	15.0
10N-18E-09N04	Yakama Nation	04-16-91	288	6.00	.12	23.0	47.0	.1	.07	.13	<.002	<.05	.56	5.60
10N-17E-05L	Yakama Nation	03-06-91	77.0	3.00	<.01	6.90	14.0	<.1	.08	.07	<.002	.70	<.07	4.50
10N-17E-17N01	Yakama Nation	03-06-91	57.0	2.00	<.01	5.40	10.0	<.1	<.05	.05	<.002	.60	<.07	2.90
12N-19E-30P01	Yakama Nation	04-16-91	127	5.00	.35	17.0	39.0	<.1	<.05	<.05	<.002	<.05	<.07	53.5
11N-19E-10Q01	Yakama Nation	12-06-90	52.0	2.00	<.01	1.10	10.0	<.1	<.05	<.05	<.002	.07	<.07	1.92
11N-19E-34B-9B	Yakama Nation	04-10-91	48.0	1.00	.01	0.59	9.90	<.1	<.05	.06	<.002	.08	.49	1.96
11N-18E-03D	Yakama Nation	03-26-91	88.0	4.00	.14	5.70	16.0	<.1	<.05	<.05	<.002	<.05	<.07	10.7
11N-18E-13N01	Yakama Nation	04-15-91	158	4.00	.04	10.0	37.0	<.1	.06	.06	<.002	3.70	<.07	22.3
11N-17E-02L2E*	Yakama Nation	11-08-90	149	4.00	<.01	11.0	24.0	<.1	.05	.06	<.002	<.05	<.07	8.50
11N-17E-16R02	Yakama Nation	09-18-90	165	2.00	<.01	16.0	35.0	<.1	<.05	<.05	<.002	4.90	<.07	22.5
11N-17E-30D01	Yakama Nation	04-10-91	100	3.00	.05	9.70	17.0	<.1	.28	.38	<.002	.50	<.07	7.40
11N-16E-21J	Yakama Nation	04-16-91	84.0	4.00	.02	6.10	14.0	<.1	<.05	.05	<.002	.50	<.07	5.30
10N-21E-32K02	Yakama Nation	12-06-90	150	9.00	.08	3.90	13.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
10N-20E-04L01	Yakama Nation	04-15-91	117	4.70	.12	5.20	13.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
10N-19E-30Q01	Yakama Nation	09-12-90	194	6.00	.31	14.0	26.0	<.1	<.05	.06	<.002	<.05	<.07	.36
10N-19E-34K01	Yakama Nation	12-06-90	235	2.00	<.01	16.0	43.0	<.1	<.05	<.05	<.002	.10	<.07	8.90
10N-18E-04A03	Yakama Nation	04-10-91	230	5.00	.21	14.0	37.0	.1	<.05	<.05	<.002	<.05	<.07	.57
10N-18E-04A07	Yakama Nation	04-12-91	254	5.00	.15	16.0	42.0	.1	<.05	<.05	<.002	<.05	<.07	<.25
10N-18E-09N03	Yakama Nation	04-16-91	220	5.00	.04	12.0	18.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
10N-18E-25E01	Yakama Nation	04-16-91	172	4.00	.12	15.0	30.0	<.1	.32	.30	<.002	<.05	.42	<.25
10N-17E-05E04	Yakama Nation	03-06-91	82.0	3.00	<.01	6.90	13.0	<.1	.11	.09	<.002	.09	<.07	2.12
10N-17E-21A01	Yakama Nation	03-26-91	81.0	1.00	.13	7.10	14.0	<.1	.09	.09	<.002	<.05	<.07	1.75
12N-18E-26C01	Yakama Nation	04-16-91	146	6.00	.04	8.50	17.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
12N-18E-27N02D1	Yakama Nation	09-30-90	133	5.00	.03	8.90	18.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
11N-21E-16P01	Yakama Nation	04-10-91	138	6.00	<.01	0.16	1.70	<.1	<.05	<.05	<.002	–	<.07	–
11N-18E-26M03	Yakama Nation	04-15-91	183	8.00	.10	11.0	21.0	.1	<.05	.50	<.002	<.05	<.07	<.25
11N-17E-02P01	Yakama Nation	03-06-91	138	5.00	.04	8.30	17.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
11N-17E-21C01	Yakama Nation	03-06-91	173	4.00	.07	10.0	18.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
11N-16E-09P01	Yakama Nation	03-06-91	99.0	4.00	<.01	11.0	15.0	<.1	.06	.11	<.002	1.03	<.07	5.01

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

Well No.	Collection agency	Date	Chloride (mg/L)	Sodium (mg/L)	SiO_2 (mg/L)	Fluoride (mg/L)	Dissolved organic carbon (mg/L)	$\delta^{18}O$ (per mil)	δD (per mil)	Tritium (TU)	^{14}C DIC (PMC)	$\delta^{13}C$ DIC (per mil)	Hydrogen sulfide (mg/L)	Oxidation reduction potential (millivolts)
10N-21E-19P01	Yakama Nation	02-28-91	3.40	28.0	66.0	0.50	0.15	-17.1	-131	1.40	–	–	–	–
10N-21E-28F01	Yakama Nation	02-28-91	5.60	12.0	33.0	<.20	.34	-14.2	-104	18.1	–	–	–	–
10N-18E-09N04	Yakama Nation	04-16-91	7.30	33.0	75.0	.60	1.5	-15.5	-114	2.50	–	–	–	–
10N-17E-05L	Yakama Nation	03-06-91	1.80	8.10	46.0	<.20	.40	-15.1	-111	9.30	–	–	–	–
10N-17E-17N01	Yakama Nation	03-06-91	1.00	5.00	38.0	.20	<.10	-15.3	-111	10.8	–	–	–	–
12N-19E-30P01	Yakama Nation	04-16-91	17.6	20.0	48.0	.20	.75	-14.7	-109	2.60	85.6	-13.6	–	–
11N-19E-10Q01	Yakama Nation	12-06-90	1.10	11.0	36.0	.20	.34	-15.1	-111	.64	–	-13.0	–	–
11N-19E-34B-9B	Yakama Nation	04-10-91	1.00	9.00	23.0	<.20	<.10	-15.0	-108	<.1	–	–	–	–
11N-18E-03D	Yakama Nation	03-26-91	2.80	16.0	63.0	.30	<.10	-17.7	-136	.60	–	–	–	–
11N-18E-13N01	Yakama Nation	04-15-91	7.90	23.0	36.0	<.20	.70	-14.5	-106	18.9	–	–	–	–
11N-17E-02L2E*	Yakama Nation	11-08-90	3.40	15.0	71.0	.40	.05	-16.2	-127	<.1	52.4	-14.5	–	–
11N-17E-16R02	Yakama Nation	09-18-90	17.5	26.0	52.0	1.20	1.0	-14.2	-102	10.0	83.8	-13.0	–	–
11N-17E-30D01	Yakama Nation	04-10-91	3.40	12.0	56.0	.30	.20	-15.0	-120	<.1	–	–	–	–
11N-16E-21J	Yakama Nation	04-16-91	2.00	9.50	35.0	.30	<.10	-15.7	-123	<.1	36.4	-14.3	–	–
10N-21E-32K02	Yakama Nation	12-06-90	4.40	39.0	56.0	.90	.67	-17.2	-132	<.1	26.4	-12.4	–	–
10N-20E-04L01	Yakama Nation	04-15-91	2.50	21.0	65.0	.60	.31	-16.7	-127	.92	17.1	-10.0	–	–
10N-19E-30Q01	Yakama Nation	09-12-90	4.30	29.0	74.0	.60	.44	-17.8	-137	<.1	5.38	-7.80	–	–
10N-19E-34K01	Yakama Nation	12-06-90	3.00	19.0	43.0	<.20	.28	-15.8	-120	<.1	81.5	-13.8	–	–
10N-18E-04A03	Yakama Nation	04-10-91	4.20	35.0	57.0	.60	.80	-18.6	-138	1.40	–	--	–	–
10N-18E-04A07	Yakama Nation	04-12-91	4.90	34.0	65.0	.60	.30	-17.8	-134	<.1	4.24	-7.10	–	–
10N-18E-09N03	Yakama Nation	04-16-91	5.20	55.0	79.0	.80	1.1	-17.5	-131	<.1	–	–	–	–
10N-18E-25E01	Yakama Nation	04-16-91	3.80	18.0	52.0	.40	.70	-16.1	-120	1.00	–	–	–	–
10N-17E-05E04	Yakama Nation	03-06-91	1.09	8.60	51.0	.20	<.10	-15.8	-120	<.1	47.2	-14.6	–	–
10N-17E-21A01	Yakama Nation	03-26-91	1.33	9.00	62.0	.30	<.10	-16.0	-121	<.1	–	–	–	–
12N-18E-26C01	Yakama Nation	04-16-91	9.50	32.0	59.0	.90	<.10	-17.2	-133	<.1	11.5	-8.40	–	–
12N-18E-27N02D1	Yakama Nation	09-30-90	5.60	22.0	60.0	.70	3.0	-17.2	-135	<.1	10.8	-9.60	–	–
11N-21E-16P01	Yakama Nation	04-10-91	–	58.0	49.0	–	4.0	-18.8	-146	<.1	3.39	-4.50	–	–
11N-18E-26M03	Yakama Nation	04-15-91	4.60	32.0	81.0	.90	.36	-18.3	-141	<.1	6.71	-9.60	–	–
11N-17E-02P01	Yakama Nation	03-06-91	5.22	29.0	68.0	.80	.43	-17.1	-135	<.1	11.0	-12.6	–	–
11N-17E-21C01	Yakama Nation	03-06-91	4.31	25.0	70.0	.70	.48	-17.1	-135	<.1	16.9	-13.1	–	–
11N-16E-09P01	Yakama Nation	03-06-91	2.26	11.0	60.0	.30	.30	-15.8	-123	1.50	83.4	-14.0	–	–

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

Well No.	Collection agency	Date	Nitrogen (mg/L)	Argon (ccSTP/g)	Neon (ccSTP/g)	Hellium (ccSTP/g)	Krypton (ccSTP/g)	Xenon (ccSTP/g)	Methane (mg/L)	δ13C-CH4 (per mil)
10N-21E-19P01	Yakama Nation	02-28-91	–	–	–	–	–	–	–	–
10N-21E-28F01	Yakama Nation	02-28-91	–	–	–	–	–	–	–	–
10N-18E-09N04	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
10N-17E-05L	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
10N-17E-17N01	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
12N-19E-30P01	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
11N-19E-10Q01	Yakama Nation	12-06-90	–	–	–	–	–	–	–	–
11N-19E-34B-9B	Yakama Nation	04-10-91	–	–	–	–	–	–	–	–
11N-18E-03D	Yakama Nation	03-26-91	–	–	–	–	–	–	–	–
11N-18E-13N01	Yakama Nation	04-15-91	–	–	–	–	–	–	–	–
11N-17E-02L2E*	Yakama Nation	11-08-90	–	–	–	–	–	–	–	–
11N-17E-16R02	Yakama Nation	09-18-90	–	–	–	–	–	–	–	–
11N-17E-30D01	Yakama Nation	04-10-91	–	–	–	–	–	–	–	–
11N-16E-21J	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
10N-21E-32K02	Yakama Nation	12-06-90	–	–	–	–	–	–	–	–
10N-20E-04L01	Yakama Nation	04-15-91	–	–	–	–	–	–	–	–
10N-19E-30Q01	Yakama Nation	09-12-90	–	–	–	–	–	–	–	–
10N-19E-34K01	Yakama Nation	12-06-90	–	–	–	–	–	–	–	–
10N-18E-04A03	Yakama Nation	04-10-91	–	–	–	–	–	–	–	–
10N-18E-04A07	Yakama Nation	04-12-91	–	–	–	–	–	–	–	–
10N-18E-09N03	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
10N-18E-25E01	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
10N-17E-05E04	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
10N-17E-21A01	Yakama Nation	03-26-91	–	–	–	–	–	–	–	–
12N-18E-26C01	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
12N-18E-27N02D1	Yakama Nation	09-30-90	–	–	–	–	–	–	–	–
11N-21E-16P01	Yakama Nation	04-10-91	–	–	–	–	–	–	–	–
11N-18E-26M03	Yakama Nation	04-15-91	–	–	–	–	–	–	–	–
11N-17E-02P01	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
11N-17E-21C01	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–
11N-16E-09P01	Yakama Nation	03-06-91	–	–	–	–	–	–	–	–

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

Well No.	Collection agency	Date	Hydrogeo-logic unit	Land-surface altitude (feet above NGVD 29)	Hole depth (ft)	Well depth (ft)	Tempe-rature (°C)	Specific conduc-tance (µS/cm)	pH	Dissolved oxygen (mg/L)	Total dissolved solids	Turbidity (NTU)	Carbonate (mg/L)	Bicar-bonate (mg/L)
11N-16E-28F02	Yakama Nation	09-18-90	WN	1,280	180	135	14.5	174	7.2	5.0	97.0	1.2	<1.0	109
11N-16E-34K03	Yakama Nation	09-18-90	GR	1,200	429	409	22.3	411	6.2	< .1	223	12	<1.0	270
10N-21E-03H01	Yakama Nation	04-16-91	SM	910	775	724	23.0	249	7.8	7.0	239	.3	<1.0	231
10N-21E-21J01	Yakama Nation	12-06-90	SM	740	554	535	22.5	314	7.9	< .1	249	.3	<1.0	184
10N-17E-06J	Yakama Nation	04-15-91	GR	1,306	1,306	1,072	20.0	525	6.8	.8	518	13	<1.0	376
10N-17E-08N03	Yakama Nation	09-12-90	WN	1,020	288	285	14.3	135	7.4	8.8	109	2.0	<1.0	90.0
10N-17E-26J01	Yakama Nation	09-12-90	GR	1,000	1,000	956	24.4	391	7.4	4.6	288	.2	<1.0	250

Well No.	Collection agency	Date	Alkalinity (mg/L)	Potassium (mg/L)	Manga-nese (mg/L)	Magne-sium (mg/L)	Calcium (mg/L)	Boron (mg/L)	Phosphate (dissolved) (mg/L)	Total phos-phorus (mg/L)	Nitrite (mg/L)	Nitrite + nitrate as N (mg/L)	Ammonia as N (mg/L)	Sulphate (mg/L)
11N-16E-28F02	Yakama Nation	09-18-90	89.0	3.00	<0.01	8.30	14.0	<0.1	0.05	0.05	<0.002	.46	<0.07	2.40
11N-16E-34K03	Yakama Nation	09-18-90	221	6.00	.07	16.0	21.0	<.1	<.05	<.05	<.002	<0.05	<.07	<.25
10N-21E-03H01	Yakama Nation	04-16-91	189	7.70	.07	8.40	36.0	<.1	<.05	<.05	<.002	<.05	<.07	2.62
10N-21E-21J01	Yakama Nation	12-06-90	151	6.40	.06	5.30	24.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25
10N-17E-06J	Yakama Nation	04-15-91	308	7.00	.08	29.0	36.0	.1	<.05	<.05	<.002	<.05	<.07	<.25
10N-17E-08N03	Yakama Nation	09-12-90	74.0	3.00	<.01	6.70	10.0	<.1	.09	.13	<.002	.11	<.07	.88
10N-17E-26J01	Yakama Nation	09-12-90	205	6.00	.02	16.0	22.0	<.1	<.05	<.05	<.002	<.05	<.07	<.25

Table B1. Hydrochemical data from groundwater samples, Yakima River basin aquifer system, Washington.—Continued

Well No.	Collection agency	Date	Chloride (mg/L)	Sodium (mg/L)	SiO_2 (mg/L)	Fluoride (mg/L)	Dissolved organic carbon (mg/L)	$\delta^{18}O$ (per mil)	δD (per mil)	Tritium (TU)	^{14}C DIC (PMC)	$\delta^{13}C$ DIC (per mil)	Hydrogen sulfide (mg/L)	Oxidation reduction potential (millivolts)
11N-16E-28F02	Yakama Nation	09-18-90	1.26	7.30	47.0	0.20	0.18	-15.9	-122	-0.70	86.3	-12.7	–	–
11N-16E-34K03	Yakama Nation	09-18-90	3.26	40.0	99.0	.40	.27	-15.8	-122	-.70	13.0	-7.30	–	–
10N-21E-03H01	Yakama Nation	04-16-91	4.70	25.0	37.0	.40	<.10	-19.0	-147	-.70	6.95	-13.1	–	–
10N-21E-21J01	Yakama Nation	12-06-90	2.70	29.0	39.0	.40	.45	-19.2	-150	-.62	9.72	-13.1	–	–
10N-17E-06J	Yakama Nation	04-15-91	7.10	41.0	90.0	.60	.41	-16.9	-128	-.73	6.01	-6.20	–	–
10N-17E-08N03	Yakama Nation	09-12-90	.70	5.50	45.0	.20	<.10	-15.8	-115	-.90	--	-13.3	–	–
10N-17E-26J01	Yakama Nation	09-12-90	4.10	36.0	83.0	.90	1.0	-17.0	-132	-.70	5.48	-5.70	–	–

Well No.	Collection agency	Date	Nitrogen (mg/L)	Argon (ccSTP/g)	Neon (ccSTP/g)	Hellium (ccSTP/g)	Krypton (ccSTP/g)	Xenon (ccSTP/g)	Methane (mg/L)	$\delta^{13}C-CH4$ (per mil)
11N-16E-28F02	Yakama Nation	09-18-90	–	–	–	–	–	–	–	–
11N-16E-34K03	Yakama Nation	09-18-90	–	–	–	–	–	–	–	–
10N-21E-03H01	Yakama Nation	04-16-91	–	–	–	–	–	–	–	–
10N-21E-21J01	Yakama Nation	12-06-90	–	–	–	–	–	–	–	–
10N-17E-06J	Yakama Nation	04-15-91	–	–	–	–	–	–	–	–
10N-17E-08N03	Yakama Nation	09-12-90	–	–	–	–	–	–	–	–
10N-17E-26J01	Yakama Nation	09-12-90	–	–	–	–	–	–	–	–

Appendix C. Methods Used to Estimate Groundwater Recharge for the Area Outside the Yakima River Basin in the Extended Study Area

Groundwater recharge was estimated for nearly the entire aquifer system within the Yakima River basin as part of this study (Vaccaro and Olsen, 2007a). Estimates of recharge were also needed for the area outside the boundary of the basin that is within the extended study area and for about 108 mi^2 within the basin where recharge was not estimated. These areas are scheduled to be included in a regional groundwater flow model being constructed as part of this study. Two methods were used to estimate mean annual and monthly recharge for a base period of water years 1960–2001. One method was used for selected areas where estimated mean annual recharge values for the period 1956–77 were available. These values were estimated by Bauer and Vaccaro (1990) using the Deep Percolation Model (DPM). The other method, using mean annual precipitation (MAP) as a predictor of mean annual recharge, was used for the remaining area.

Recharge estimates for the two areas needed to be referenced to approximately the same base period. Estimated recharge values for one of the modeled areas of Vaccaro and Olsen (2007a) were selected as the base-period (1960–2001) reference values based on the area's proximity, mean annual recharge values, and land-use/land-cover. The modeled area, called Cold-Dry, included 321.3 mi^2 and had a mean annual recharge estimate of 0.44 in. for predevelopment land-use/land-cover conditions for the period 1950–2003 (Vaccaro and Olsen, 2007a). The ratio of mean annual recharge for the period 1960–2001 to that for the period 1956–1977 for Cold-Dry was calculated as 0.934; the ratio is less than 1.0 because of the general trend of declining precipitation starting in 1977 (Vaccaro, 1995, 2002). This is the adjustment factor used to multiply the mean annual estimates of Bauer and Vaccaro (1990) to obtain consistent, mean annual base-period values. The precipitation-derived mean annual recharge values were multiplied by an adjustment factor of 0.858 to make them consistent with the 1960–2001 base period; the adjustment factor was the ratio of Cold-Dry mean annual recharge for 1960–2001 to that for 1971–2000, which is the climate normal for the MAP.

Next, for each water year, the ratio of the annual value to the 1960–2001 mean annual value for Cold-Dry was calculated. This ratio is the factor (F) used to adjust the base-period referenced mean annual values of Bauer and Vaccaro (1990) to water year values. Last, the monthly percent of annual recharge for water years 1960–2001 was calculated for Cold-Dry. These 42 water years of monthly factors are used to estimate the monthly values using the adjusted mean annual values to water year values. For example, to estimate recharge for month 1 (October) of water year 1960 for a cell in a modeled area:

$$\text{Mon1, 1960} = ((\text{BVMA})*0.934)*\text{F60})*\text{Mon\%Ann}\ , \qquad (1)$$

where

Mon1, 1960 is estimated recharge for month 1 of 1960;

BVMA is mean annual recharge value for a model cell of Bauer and Vaccaro (1990);

F60 is the ratio of Cold-Dry annual recharge for 1960 to its 1960-2001 mean annual value, and

Mon%Ann is month 1 percent of annual for 1960 for Cold-Dry.

Model-calculated mean annual recharge values for three modeled areas (zones 4, 8, and 14 of Bauer and Vaccaro [1990]) were selected for applying the first method. These zones covered 682 mi^2 and included 1,840 cells that had an approximate area of 0.375 mi^2. Of this total area, 385 mi^2 already had estimates of recharge available from Vaccaro and Olsen (2007a), so that recharge was needed only for the remaining 297 mi^2. A 500 ft raster grid was overlaid on the combined three areas, and an association between a raster grid cell outside of the areas modeled by Vaccaro and Olsen (2007a) and a cell in one of the three models was established. There were a total of 48,838 500-ft grid cells and there were 694 unique DPM cells for the grid cells. The spatial distribution of the mean annual recharge values were preserved because the DPM model cells of Bauer and Vaccaro (1990) were much larger than the raster grid cell size. The mean annual value for each of the 694 unique cells was then multiplied by 0.934 in order to obtain a spatial distribution of mean annual values adjusted to the base period. This method of relating recharge values (both monthly and annual) to a raster grid is similar to that used by Vaccaro and Olsen (2007b).

To obtain monthly values from the adjusted mean annual values, the values were first multiplied by the factor F, described above (for example, F60), for each water year. The resulting 42 values were then multiplied by the 12 Mon%Ann values for each month of a water year. These calculations yielded one ASCII file of estimated monthly recharge that is spatially related using the raster grid with its unique identification numbers.

Model-calculated estimates of recharge were not available for the remaining 511 mi^2. For this area, a polynomial relation between mean annual recharge and MAP of Bauer and Vaccaro (1990) was used. MAP was for the climate normal of 1971–2000, and is a raster grid with an 800-m size cell (Spatial Climate Analysis Service–Oregon State University, 2006). Mean annual recharge was calculated using the relation of Bauer and Vaccaro (1990) for each MAP cell in this area. These values were multiplied by the 0.858 factor to adjust them to the 1960–2001 base period. The mean annual recharge values were then made into an ASCII file that had the 800-m raster grid identification number and an associated mean annual recharge value. Using the same methods and information described above, one ASCII file containing monthly values for each water year was then generated.

The 500-ft raster grid described in Vaccaro and Olsen (2007b) was then overlaid on the two raster grids described above in order to assign a unique identifier to the grid. Thus, a single 500 ft raster grid was developed for the extended study area that contains unique identification numbers that can be directly related to ASCII files containing monthly recharge estimates.

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Appendix D. Estimation of Mean Annual Values and Development of Time-Series of Monthly Values for Surface-Water Diversions, Yakima River Basin, Washington

Monthly values for 65 surface-water diversions within the Yakima River basin were obtained from the Bureau of Reclamation (R. Larson, written commun., 2007 and R. Sonnichsen, written commun., 2007; and Hydromet— http://www.usbr.gov/pn/hydromet/yakima/yakwebarcread.html). Monthly observed or estimated values for the base period of water years 1960–2001 were available for Kittitas, Sunnyside, Tieton, Chandler, New Reservation, Kennewick, Roza, Naches-Selah, South Naches, and Westside canals. For most of the other 54 diversions, monthly observed or estimated information was available only through water year 1999 for the information provided by Larson and through 2001 for the information provided by Sonnichsen. However, the latter information started in water year 1981. In order to extend the records of Larson through 2001, diversions in 2000 were assumed to be the same as those in 1998. Both 1998 and 2000 had nearly the same Total Water Supply Available (forecasted water supply for the irrigation season) and the storage control date was the same. For diversions in 2001, it was assumed that those diversions were the same as those in 1994, which had the same April 1 prorating percent of 37 (junior surface-water users obtained only 37 percent of their entitlement). As an indication of the accuracy of these estimates, for canals with records through 2001 the mean annual diversion in 1994 was as much as 12 percent smaller and as much as 8 percent higher than 2001. In addition, monthly values for the two diversions of the Cascade Irrigation District (CID) were only through 1991 and 1994. The remaining monthly values were provided by CID (Cascade Irrigation District, written commun., 2007).

The mean annual values were then calculated for 49 canals (a subset of the 54 described above) common to both Reclamation data sets. The values were compared to each other and to their percent of the entitlement. It was determined that for 28 diversions the monthly values may have been estimated as either too large or too small. Based on a ratio of mean annual values from the two sources, the monthly values for 1960–2001 were modified. This resulted in a net increase of 68 ft^3/s or about 10 percent of the total mean annual diversion for the 49 canals. The 68 ft^3/s is only about 1 percent of the total for the 65 diversions. The resulting total mean annual value for the 65 diversions is about 86 percent of the total entitlement for these canals.

Most of the remaining surface-water diversion quantities were obtained from a file provided by WaDOE (B. Johnson, Washington State Department of Ecology, written commun., 2007). For each diversion, the file contained its maximum allowable instantaneous rate, annual entitlement, and in most cases, a name and location. The information provided by WaDOE contained no measured or estimated monthly or annual diversion quantities. First, diversions identified as springs or spring-fed creeks were not included in this analysis. However, these springs are presented on a map of the springs in the basin in this report. Next, diversions with small annual entitlements (generally less than 5 acre-ft) and diversions without location information also were not included; together these accounted for about 6 percent of the total annual entitlement of all of the diversions in the file. Last, diversions that were part of the above 65 were excluded. Except for the Yakima and Naches Rivers, diversions were then generally aggregated to one or several locations for each stream based on the locations of the largest entitlements. This aggregation resulted in 70 points of withdrawals. The full annual entitlement value was used as the estimate of the mean annual diversion (B. Johnson, Washington State Department of Ecology, written commun., 2007), and thus, was assumed to be constant for every year. Information for eight other diversions in the Toppenish basin was obtained from the Yakama Nation (D. Lind, Yakama Nation, written commun., 2007). The information included location and the estimated mean annual and maximum diversion quantities. Similar to a diversion in the Ahtanum subbasin, seven of the eight diversions are shut off in the summer to maintain instream flows (D. Lind, Yakama Nation, written commun., 2007). Based on the average of the ratio of the mean monthly to the mean annual quantities for the 65 diversions, monthly multiplying factors were calculated for distributing the final 78 mean annual values over the April through October irrigation season. These factors ranged from 0.68 in October to 2.17 in July. The monthly estimates for these 78 points and the 65 other diversions are needed for other parts of this study and provide a comparison with the groundwater pumpage in the basin.

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Director, Washington Water Science Center
U.S. Geological Survey, 934 Broadway — Suite 300
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