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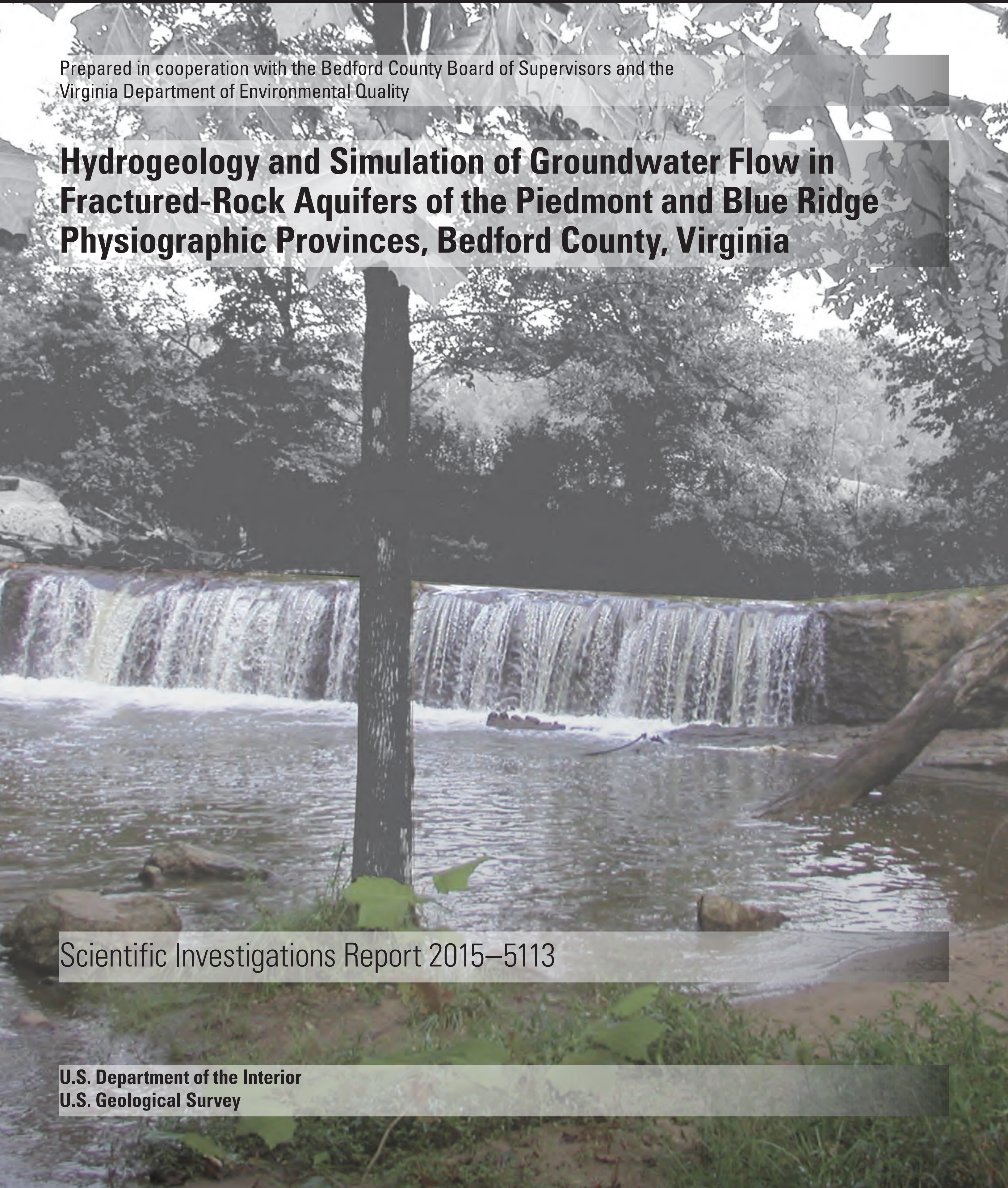
Hydrogeology and Simulation of Groundwater Flow in Fractured-Rock Aquifers of the Piedmont and Blue Ridge Physiographic Provinces, Bedford County, Virginia



Scientific Investigations Report 2015–5113

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

Front cover: Small waterfall on North Otter River in Bedford County, Virginia. Photograph by Don Hayes, U.S. Geological Survey.

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By Kurt J. McCoy, Bradley A. White, Richard M. Yager, and George E. Harlow, Jr.

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SALLY JEWELL, Secretary

U.S. Geological Survey
Suzette M. Kimball, Acting Director

U.S. Geological Survey, Reston, Virginia: 2015

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Conversion Factors and Datums

Inch/Pound to SI

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	
square mile (mi^2)	259.0	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
gallon (gal)	0.003785	cubic meter (m^3)
gallon (gal)	3.785	cubic decimeter (dm^3)
million gallons (Mgal)	3,785	cubic meter (m^3)
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
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)
gallon per day (gal/d)	0.003785	cubic meter per day (m^3/d)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Pressure	
atmosphere, standard (atm)	101.3	kilopascal (kPa)
bar	100	kilopascal (kPa)
pound per square inch (lb/in^2)	6.895	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×°C)+32

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:
°C=(°F–32)/1.8

*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.

Datum

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD 29).

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

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

Acknowledgments

The authors gratefully acknowledge the many individuals, establishments, and authorities in the county that permitted access to their wells to measure water levels during this study. The authors also thank the property owners that granted access to their property for collection of streamflow data. Special thanks are given to Frank J. Rogers, Deputy County Administrator of Bedford County, for his efforts with project logistics and support. Todd Fowler of the Bedford County Health Department provided well completion information and guidance in regards to the distribution of well yields.

Hydrogeology and Simulation of Groundwater Flow in Fractured-Rock Aquifers of the Piedmont and Blue Ridge Physiographic Provinces, Bedford County, Virginia

By Kurt J. McCoy,[1] Bradley A. White,[2] Richard M. Yager,[1] and George E. Harlow, Jr.[1]

Abstract

An annual groundwater budget was computed as part of a hydrogeologic characterization and monitoring effort of fractured-rock aquifers in Bedford County, Virginia, a growing 764-square-mile (mi^2) rural area between the cities of Roanoke and Lynchburg, Virginia. Data collection in Bedford County began in the 1930s when continuous stream gages were installed on Goose Creek and Big Otter River, the two major tributaries of the Roanoke River within the county. Between 2006 and 2014, an additional 2 stream gages, 3 groundwater monitoring wells, and 12 partial-record stream gages were operated. Hydrograph separation methods were used to compute base-flow recharge rates from the continuous data collected from the continuous stream gages. Mean annual base-flow recharge ranged from 8.3 inches per year (in/yr) for the period 1931–2012 at Goose Creek near Huddleston (drainage area 188 mi^2) to 9.3 in/yr for the period 1938–2012 at Big Otter River near Evington (drainage area 315 mi^2). Mean annual base-flow recharge was estimated to be 6.5 in/yr for the period 2007–2012 at Goose Creek at Route 747 near Bunker Hill (drainage area 125 mi^2) and 8.9 in/yr for the period 2007–2012 at Big Otter River at Route 221 near Bedford (drainage area 114 mi^2). Base-flow recharge computed from the partial-record data ranged from 5.0 in/yr in the headwaters of Goose Creek to 10.5 in/yr in the headwaters of Big Otter River.

A steady-state groundwater-flow simulation for Bedford County was developed to test the conceptual understanding of flow in the fractured-rock aquifers and to compute a groundwater budget for the four major drainages: James River, Smith Mountain and Leesville Lakes, Goose Creek, and Big Otter River. Model results indicate that groundwater levels mimic topography and that minimal differences in aquifer properties exist between the Proterozoic basement crystalline rocks and Late Proterozoic-Cambrian cover crystalline rocks. The Big Otter River receives 40.8 percent of the total daily groundwater outflow from fractured-rock aquifers in Bedford County; Goose Creek receives 25.8 percent, the James River receives 18.2 percent, and Smith Mountain and Leesville Lakes receive 15.2 percent. The remaining percentage of outflow is attributed to pumping from the aquifer (consumptive use).

[1]U.S. Geological Survey.

[2]Virginia Department of Environmental Quality.

Introduction

Groundwater resources in Bedford County, Virginia (Va.), are increasingly relied upon to supply water to local communities, industry, and individual residences. Groundwater withdrawals from fractured-rock aquifers are the primary source of water for most rural households and the majority of the county's residents. Since 2003, more than 2,000 new wells have been permitted and drilled in Bedford County to meet the needs of individual residences (T.R. Fowler, Bedford County Health Department, oral commun., 2012). The area has a growing rural population which has expanded from approximately 38,300 residents in 1985 to 68,700 residents in 2010 (Maupin and others, 2014). To meet future water needs of individual residences, additional domestic development of these bedrock aquifers is likely. Previous hydrologic work in rural areas of the central Piedmont and Blue Ridge Physiographic Provinces of Virginia is limited, and basic knowledge of aquifer systems in this area is needed to support the expanding economy and growing population of Bedford County.

From 2006 to 2014, the U.S. Geological Survey (USGS), in cooperation with the Bedford County Board of Supervisors and the Virginia Department of Environmental Quality (DEQ), collected hydrologic data in Bedford County to assess county-wide groundwater conditions and provide technical data and a scientific foundation that could be used as a basis for management and future planning of Bedford County water resources. A conceptual model of groundwater flow in Bedford County was developed based on (1) previous studies in the Piedmont and Blue Ridge fractured-rock aquifers, (2) compilation of existing data, and (3) results of new hydrologic data collected from wells and streams. Base-flow yields, general well construction information, and borehole logs were summarized to support conceptualization of geologic features controlling the occurrence of groundwater in the Piedmont and Blue Ridge fractured-rock aquifers of Bedford County. A numerical model simulating groundwater flow in the aquifers was constructed as a component of this investigation to evaluate the conceptual model and estimate steady-state groundwater budgets for areas within Bedford County that drain to the Big Otter River, Goose Creek, the James River, and Smith Mountain and Leesville Lakes.

Purpose and Scope

This report provides a description of the hydrogeology and groundwater availability of the fractured-rock aquifer systems in Bedford County, Va. The primary purpose of the data collection and groundwater-flow simulation conducted as part of this study in Bedford County is to provide hydrogeologic information that can be used to guide the development and management of these important water resources in context of long-term aquifer inflows and outflows. The scope of this study included (1) the drilling of three new bedrock monitoring wells; (2) establishment of a continuous and biannual groundwater-level network; (3) continuous and partial-record measurement of stream discharge in the Big Otter River and Goose Creek Basins; and (4) borehole geophysical logging of five wells in Bedford County. Well completion reports from local and State health departments and archival State and Federal records were synthesized to document the variability in well construction and yields among hydrogeologic units.

This report also documents the development of a numerical model to synthesize all currently available data and evaluate the conceptualization of groundwater flow in the fractured-rock aquifers of Bedford County at a scale of hundreds of square miles. Extrapolation of model results to smaller-scale domains would require more hydrogeologic detail than is currently (2015) available.

Description of Study Area

Bedford County encompasses 764 square miles (mi^2) in Virginia's Piedmont and Blue Ridge Physiographic Provinces (fig. 1), two physiographic regions that extend over much of the central portion of Virginia. The two physiographic regions are defined by large topographic differences. The Piedmont is characterized by rolling and hilly terrain while the Blue Ridge has much steeper slopes. The Piedmont in Bedford County ranges in elevation from 800 feet (ft) to 2,100 ft above sea level, while elevations in the Blue Ridge are as much as 4,000 ft. The county is bounded by the Blue Ridge Mountains on the west, the James River on the northeast, Smith Mountain and Leesville Lakes on the south, and Campbell County on the east. The county contains the headwaters of Goose Creek and Big Otter River, which are major tributaries to the Roanoke River.

Bedford County has a mild climate with an average annual precipitation of 45.6 inches per year (in/yr) and a mean maximum daily temperature of 67.4 degrees Fahrenheit (PRISM Climate Group, Oregon State University, 2014). Climate station data for Bedford County were obtained from the National Oceanic and Atmospheric Administration (NOAA) National Climate Data Center (National Oceanic and Atmospheric Administration, 2014). Available precipitation data within or near Bedford County for which periods of data were available for the current normal climatological period 1981–2010 included five sites (table 1; fig. 1). Two of the sites (Holcomb Rock and Lynchburg #2) are within the James River Basin, two of the sites (Bedford 4 NW and Bedford) are within the Big Otter River Basin, and the remaining site (Huddleston 4 SW) is within the Goose Creek Basin.

Mean annual precipitation for the climatological period 1981–2010 decreases in an easterly direction, ranging from 46.5 in/yr (Holcomb Rock) to 41.5 in/yr (Lynchburg #2) (fig. 1; table 1). Mean monthly precipitation amounts among

Table 1. NOAA climate stations in Bedford County, Virginia.

[Site locations shown in figure 1. Abbreviations: ft, feet above National Geodetic Vertical Datum of 1929; NOAA, National Oceanic and Atmospheric Administration; in/yr, inches per year; NAD 83, North American Datum of 1983]

Station identification number	Station name	Latitude (decimal degrees)	Longitude (decimal degrees)	Datum	Elevation (ft)	Operating agency	Period of record (calendar years)[1]	NOAA climatological period 1981–2010 mean annual precipitation (in/yr)
440561	Bedford 4 NW	37.380	–79.561	NAD 83	1,220	NOAA	1973–2014	44.2
440551	Bedford	37.348	–79.523	NAD 83	975	NOAA	1948–2006	45.1
444039	Holcomb Rock	37.544	–79.403	NAD 83	620	NOAA	1960–2014	46.5
444148	Huddleston 4 SW	37.126	–79.526	NAD 83	1,045	NOAA	1950–2014	42.5
445117	Lynchburg #2	37.385	–79.229	NAD 83	740	NOAA	1997–2014	41.5

[1]Discontinuous record and data gaps may exist within ranges of years.

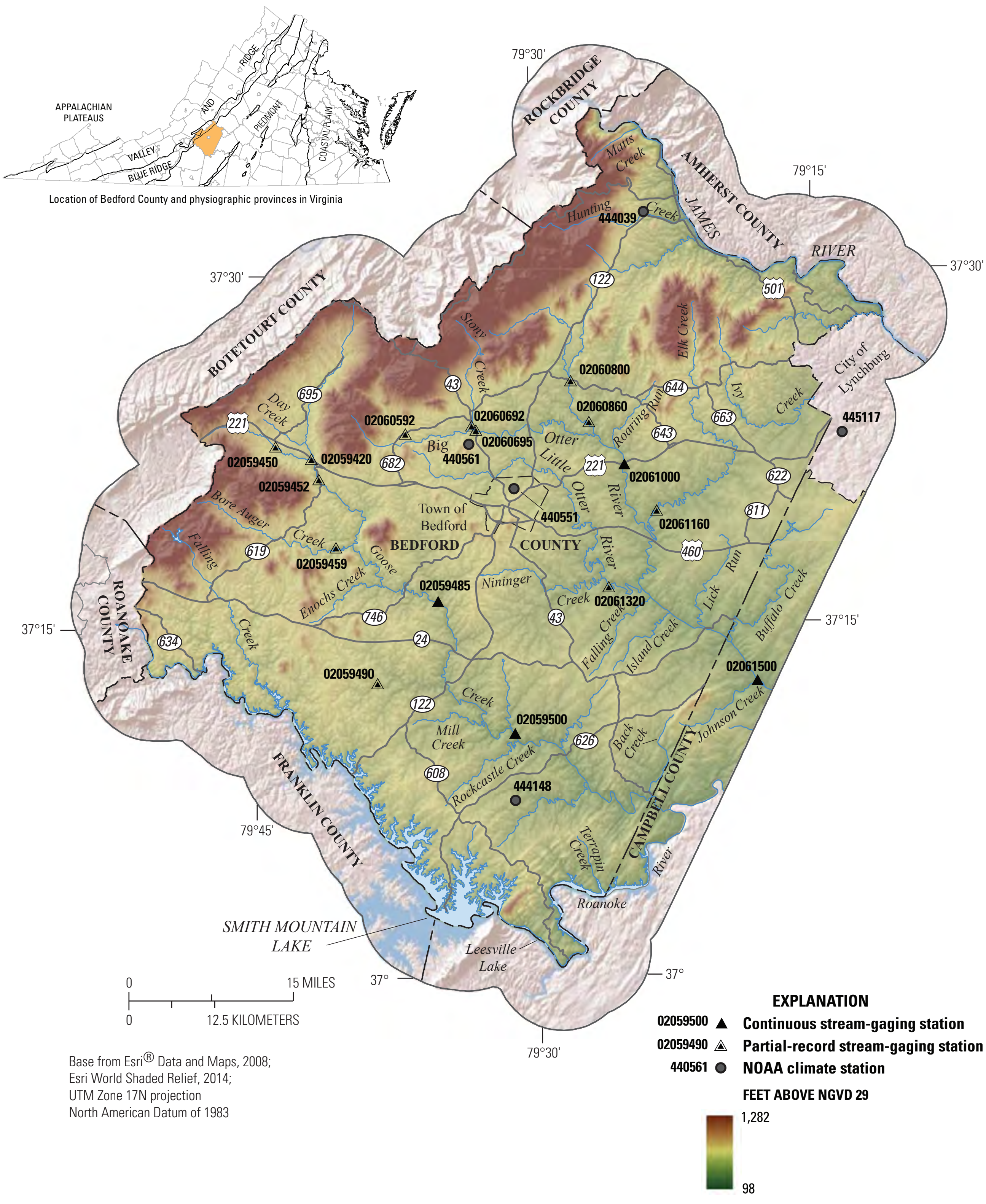


Figure 1. Locations of stream-gaging and climate stations within the study area, Bedford County, Virginia.

all five climate stations are similar; the largest differences generally occur between the months of April and July. In the center of the county, the Bedford 4 NW station indicates that monthly precipitation amounts peak mid-summer with 45 percent of annual precipitation falling during May through September; generally, precipitation amounts vary only slightly from month to month (fig. 2). Mean monthly precipitation is lowest from November through February.

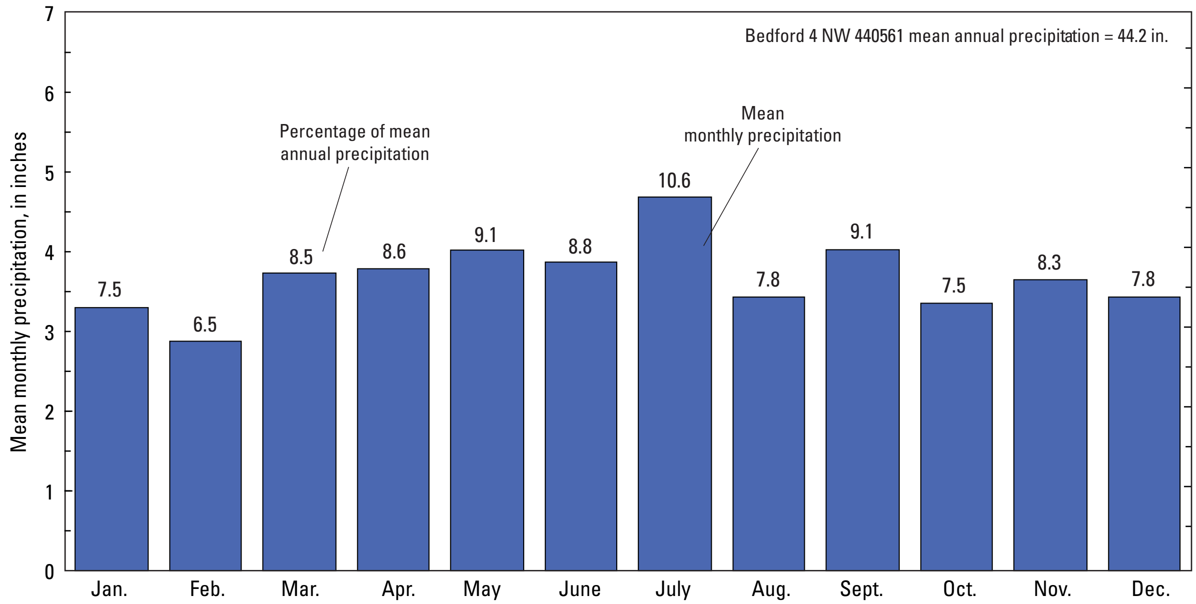


Figure 2. Mean monthly precipitation at National Oceanic and Atmospheric Administration climate station Bedford 4 NW 440561, 1981–2010.

Previous Investigations

During the mid- to late 1990s, the USGS Regional Aquifer System Analysis (RASA) Program used discharge data from surface-water basins to calculate base-flow yields and effective recharge for the Blue Ridge and Piedmont Physiographic Provinces (Rutledge and Mesko, 1996; Nelms and others, 1997; Swain and others, 2004). Rutledge and Mesko (1996) found a positive correlation between rates of recharge and base flow, or the water-yielding capacity of rocks in the Piedmont. Base-flow yields in the Piedmont, however, are lower in comparison to yields in the adjacent Valley and Ridge and Blue Ridge Physiographic Provinces (Nelms and others, 1997). By using data from the stream gage at Goose Creek near Huddleston (02059500, fig. 1) in Bedford County, mean effective recharge rates calculated by hydrograph separation techniques ranged from 10.45 in/yr during 1961–1990 (Rutledge and Mesko, 1996) to 8.40 in/yr during 1956–1984 (Nelms and others, 1997). For Big Otter River near Bedford (02061000, fig. 1), the mean effective recharge rate was 11.48 in/yr for the period 1945–1960 (Nelms and others, 1997).

Sanford and others (2012) developed a series of regression equations for the entire State of Virginia to quantify components of the water budget on a watershed and locality basis, including the town of Bedford and Bedford County. Groundwater recharge calculated by regression methods was 12.3 in/yr and 13.7 in/yr, respectively, for the town of Bedford and Bedford County. Their results for Bedford County indicate that 64.2 percent of average annual precipitation is lost to evapotranspiration processes or evaporation plus uptake by vegetation. Direct runoff to streams accounted for only 8.3 percent of total annual precipitation.

Heller (2008) summarized yields from public-supply wells in Virginia based on geologic province. Data indicated that public-supply wells from the Blue Ridge and Piedmont Physiographic Provinces were less likely to be high yielding (less than100 gallons per minute [gal/min]) than wells in other physiographic provinces in the State. Powell and Abe (1985) identified the major factors affecting well yields in the Piedmont of Virginia as topographic setting and the occurrence of clastic rock units; White (2012) later found topographic setting to also influence rates of recharge. In summarizing studies from the Piedmont in North Carolina and Maryland, neither Daniel and Payne (1990) nor Powell and Abe (1985) identified differences in the water-producing capacity of the various crystalline rock units in the Piedmont.

The occurrence of groundwater in crystalline rocks depends on the location and density of permeable fractures, some of which may be localized along regional structures (Keane and Gilstrap, 2011). Fracture-related heterogeneities in the Blue Ridge Physiographic Province are ubiquitous at a local scale (White, 2012) and can yield significant quantities of water if associated with faulting (Seaton and Burbey, 2005; White and Burbey, 2007). Fault zones have been observed to have permeabilities up to 6 orders of magnitude higher than the surrounding crystalline rock mass (Seaton and Burbey, 2005) and strongly control recharge processes in the Blue Ridge Physiographic Province (White and Burbey, 2007).

Lithology and structure are also important controls on groundwater yields in fractured rocks of the Piedmont Physiographic Province near Atlanta, Georgia (Chapman and others, 1998; Williams and others, 2005). Chapman and others (2011) studied a developed area in the Piedmont of North Carolina where annual water-level fluctuations exceeded 200 ft. Prior to development, water levels fluctuated only 15 to 50 ft. They concluded that the occurrence of intermittent dry wells during a period of drought correlated with geologic structure and a temporary loss of groundwater in storage potentially related to increased pumpage at community wells.

Much of the conceptual understanding of groundwater storage in the regolith, or weathered overburden, overlying fractured rocks of the Piedmont was developed by Heath (1984) and later expanded by Daniel and Harned (1998). White and Burbey (2007) and Seaton and Burbey (2005) characterize regolith of the Blue Ridge Physiographic Province as having significant geologic variability with confined conditions often found along a thin transition zone at the top of bedrock. The interaction of shallow water stored in the regolith with deeper fractures varies depending on the hydrologic connection from recharge to discharge zones (White and Burbey, 2007).

Hydrologic Data Collection

Various hydrologic data from Bedford County have been collected by the USGS and the DEQ since 1930. The longest period of record in the study area is at continuous stream-gaging station 02059500, Goose Creek near Huddleston, Va. Between 2006 and 2009, 12 partial-record streamflow gaging stations, 3 continuous groundwater-level observation wells, and 2 additional continuous stream-gaging stations were instrumented to record hydrologic data. Geologic data collected from 5 wells in Bedford County and more than 2,000 well records are presented to investigate the geohydrology in Bedford County.

Well Installation

Between May 13 and 15, 2008, monitoring wells were drilled at three county-owned locations for the purpose of monitoring water levels in an assumed recharge area of western Bedford County (site numbers 33G 1 SOW 224, 33G 2 SOW 225, and 35H 1 SOW 226; fig. 3; table 2). All three wells were drilled as 10-inch (in.)-diameter boreholes to the depth of competent bedrock where a 6.625-in. inside-diameter black steel casing was seated, and the annulus space exterior to the casing was backfilled with a bentonite slurry to land surface. Below the casing interval, wells were completed as open boreholes drilled to various field-determined depths at which a desired yield was obtained from bedrock aquifers. Well cuttings were collected from each drill hole. Detailed lithologic descriptions were made at the site by DEQ personnel using a hand lens and a Munsell color chart.

On October 16, 2011, well 35H 1 SOW 226 near Otterville, Va., was converted to a dual-zone groundwater monitoring well for independently monitoring hydraulic heads associated with transmissive fractures at 37 and 168 ft below land-surface datum. Interior 2-in. PVC monitoring wells were nested inside the open bedrock wellbore in a deep and shallow configuration with 10-ft sections of 0.2-in. slotted screen open to the transmissive fractures. Hydraulic heads were isolated by surrounding the screened intervals with #3 sand filter pack and placing bentonite pellets in lifts between the screened intervals of the deep and shallow 2-in. wells. Well 35H 1 SOW 226 was then discontinued, and the deep and shallow nested wells were assigned well numbers 35H 3 SOW 226A and 35H 4 SOW 226B, respectively.

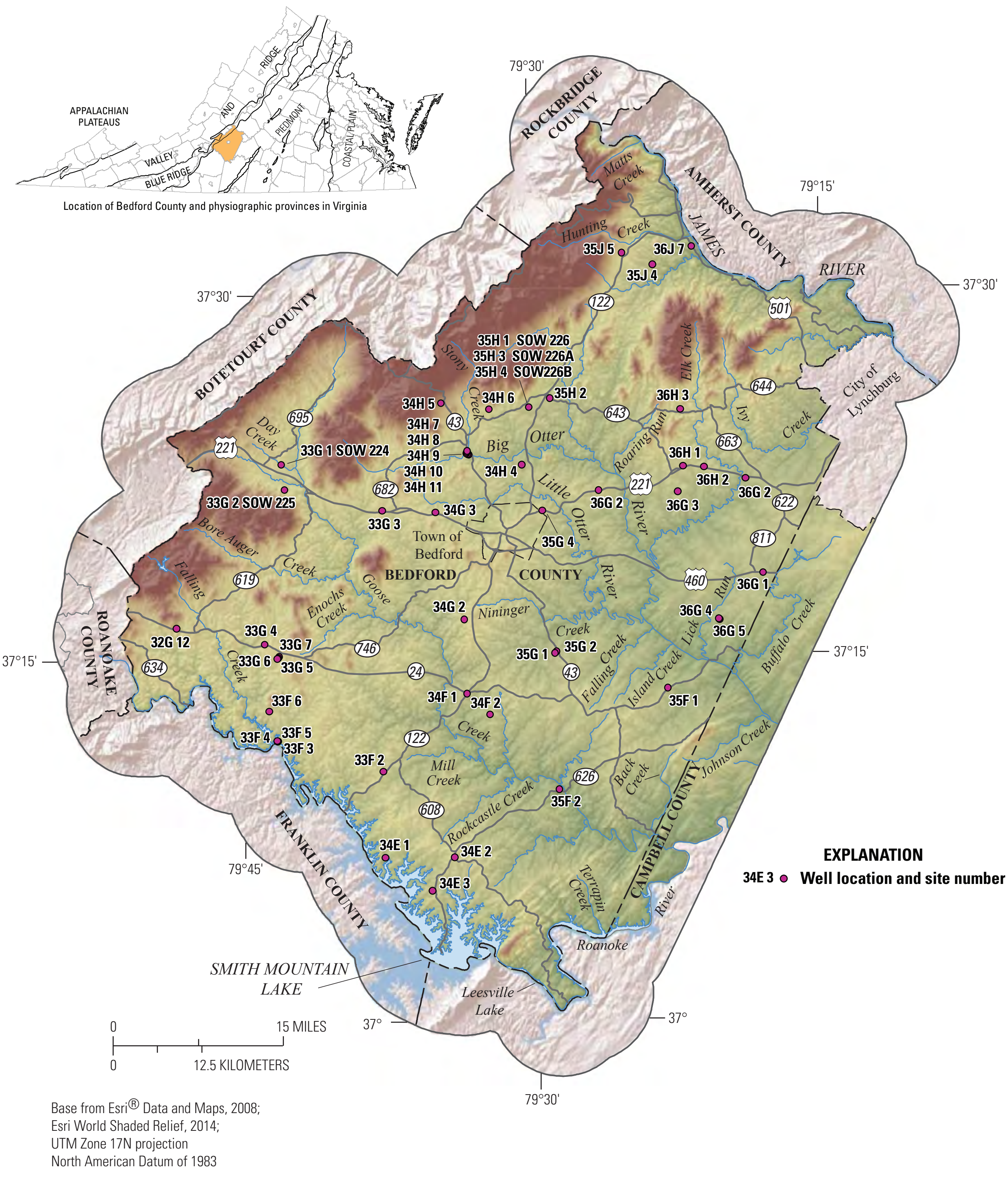


Figure 3. Locations of wells within the study area, Bedford County, Virginia.

Table 2. Site information for monitoring wells in Bedford County, Virginia.

[Site locations shown in figure 3. Abbreviations: USGS, U.S. Geological Survey; DDMMSS, degrees, minutes and seconds; Elev, elevation above National Geodetic Vertical Datum of 1929; ft, feet; ft-blsd, feet below land-surface datum; --, no data reported]

Site number	USGS station number	Latitude (DDMMSS)	Longitude (DDMMSS)	Elev (ft)	Well depth (ft-blsd)	Casing length (ft)	Screened interval (ft-blsd)	Water level data type	Hydrogeologic unit
[1]33G 1 SOW 224	372224079423601	372224	794236	930	101	51	--	hourly	Cambro-Ordovician metasedimentary rocks
[1]33G 2 SOW 225	372150079422301	372150	794223	1,070	201	95	--	hourly	Proterozoic basement crystalline rocks
[1, 2]35H 1 SOW 226	372543079295401	372543	792954	940	181	38	--	hourly	Proterozoic basement crystalline rocks
35H 3 SOW 226A	372543079295402	372543	792954	940	181	--	162–172	hourly	Proterozoic basement crystalline rocks
35H 4 SOW 226B	372543079295403	372543	792954	940	50	--	30–40	hourly	Proterozoic basement crystalline rocks
32G 12	371621079482801	371621	794828	1,053	--	--	--	biannual	Proterozoic basement crystalline rocks
33F 2	371019079375301	371019	793753	977	420	70	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
33F 3	371136079431901	371137	794319	823	--	--	--	biannual	Proterozoic basement crystalline rocks
33F 4	371138079431601	371138	794316	882	385	60	--	biannual	Proterozoic basement crystalline rocks
33F 5	371139079432001	371139	794320	844	445	31	--	biannual	Proterozoic basement crystalline rocks
33G 3	372106079374501	372106	793745	946	305	60	--	biannual	Proterozoic basement crystalline rocks
33G 4	371538079435501	371538	794355	997	246	32	--	biannual	Proterozoic basement crystalline rocks
33G 5	371508079431001	371508	794310	1,009	260	78	--	biannual	Proterozoic basement crystalline rocks
33G 6	371505079431301	371505	794313	985	280	42	--	biannual	Proterozoic basement crystalline rocks
[1]33G 7	371503079431701	371503	794317	982	300	55	--	biannual	Proterozoic basement crystalline rocks
[1]34E 1	370636079370601	370636	793706	940	350	85	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
34E 2	370645079341501	370645	793415	1,011	160	110	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
34E 3	370522079352601	370522	793526	932	--	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
33F 6	371252079434501	371252	794345	998	700	55	--	biannual	Proterozoic basement crystalline rocks
34F 1	371330079333001	371330	793330	911	254	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
34F 2	371237079321901	371237	793219	923	--	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
34G 2	371634079333501	371634	793335	1,002	--	--	--	biannual	Proterozoic basement crystalline rocks
34G 3	372100079345901	372100	793459	955	--	--	--	biannual	Proterozoic basement crystalline rocks
34H 4	372256079303001	372256	793030	1,055	--	--	--	biannual	Proterozoic basement crystalline rocks
34H 5	372531079343701	372531	793437	1,319	181	60	--	biannual	Proterozoic basement crystalline rocks

Table 2. Site information for monitoring wells in Bedford County, Virginia.—Continued

[Site locations shown in figure 3. Abbreviations: USGS, U.S. Geological Survey; DDMMSS, degrees, minutes and seconds; Elev, elevation above National Geodetic Vertical Datum of 1929; ft, feet; ft-blsd, feet below land-surface datum; --, no data reported]

Site number	USGS station number	Latitude (DDMMSS)	Longitude (DDMMSS)	Elev (ft)	Well depth (ft-blsd)	Casing length (ft)	Screened interval (ft-blsd)	Water level data type	Hydrogeologic unit
34H 6	372514079320801	372514	793208	1,006	450	21	--	biannual	Proterozoic basement crystalline rocks
34H 7	372323079331301	372323	793313	853	450	--	--	biannual	Proterozoic basement crystalline rocks
34H 8	372324079331701	372324	793317	854	413	--	--	biannual	Proterozoic basement crystalline rocks
34H 9	372326079332001	372326	793320	857	500	--	--	biannual	Proterozoic basement crystalline rocks
34H 10	372329079331901	372329	793319	861	450	--	--	biannual	Proterozoic basement crystalline rocks
34H 11	372332079331801	372332	793318	865	500	--	--	biannual	Proterozoic basement crystalline rocks
35F 1	371337079230701	371337	792307	858	305	65	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
35F 2	370929079284801	370929	792848	705	320	100	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
35G 1	371511079284901	371511	792849	874	--	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
35G 2	371508079285501	371508	792855	881	240	40	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
35G 3	372150079263101	372150	792631	910	--	--	--	biannual	Proterozoic basement crystalline rocks
35G 4	372102079292801	372102	792928	926	--	--	--	biannual	Proterozoic basement crystalline rocks
35H 2	372539079285901	372539	792859	922	206	90	--	biannual	Proterozoic basement crystalline rocks
35J 4	373108079233301	373108	792333	889	305	105	--	biannual	Proterozoic basement crystalline rocks
35J 5	373138079250801	373138	792508	1,056	--	--	--	biannual	Proterozoic basement crystalline rocks
36G 1	371819079180501	371819	791805	871	296	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
36G 2	372214079185501	372214	791855	811	175	67	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
36G 3	372143079222501	372143	792225	852	245	52	--	biannual	Proterozoic basement crystalline rocks
36G 4	371625079202301	371625	792023	816	--	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
36G 5	371627079202501	371627	792025	845	--	--	--	biannual	Late Proterozoic-Cambrian cover crystalline rocks
36H 1	372246079220901	372246	792209	892	--	--	--	biannual	Proterozoic basement crystalline rocks
36H 2	372244079210401	372244	792104	900	305	53	--	biannual	Proterozoic basement crystalline rocks
36H 3	372508079221401	372508	792214	1,013	--	--	--	biannual	Proterozoic basement crystalline rocks
36J 7	373151079213001	373151	792130	726	--	--	--	biannual	Proterozoic basement crystalline rocks

[1] Denotes well with borehole geophysical log.

[2] Well later completed as a multiple-zone monitoring well.

Groundwater Data

Pressure transducers with dataloggers were installed by the USGS in wells 33G 1 SOW 224 and 33G 2 SOW 225 and by the DEQ in well 35H 1 SOW 226, and subsequently installed by the DEQ in wells 35H 3 SOW 226A and 35H 4 SOW 226B to monitor and record water levels continuously. Water-level data were uploaded automatically via real-time satellite platforms to the USGS National Water Information System (NWIS, http://waterdata.usgs.gov/nwis) and field verified with manual tapedowns every 6 to 8 weeks.

Synoptic surveys of water levels in 47 wells (table 2; fig. 3) distributed across Bedford County were conducted during 1-week periods in December 2010 and April 2011. Water levels were measured manually with a steel tape to the nearest one-hundredth of a foot. An electric tape was substituted for the steel tape in wells where the depth to water could not be successfully determined with the steel tape, typically because of moisture inside the well casing. At each visit, two water-level measurements were made, at least 5 minutes apart, to detect any short-term water-level fluctuations attributed to recent pumpage. The intent of the survey was to evaluate the seasonal water-table fluctuations across the county that coincide with periods of water-level lows in early winter and water-level highs in early spring. The water-level data were used to construct a water-level contour map for the county. Available well information was inventoried from well owner or local health department records.

Borehole Logging

Borehole geophysical logging of the three wells constructed during this study, as well as two additional wells (table 2), included caliper; ambient and pumping fluid resistivity and temperature; ambient and pumping electromagnetic flow; multiparameter logs (gamma, formation resistivity); and acoustic televiewer (ATV) and borehole video. ATV, borehole video, and caliper logs were collected to characterize the locations and orientation of subsurface features in the fractured bedrock aquifers. ATV logging produces a high-resolution, magnetically oriented, digital image that was used to delineate and measure orientations of fractures at depth that intersect the borehole. When used in conjunction with other logs, data from the caliper logs are diagnostic in pinpointing locations of open and transmissive fractures. Ambient and pumping electromagnetic (EM) flow and fluid resistivity and temperature logs were run to ascertain the number and approximate position of transmissive fractures for four of the five wells in the study area. Gamma and formation resistivity logs were collected from all wells for evaluation of changes in rock type and formation competency adjacent to the wellbore.

Well Construction Report Compilation

Digital water-well records were compiled from the Virginia Department of Health Virginia Environmental Information System (VENIS), the Virginia Department of Health Safe Drinking Water Information System (SDWIS), the USGS Groundwater Site Inventory (GWSI), the DEQ Virginia Water Use Data System (VWUDS), and the U.S. Environmental Protection Agency (EPA) Storage and Retrieval (STORET) database. The location accuracy and completeness of the water-well data vary within and among databases. Coordinate information for SDWIS public water supply wells is typically of Global Positioning System (GPS) or Differential Global Positioning System (DGPS) quality, and location data for most of the wells originating in the GWSI and STORET databases were determined from locations using 7.5-minute quadrangles. Virginia Department of Health well data were located by assigning the coordinates for the tax parcel centroid associated with the well because original coordinate data were not given. A total of 2,140 well records were compiled and stored in a geographic information system (GIS) shapefile for geospatial analysis.

Streamflow Measurements

Streamflow was measured with continuous data recorders every 15 minutes at three gaging locations in Bedford County and one gaging location in Campbell County within the Roanoke River Basin (fig. 1; table 3). Two pre-existing stream gages—one on Goose Creek near Huddleston, Va. (02059500) and one on Big Otter River near Evington, Va. (02061500)—had been in operation since 1930 and 1936, respectively. In 2006, two additional stream gages were established as part of the current investigation—one on Goose Creek at Route 747 near Bunker Hill, Va. (02059485) and the other on Big Otter River at Route 221 near Bedford, Va. (02061000). Partial-record streamflow measurements (sites where discrete streamflow measurements were obtained over a period of time without continuous data being recorded) were also conducted at 12 sites within the county (fig. 1; table 3).

Table 3. Streamflow measurement gages in Bedford County, Virginia.

[Site locations shown in figure 1. Abbreviations: USGS, U.S. Geological Survey; DDMMSS, degrees, minutes, and seconds; mi^2, square miles]

Station number	USGS station name	Latitude (DDMMSS)	Longitude (DDMMSS)	Drainage area (mi^2)	Station type	Period of record
02059485	Goose Creek at Route 747 near Bunker Hill, VA	371559	793516	125	Continuous	2007–2012
02061000	Big Otter River at Route 221 near Bedford, VA	372150	792510	114	Continuous	2007–2012
02059500	Goose Creek near Huddleston, VA	371023	793114	188	Continuous	1931–2012
02061500	Big Otter River near Evington, VA	371230	791814	315	Continuous	1938–2012
02059420	North Fork Goose Creek at Route 460 near Montvale, VA	372214	794155	31.5	Partial Record	2006–2009
02059450	South Fork Goose Creek at Route 607 at Montvale, VA	372247	794350	11.0	Partial Record	2006–2009
02059452	Goose Creek at Route 726 near Irving, VA	372122	794132	46.8	Partial Record	2006–2009
02059459	Bore Auger Creek at Route 754 near Irving, VA	371826	794040	17.6	Partial Record	2006–2009
02059490	Stony Fork at Route 608 near Moneta, VA	371236	793833	7.01	Partial Record	2006–2009
02060592	Sheep Creek at Route 688 near Thaxton, VA	372316	793653	10.3	Partial Record	2006–2009
02060692	Stony Creek at Route 43 near Peaks of Otter, VA	372334	793319	14	Partial Record	2006–2009
02060695	Big Otter River at Route 43 near Peaks of Otter, VA	372323	793305	37	Partial Record	2006–2009
02060800	North Otter Creek at Route 643 near Cifax, VA	372526	792758	25.8	Partial Record	2006–2009
02060860	Oslin Creek at Route 644 near Cifax, VA	372340	792702	12	Partial Record	2006–2009
02061160	Elk Creek at Route 668 near Goode, VA	371950	792329	41	Partial Record	2006–2009
02061320	Little Otter River at Route 715 near Otter Hill, VA	371636	792606	66.4	Partial Record	2006–2009

Hydrogeology of Piedmont and Blue Ridge Aquifers

Hydrogeologic conditions and processes that control groundwater availability in the Piedmont and Blue Ridge fractured-rock aquifer systems in Bedford County were investigated by a synthesis of the literature and of data sources and analyses that are described in the following sections. The rates of groundwater inflow, outflow, and volume of water stored in Piedmont and Blue Ridge aquifers are defined by the prevailing climate conditions and topography and by the complex distributions of aquifer properties in the regolith and fractured-rock parts of the aquifer.

Hydrogeologic Units

Piedmont and Blue Ridge fractured-rock aquifers within Bedford County were subdivided into three distinct hydrogeologic units following White (2012): (1) Cambro-Ordovician metasedimentary rocks, (2) Proterozoic basement crystalline rocks, and (3) Late Proterozoic-Cambrian cover crystalline rocks (figs. 4 and 5). For the Proterozoic basement crystalline rock and Late Proterozoic-Cambrian cover crystalline rock hydrogeologic units, subunits are defined by grouping geologic map units with similar metamorphic history or textural characteristics. While local hydrogeologic conditions are highly variable within and across all three hydrogeologic units, these groups warrant distinction because of differences in texture and structure that can influence the storage and movement of groundwater (Daniel and Payne, 1990). The three hydrogeologic units occur within the Blue Ridge Anticlinorium, a larger geologic feature in Virginia that is composed of a stack of northeast- to southwest-striking thrust sheets (Bailey and others, 2006; Southworth and others, 2009).

Orientations of fractures and foliation (the preferential orientation of minerals and fabric in metamorphic rocks) were collected from borehole geophysical logs and plotted to describe general fracture trends of the three hydrogeologic units in the study area. Trends were assessed from borehole ATV geophysical log data collected in three wells in the Proterozoic basement crystalline rocks, one well in the Late Proterozoic-Cambrian cover crystalline rocks, and one well in the Cambro-Ordovician metasedimentary rocks (fig. 6). Foliation measurements were collected only in portions of wells that exhibited a metamorphic fabric or texture (schistosity and [or] cleavage) because of the inability of the ATV to resolve lineations present on otherwise smooth borehole wall. Fluid resistivity, fluid temperature, formation resistivity, natural gamma, EM flow, and ATV logs from the five wells are shown in Appendix 1. Several general statements can be made about the distribution and orientation of structural features collected from ATV logs in the study area:

1. Fracturing in the Proterozoic basement crystalline rocks appears to be systematic at the borehole level but not at a regional scale. Similar observations have been made by workers studying fracturing at the outcrop and the regional level elsewhere in the Blue Ridge (Bailey and others, 2003; Hasty and Bailey, 2005).

2. A pervasive southeast dipping foliation can occur in bedrock wells across all hydrogeologic units. This foliation is more prevalent in the Late Proterozoic-Cambrian cover crystalline rocks and Cambro-Ordovician metasedimentary rocks, and fracturing related to foliation within these units may be a discernible regional trend.

3. Water-bearing fractures occur at a variety of orientations. Notable trends in the orientation of water-bearing fractures across all hydrogeologic units are as follows:

   a. Fractures parallel or subparallel to foliation (parting along foliation or schistosity).

   b. Horizontal or subhorizontal water-bearing joints (sheet jointing).

   c. Water-bearing fractures occurring as a component of a conjugate fracture set (tectonically induced fractures).

Similar trends have been noted in other locations along the Atlantic Seaboard by workers conducting hydrogeologic studies in crystalline rock (Chapman and others, 2005; Williams and others, 2005; Manda and others, 2008).

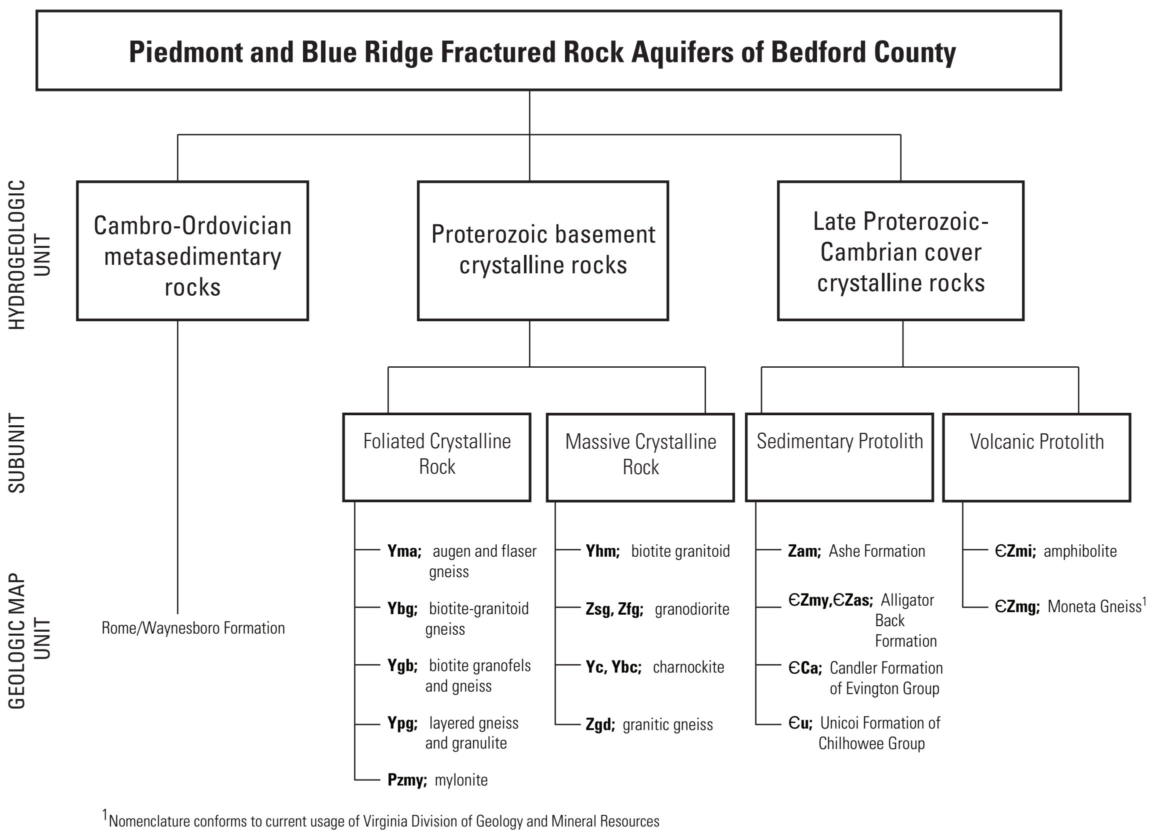


Figure 4. Major hydrogeologic units, subunits, and geologic map units in Bedford County, Virginia.

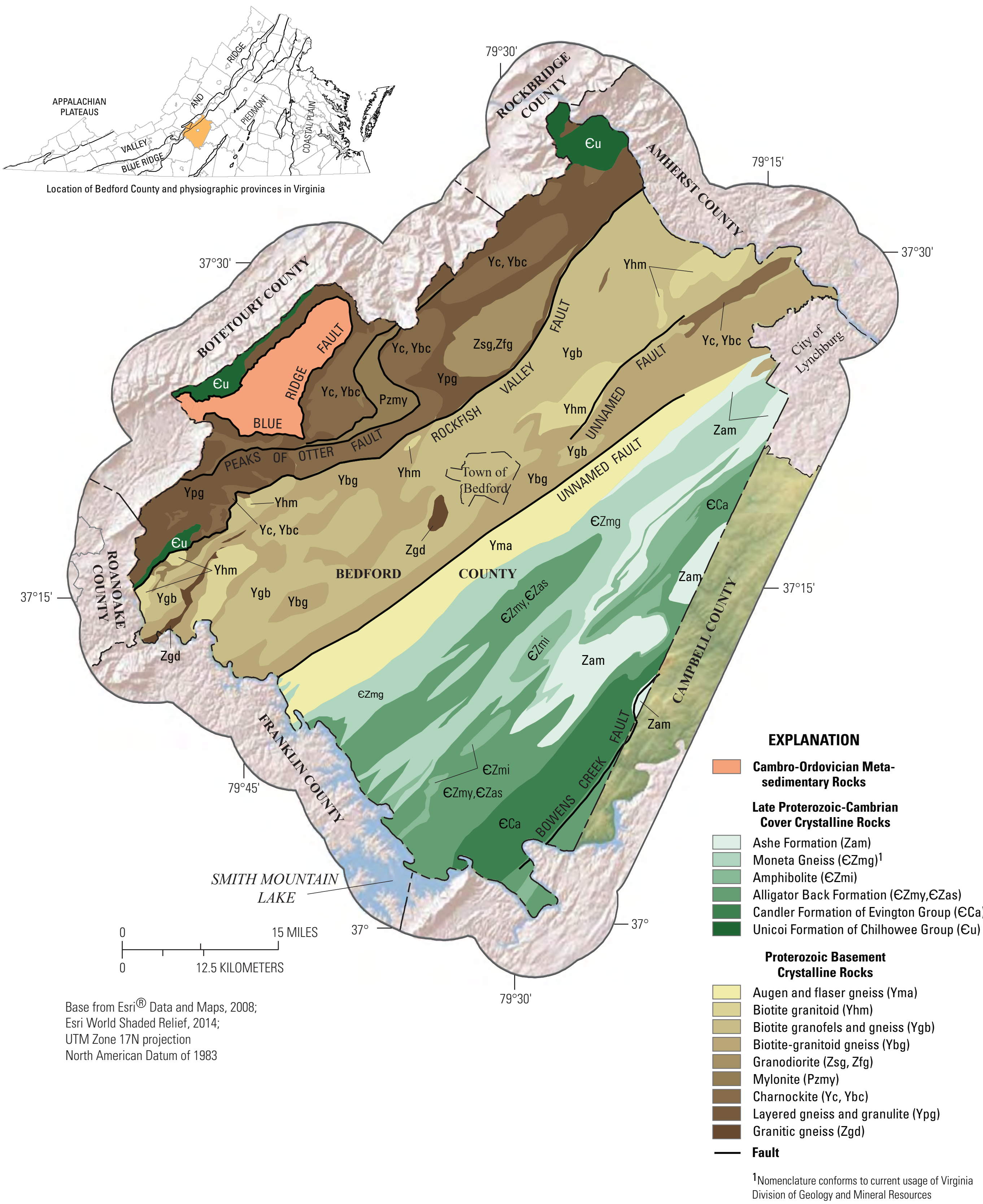


Figure 5. Major hydrogeologic units and geologic map units in Bedford County, Virginia. [Modified from Henika (1997).]

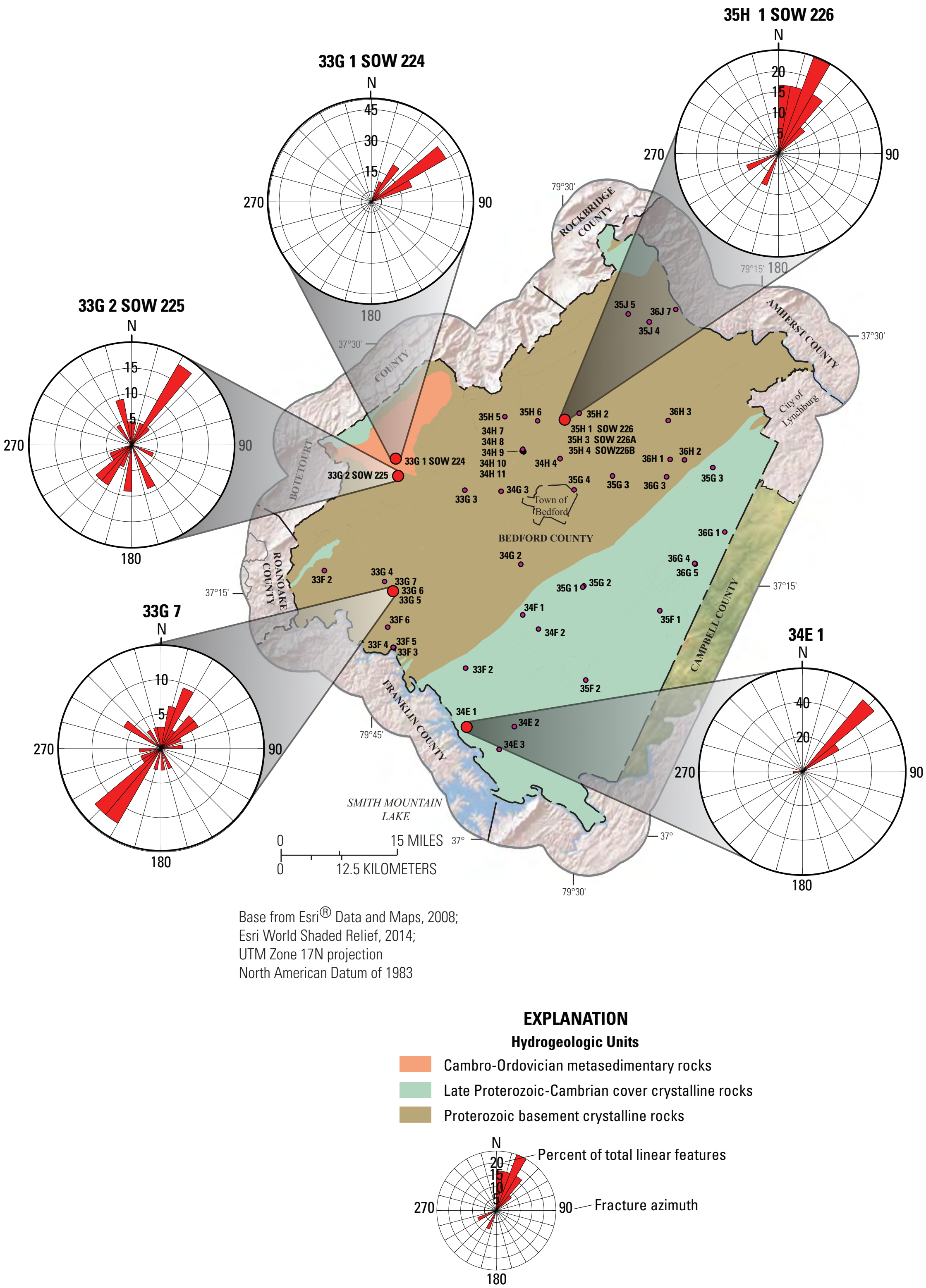


Figure 6. Rose diagrams showing all linear features from selected wells in Bedford County, Virginia. [Fracture data obtained from acoustic televiewer logs. Geology modified from Henika (1997).]

Cambro-Ordovician Metasedimentary Rocks

The Cambro-Ordovician-aged metasedimentary rocks occur in the valleys along the north and south forks of Goose Creek in the northwestern portion of the county (fig. 5) and are exposed at land surface where erosion has removed the overlying Proterozoic basement crystalline rocks (Henika, 1981). In Bedford County, Cambro-Ordovician metasedimentary rocks are composed predominantly of the Rome-Waynesboro Formation—an interbedded sequence of mudstone, argillite, quartzite, and dolomite. Carbonate rocks of the Elbrook Limestone, Conococheague Limestone, and the Stonehenge Limestone of Beekmantown Group occur in the extreme northwestern end of Bedford County and are also included in this unit. These rocks originated as cyclic shallow marine sequences deposited in a cratonic basin during the late Cambrian and early Ordovician (Butts, 1940). Subsequent to lithification, these rocks were folded and faulted during the mid- to late Paleozoic. In Bedford County, rocks of the Rome-Waynesboro Formation are thrust over the younger carbonate units (Henika, 1981).

Well 33G 1 SOW 224 was drilled in the Rome-Waynesboro Formation. All noted fractures in the well strike between 15° and 70°, and although no prevalent schistosity was noted in the well with the televiewer, these fractures strike in the same general direction as the foliation noted in wells 35H 1 SOW 226 and 34E 1, which are located in the Proterozoic basement crystalline rocks and Late Proterozoic-Cambrian cover crystalline rocks, respectively (fig. 6). Dip angles for noted fractures in well 33G 1 SOW 224 varied between 29° and 74°, with most dip angles greater than 50° (Appendix 1). Orientations for water-bearing fractures noted in well 33G 1 SOW 224 identified from EM flow and camera logging are coincident with the strikes and dip angles of other non-transmissive fractures in the well. Visual logs of the well via borehole camera show zones of complex folding and faulting with a persistent cleavage striking to the northeast. All open fractures observed on the camera log for well 33G 1 SOW 224 occurred within carbonate intervals.

The Rome-Waynesboro Formation was the only geologic unit in the Cambro-Ordovician metasedimentary rocks hydrogeologic unit with sufficient well completion data for summary evaluation. Wells drilled in the Rome-Waynesboro Formation are typically less than 300 ft deep (median well depth of 225 ft), with median yield of 20 gal/min, median depth to bedrock of 71.5 ft, and median casing depth of 108 ft (figure 7; table 4). Median depths for first and lowest reported water-bearing zones occur at 145 ft (15 wells) and 150 ft (7 wells), respectively, approximately 70 to 80 ft below the median depth to the bedrock interface.

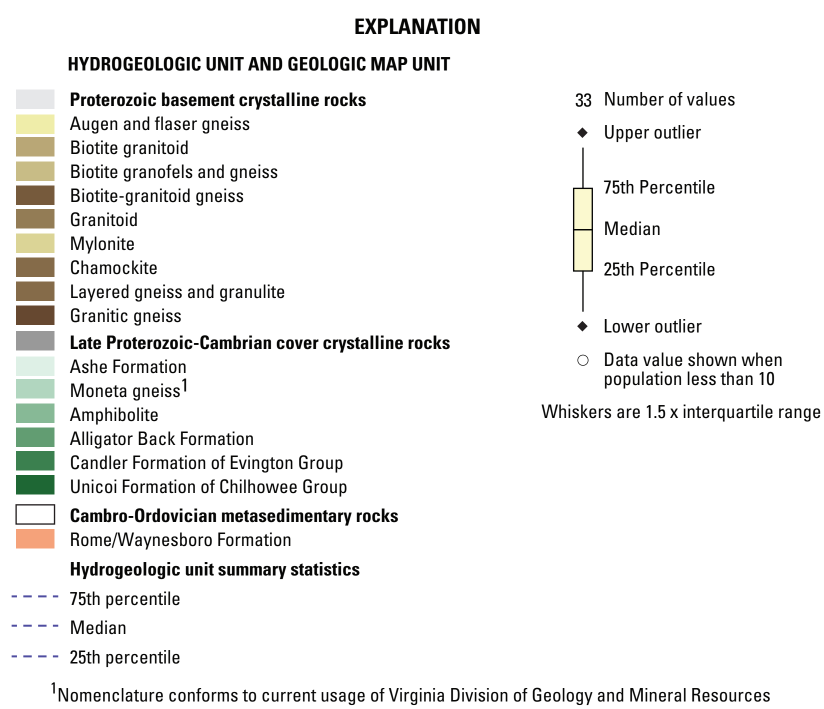


Figure 7. Well characteristics by hydrogeologic unit in Bedford County, Virginia.

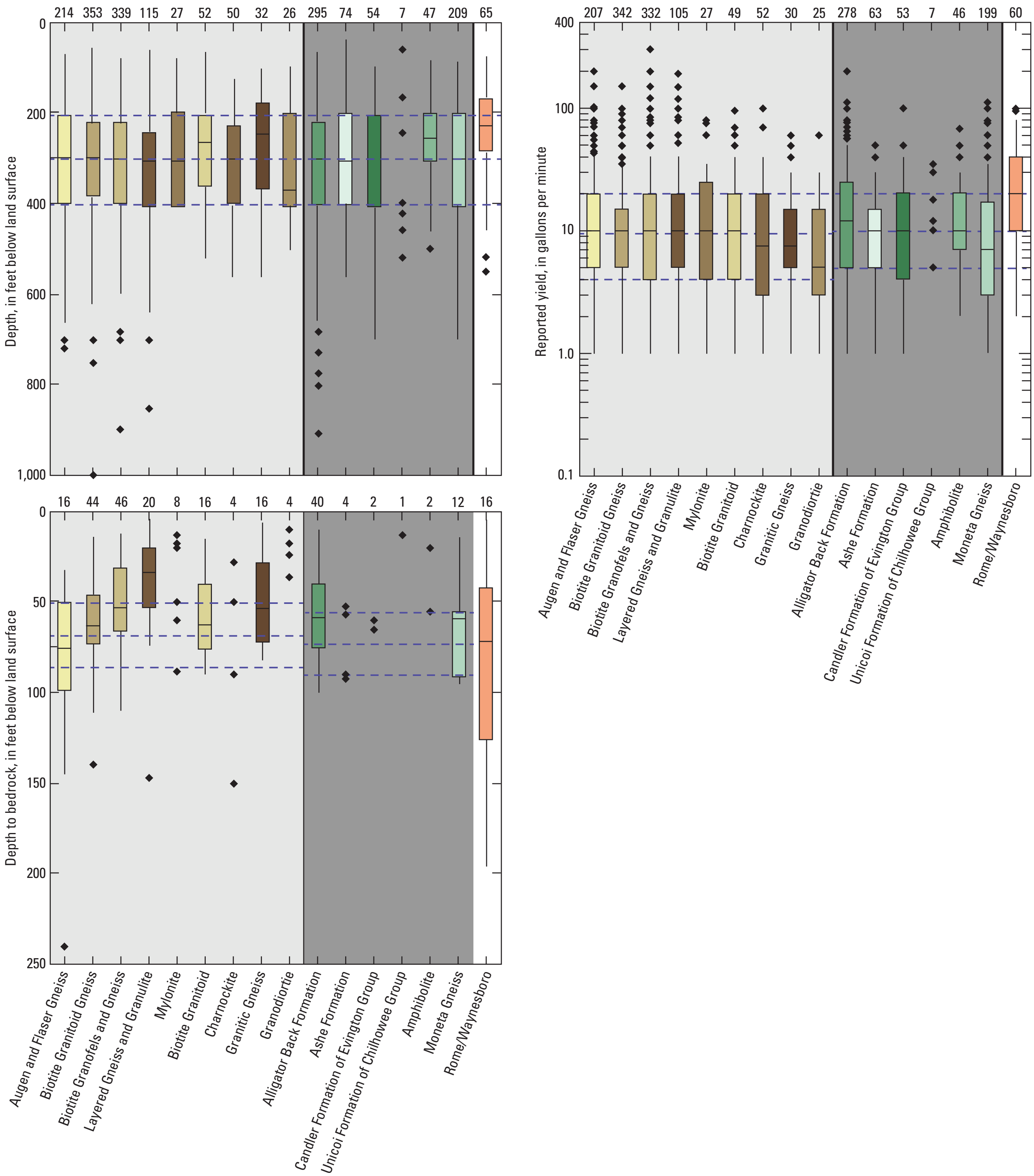


Figure 7. Well characteristics by hydrogeologic unit in Bedford County, Virginia.—Continued

Table 4. Descriptive statistics of hydrogeologic parameters by rock type associated with a selection of drilled water wells in Bedford County, Virginia.

Hydrogeologic unit	Proterozoic basement crystalline rocks											
Sub unit	Foliated crystalline rock						Massive crystalline rock					
Geologic map unit	Augen and flaser gneiss	Biotite-granitoid gneiss	Biotite granofels and gneiss	Layered gneiss and granulite	Mylonite	All foliated	Biotite granitoid	Charnockite	Granitic gneiss	Granodiorite	All massive	All
	Reported yield at drilling, in gallons per minute											
Median	10	10	10	10	10	10	8	7.5	7.5	5	7	10
Minimum	1	1	1	1	1	1	1	1	1	1	1	1
Maximum	200	150	300	191	80	300	95	100	60	60	100	300
Number of observations	207	342	332	105	27	1,013	49	52	30	25	156	1,169
	Well depth, in feet											
Median	300	300	300	305	305	300	265	300	245	370	300	300
Minimum	65	54	80	60	75	54	65	125	100	95	65	54
Maximum	720	1,000	900	850	560	1,000	520	560	560	500	560	1,000
Number of observations	214	353	339	115	27	1,048	52	50	32	26	160	1,208
	Depth to bedrock, in feet											
Median	75.5	63	53	33.5	20	54	62.5	70	53.5	21	52.5	53.5
Minimum	32	19	12	2	13	2	15	28	6	10	6	2
Maximum	240	140	110	147	88	240	90	150	82	36	150	240
Number of observations	16	44	46	20	8	134	16	4	16	4	40	174
	Casing depth, in feet											
Median	75	67.5	63	70	65	68	60.5	60	65.5	40	60	66
Minimum	20	9	13	16	20	9	30	23	20	19	19	9
Maximum	160	180	140	148	120	180	103	124	100	80	124	180
Number of observations	199	292	302	102	20	915	40	42	26	23	131	1,046
	Depth to first water-bearing zone, in feet											
Median	115	102	105	101	88	105	122	115	123	96.6	117	110
Minimum	42	40	31	29	35	29	50	70	50	38	38	29
Maximum	210	355	340	315	230	355	260	250	215	180	260	355
Number of observations	15	56	51	18	9	149	18	6	19	5	48	197
	Depth to lowest water-bearing zone, in feet											
Median	180	205	185	196	110	192	209	170	210	199	202	195
Minimum	95	110	50	90	60	50	183	150	115	168	115	50
Maximum	285	478	410	390	180	478	434	190	295	231	434	478
Number of observations	8	33	37	10	4	92	6	2	10	2	20	112

Table 4. Descriptive statistics of hydrogeologic parameters by rock type associated with a selection of drilled water wells in Bedford County, VA.—Continued

Hydrogeologic unit	Late Proterozoic-Cambrian cover crystalline rocks									Cambro-Ordovician metasedimentary rocks	
Sub unit	Sedimentary Protolith					Volcanic Protolith					
Hydrogeologic parameter	Alligator Back Formation	Ashe Formation	Candler Formation of Evington Group	Unicoi Formation of Chilhowee Group	All sedimentary	Amphibolite	Moneta gneiss[1]	All volcanic	All	Rome-Waynesboro Formation	Elbrook Limestone
	Reported yield at drilling, in gallons per minute										
Median	12	10	10	12	10	12	7	8	10	20	
Minimum	1	1	1	5	1	2	1	1	1	2	
Maximum	200	50	100	35	200	68	112	112	200	100	
Number of observations	278	63	53	7	401	46	199	245	646	60	
	Well depth, in feet										
Median	300	305	305	400	300	255	300	300	300	225	300
Minimum	65	6	96	60	37	84	85	84	37	74	300
Maximum	908	560	700	520	908	500	700	700	908	550	300
Number of observations	295	74	54	7	429	47	209	256	685	65	1
	Depth to bedrock, in feet										
Median	54.1	72.7	62.5	13	57	37.5	66.2	57	58	71.5	
Minimum	10	52	60	13	10	20	14	14	10	4	
Maximum	100	92	65	13	100	55	95	95	100	196	
Number of observations	40	4	2	1	49	2	12	14	61	16	
	Casing depth, in feet										
Median	71.6	64	79	87.2	70	68.1	71.1	68	70	108	82
Minimum	16	20	30	51	16	20	10	10	10	20	82
Maximum	142	122	140	123	142	140	130	140	142	441	82
Number of observations	158	66	52	7	383	43	198	241	624	56	1
	Depth to first water-bearing zone, in feet										
Median	160	136	105	140	145	175	138	132	142	145	
Minimum	25	45	60	140	25	140	75	75	25	62	
Maximum	380	250	150	140	380	210	280	280	380	235	
Number of observations	47	4	2	1	54	2	12	14	68	15	
	Depth to lowest water-bearing zone, in feet										
Median	220		200	150	225	240	181	185	202	150	
Minimum	53		150	150	53	230	92	92	53	100	
Maximum	440		250	150	440	250	320	320	440	260	
Number of observations	18		2	1	21	2	8	10	31	7	

[1]Nomenclature conforms to current usage of Virginia Division of Geology and Mineral Resources.

Proterozoic Basement Crystalline Rocks

Proterozoic-aged basement crystalline rocks occur to the northwest of a northeast-trending contact with the Late Proterozoic-Cambrian cover crystalline rocks (fig. 5). Proterozoic basement crystalline rocks within Bedford County are composed predominantly of biotite-rich gneisses surrounding pods of relatively undeformed granitic rocks to the south of the Rockfish Valley Fault, and more pyroxene-rich granulites, gneisses, and charnockitic rocks to the north of the Rockfish Valley Fault (Henika, 1997). These are the oldest rocks in the study area, and many bear evidence of multiple periods of deformation. These rocks are variably foliated, granitoid gneisses and granitoids defined by a characteristic crystalline structure as well as by their lowest stratigraphic position within the Blue Ridge Anticlinorium. Basement crystalline rocks include the augen and flaser gneiss (Yma), biotite-granitoid gneiss (Ybg), biotite granofels and gneiss (Ygb), biotite granitoid (Yhm), granitic gneiss (Zgd), charnockite (Yc,Ybc), layered gneiss and granulite (Ypg), mylonite (Pzmy), and granodiorite (Zsg, Zfg) geologic map units described in the geologic map of the Roanoke 30×60-minute quadrangle (Henika, 1997).

Fracturing within basement crystalline rocks is highly variable and dependent on the orientation of local stress fields in the rock imparted during deformation and exhumation, and in some cases by the presence or absence of metamorphic fabric (foliation and schistosity). Observations by the authors of basement outcrop within the study area and analysis of limited borehole geophysical log data obtained from three wells within the crystalline basement portion of the county indicate that jointing and parting along metamorphic fabric appear to be the predominant styles of fracturing within these rocks.

Conjugate sets of fractures striking between 20° to 60° and 210° to 230° were noted in well 33G 7 (fig. 6). Dip angles for noted fractures in well 33G 7 varied between 17° and 84° with the majority of noted fractures dipping at angles greater than 50° (Appendix 1). Four of the six water-bearing fractures identified in well 33G 7 through EM flow and fluid logging have dip angles of 50° or less. The main water-producing zones for well 33G 7 are associated with a steeply dipping fracture (dip angle of 72°) at 119 ft below land surface (bls) and a moderately dipping fracture (dip angle of 17°) at 135 ft bls.

Conjugate fracture sets in well 33G 2 SOW 225 strike between 325° to 50° and 115° to 245°. Four water-bearing fractures were noted in well 33G 2 SOW 225. Two of the noted water-bearing fractures strike within the easterly dipping fracture set (strike angles of 26° and 343°), and two water-bearing fractures strike within the westward dipping fracture set (strike angles of 216° and 222°). Dip angles for noted fractures within well 33G2 SOW 225 vary between 11° and 74° with predominant dip angles for noted fractures ranging between 50° and 70°. Three of the noted water-bearing fractures in well 33G 2 SOW 225 have dip angles between 18° and 28° and collectively yielded about 3 gal/min during air lifting as the well was drilled. The main water-bearing fracture at 188 ft bls yielded approximately 10 gal/min and has a dip angle of 68°, which is coincident with the predominant dip of fracturing within the wellbore.

Well 35H 1 SOW 226 was drilled in a mylonitic section of granodiorite and has a pervasive schistosity striking between 0° and 50° (Appendix 1). Two water-bearing zones were identified in this well—a nearly horizontal joint at 38 ft bls dipping at 2° and striking at 17° that yielded less than 1 gal/min during air lifting and a joint at 169 ft bls striking at 201° and dipping northwest at nearly 20°. Yield from the lower fracture during air lifting was approximately 60 gal/min. EM flow and fluid profiling confirmed observations about the positions and yields of transmissive fractures recorded during drilling of the well.

Wells drilled in the Proterozoic basement crystalline rocks typically have a median depth ranging from 300 to 305 ft (table 4; fig. 7) with median casing depth of 60 to 75 ft. Drilling depths of less than 300 ft are common for wells completed in the biotite granitoid and granitic gneiss map units, and drilling depths of greater than 350 ft and shallower casing depths (median value of 40 ft) can be expected for wells drilled in the granodiorite map unit. Reported median yield values for all rocks within the basement crystalline rocks ranged between 5 and 10 gal/min. The highest reported well yield of 300 gal/min occurred in the biotite granofels and gneiss map unit. Reported water-bearing zones occur at median depths between 110 ft and 195 ft, and median depth to bedrock occurs at 53.5 ft.

Late Proterozoic-Cambrian Cover Crystalline Rocks

Late Proterozoic-Cambrian cover crystalline rocks occur primarily to the southeast of the northeast trending contact with the basement crystalline rocks and in smaller areas of the county along the northwestern borders with Botetourt, Rockbridge, Roanoke, and Amherst Counties (fig. 5). These rocks were unconformably extruded or deposited on basement crystalline rocks and include rocks of volcanic and sedimentary origin. The rocks in this hydrogeologic unit include the Alligator Back Formation (ЄZmy, ЄZas), amphibolite (ЄZmi), Moneta Gneiss (ЄZmg—nomenclature conforms to current usage of the Virginia Division of Geology and Mineral Resources), Ashe Formation (Zam), Candler Formation of the Evington Group (ЄCa), and Unicoi Formation of the Chilhowee Group (Єu) described in Henika (1997).

Preserved textures in these rocks generally progress from coarse-grained conglomeratic metagraywacke and gneiss in the Ashe Formation to finer grained schistose and phyllitic rocks in the Alligator Back Formation and Candler Formation of the Evington Group. Along the northwestern border of the county, rocks of the Unicoi Formation of the Chilhowee group are composed of heterogeneous packages of coarse-grained basal units that typically fine upward into phyllitic and quartzose sequences (Henika, 1981). Well 34E 1 was completed in the Alligator Back Formation. Foliation was resolved with the optical televiewer throughout the well and strikes uniformly between 40° and 60° with dip angles between 50° and 79° (fig. 6; Appendix 1).

Fracturing along and subangular to foliation is prevalent. A single transmissive joint at 292 ft bls strikes at 71° and dips 8°. Cascading conditions were observed in the well while camera logging. Cascading conditions indicate a downward hydraulic gradient between a flowing fracture observed on the camera log at 103 ft bls and the transmissive joint at 292 ft bls.

Wells drilled in the Late Proterozoic-Cambrian cover crystalline rocks have a median depth of 300 ft with a median casing depth of 70 ft (table 4; fig. 7). The median depth to bedrock in the Late Proterozoic-Cambrian cover crystalline rocks was 58 ft. Median well yields among geologic map units in the Late Proterozoic-Cambrian cover crystalline rocks range between 7 gal/min (Moneta Gneiss) and 12 gal/min (Alligator Back Formation, Unicoi Formation of the Chilhowee Group, and amphibolite). Median depths to first and lowest reported water-bearing zones are 142 and 202 ft, respectively.

Low-yielding or seasonally dry wells have been reported in southeastern areas of the county (Todd Fowler, Bedford County Health Department, written commun., 2012) in areas underlain by Late Proterozoic-Cambrian cover crystalline rocks. A simple one-way analysis of variance (ANOVA) test on drillers' well yields was conducted to assess the hypothesis that low-yielding wells were more common in Late Proterozoic-Cambrian cover crystalline rocks than in the Proterozoic basement crystalline rocks. Yields were binned by hydrogeologic unit, and the mean yields for Proterozoic basement crystalline rocks and Late Proterozoic-Cambrian cover crystalline rocks were found to be identical (fig. 7). The ANOVA test resulted in a p-value of 0.26 that indicates no differences in well yields in the basement and cover crystalline rocks across the scale of the entire county. Daniel and Harned (1998) similarly found no difference in yields among crystalline rock units in the Piedmont of North Carolina. Reported low-yielding wells in the southeastern part of the county likely result from local heterogeneities of which insufficient data exist to characterize and which are not reflected in the current conceptual model.

Conceptual Model of Groundwater Flow

In the Blue Ridge Physiographic Province, topography provides the driving force for groundwater flow, and the water table is generally a subdued reflection of the land surface (figs. 8, 9). The high density of streams in the Blue Ridge Physiographic Province is probably the result of the underlying low-permeability rocks. The porous nature of the overlying regolith material (overburden) and the sharp contrast in permeability with the underlying hydrogeologic units coupled with numerous streams and hilly terrain are probably conducive for the development of local flow systems. Groundwater discharges at low elevation points along small streams and headwater springs (White, 2012). Groundwater flow in the bedrock is primarily controlled by the irregular fracture network and steep terrain (Wright, 1990) and by the development of complex potentiometric head relations between recharge and discharge zones (White, 2012). Local flow systems may be controlled by stress relief or other brittle fracturing; however, bedding-plane partings in metasedimentary rocks and foliations in the metamorphic rocks may facilitate more of a subregional- to regional-type flow.

The groundwater-flow system of the Piedmont Physiographic Province can be idealized to consist of four components (fig. 9): (1) unsaturated regolith, (2) saturated regolith, (3) a transition zone between regolith and bedrock aquifer, and (4) fractured bedrock aquifer or hydrogeologic unit as described previously (Daniel and Harned, 1998). The regolith is a mantle of organic materials in surface soils, residuum, and saprolite, as well as alluvium and colluvium, that overlies bedrock in most locations. The sandy-clay residuum or saprolite that composes the majority of the regolith thickness is derived from in-place weathering of bedrock and commonly contains fragments of solid rock that is highly fractured near a transition zone (Daniel and Harned, 1998) to bedrock. Residuum is produced from the weathering of feldspars and micas found in crystalline rocks to clays and differs from saprolite, which is weathered material that retains the structural characteristics of the parent rock (Chapman and others, 2005). The thickness of the regolith may vary considerably from near zero to greater than 150 ft. The regolith and transition zone contain intergranular porosity that provides the majority of water storage within the Piedmont and Blue Ridge aquifers (Heath, 1980). Storage potential of the regolith is dependent on a number of factors, including topographic position, thickness of the material, hydraulic gradients, and orientation of relict structure (saprolite) (White, 2012). An abrupt change in storage potential occurs at the bedrock interface because of the reduction in porosity from typical values of 35 to 55 percent in the regolith to 1 to 3 percent in the fractured bedrock (Daniel and Harned, 1998) (fig. 9). Hydraulic gradients are downward in recharge areas where water stored in the regolith slowly leaks downward into the bedrock fractures or moves laterally along the interface towards discharge points. In discharge areas, hydraulic gradients are reversed, and water from the fractured bedrock flows upward into the regolith (White, 2012).

Groundwater-level altitudes that were measured in December 2010 in Bedford County range from about 1,300 ft NGVD 29 in the northwest part of the county to about 600 ft NGVD 29 in the southeastern part of the county (fig. 10). The water-level map shown in figure 10 assumes streams are a surficial expression of the water level in adjacent aquifers, and the water-level altitude associated with streams were assigned by using a 20-meter digital elevation model (DEM). Groundwater generally flows from northwest to southeast in Bedford County, although groundwater flows locally towards drainages in eastern portions of the county where surface-water channels are deeply incised. Although the spatial coverage of the wells used to measure groundwater levels as part of this study was limited, in general, groundwater divides appear coincident with surface-water divides that separate the county's major surface drainages: Smith Mountain Lake, Goose Creek, Big Otter River, and the James River (fig. 10).

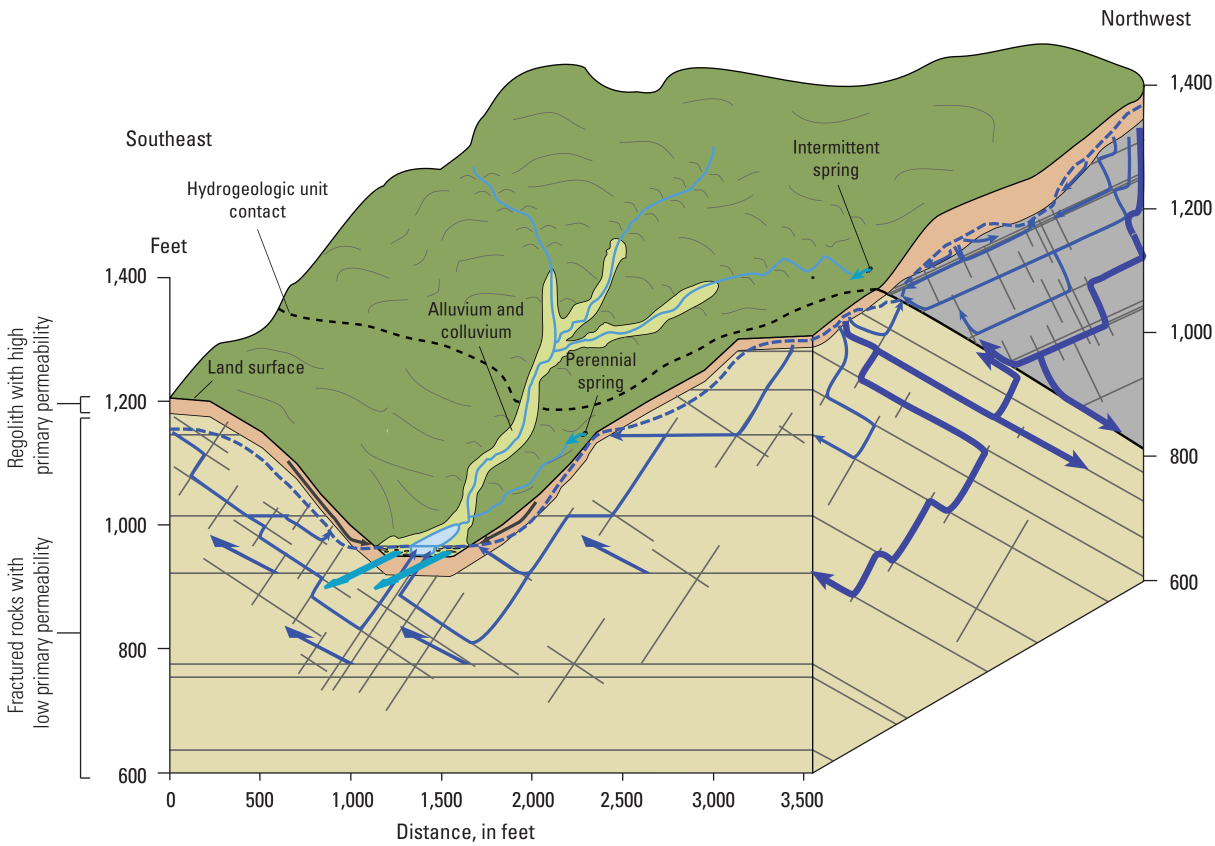


Figure 8. Generalized conceptual model of the flow system in Blue Ridge fracture rock aquifers in Bedford County, Virginia. [Modified from Nelms and Moberg (2010).]

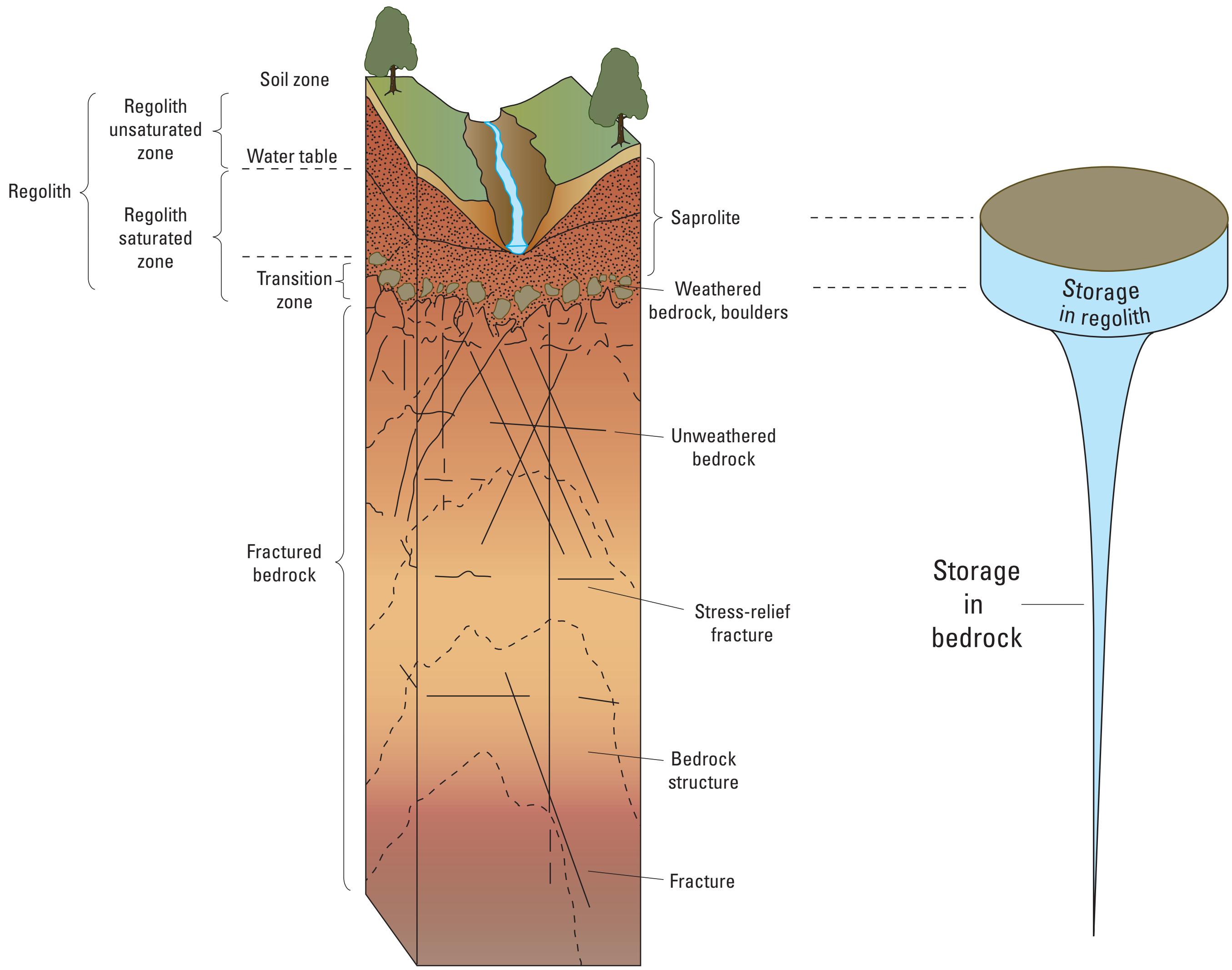


Figure 9. Components of the groundwater-flow system and relative storage volumes in the regolith and fractured-rock aquifers of the Piedmont Physiographic Province, Bedford County, Virginia. [Modified from Heath (1984) and Daniel and Harned (1998).]

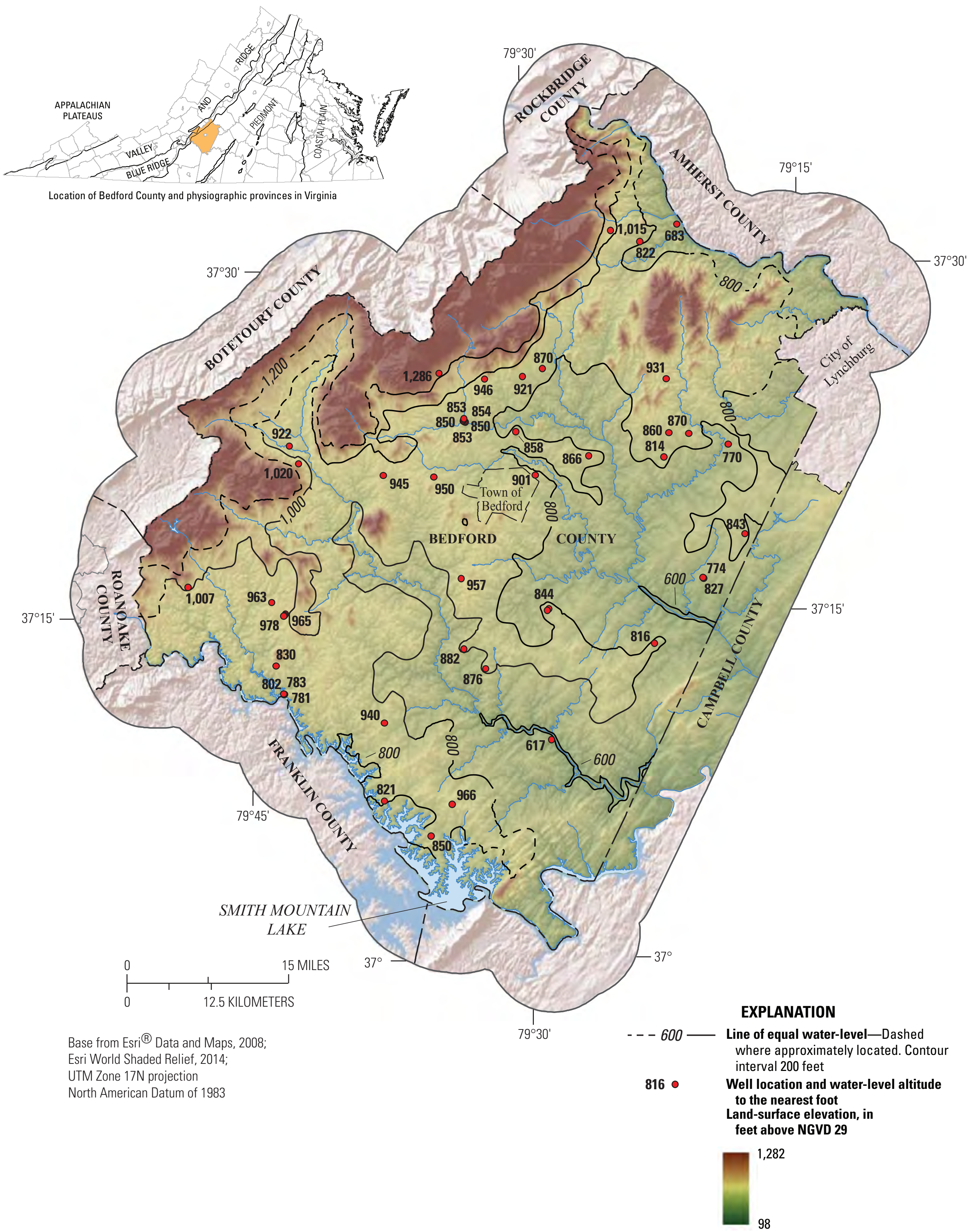


Figure 10. Groundwater levels in Bedford County, Virginia, December 2010.

Groundwater Response to Precipitation

The fractured-bedrock aquifers in the study area respond to natural transient atmospheric conditions, and the response is measured as changes in groundwater levels and base-level conditions in streams. The range of natural water-level fluctuation in the fractured-rock aquifer system varies across Bedford County and is controlled by (1) the location and magnitude of recharge, (2) topographic relief, or position in the local flow system, (3) amount of evapotranspiration (ET), particularly during the growing season, (4) hydraulic properties of the aquifer, (5) groundwater discharge to springs and streams, and (6) the location of pumped wells.

The differences between the December 2010 and April 2011 water levels measured at 47 wells averaged 1.0 ft (table 5); seasonal water-level fluctuations between the two measurement time periods were minimal (as noted from continuous hourly groundwater levels recorded in the study area for this time period) and are less than the resolution of the 200-ft contour intervals shown in figure 10. With the exception of two wells (33F 5 and 34H 6) that are thought to be influenced by recent pumpage, the maximum change from December 2010 to April 2011 was 6.12 ft (table 5).

Daily water-level data from October 2009 through September 2013 in continuous monitoring wells 33G 1 SOW 224, 33G 2 SOW 225, 35H 1 SOW 226, 35H 3 SOW 226A, and 35H 4 SOW 226B are shown in figure 11 with daily precipitation data from NOAA site 440561 Bedford 4 NW, the site closest to the three well locations (figs. 1, 3). The hydrographs for these wells indicate annual water-level variation of less than 10 ft, with subtle differences in amplitude and duration of interannual peaks. White (2012) prescribed the differences in response to precipitation to the variation in regolith storage, hydrologic communication with the surface, and depths of water-bearing fractures in the respective wells.

Water-level variation attributed to changing climate conditions during the study period can be quantified by using the Standardized Precipitation Index (SPI) (McKee and others, 1993). In the SPI, raw observations of precipitation are transformed to follow a normal distribution, and a single numeric value, equivalent to the number of standard deviations the observed data differ from the long-term mean, is assigned to observed precipitation data. For Bedford County, a 3-month time period was used to calculate SPI values from the precipitation data collected at NOAA site 440561 Bedford 4 NW, although a variety of time scales is possible. The SPI results during the water years 2009 to 2013 (fig. 11) indicate periods of below-normal precipitation during the winter of 2008–2009 and from March 2010 through June 2011, and periods of above-normal precipitation from July 2011 through May 2012 and January 2013 through September 2013.

Wells 33G 1 SOW 224, 33G 2 SOW 225, and 35H 1 SOW 226 had similar water-level trends that followed the 3-month SPI trends with broad peak lags of 3 to 4 months (White, 2012). The general absence of flashiness or response to individual precipitation events in the hydrographs of 33G 1 SOW 224, 33G 2 SOW 225, and 35H 1 SOW 226 for any of the years of record qualitatively implies either (1) a delayed response to subsurface flow within adjacent hillslopes, or (2) unconsolidated sediment in the regolith contains large accessible storage to modulate climatic inputs. For the first case, the rounded peaks are considered a result of an increase in distance, depth, or areal distribution of recharge to the aquifer. In the second case, permeability contrast at the bedrock interface likely provides additional moderation of the downgradient rate of leakage from the overlying regolith to the fractured-rock aquifers. Other factors such as prevailing and antecedent moisture conditions, hydrogeologic framework of the aquifer, proximity of a well to streams or other point of aquifer discharge, degree of aquifer confinement, or duration and intensity of seasonal events also may contribute to low-amplitude peaks and longer response times to precipitation.

Winter precipitation is a critical source of recharge to aquifers and is directly related to base flow in receiving streams during later periods of low flow in late summer and early fall (Austin, 2014). The response of aquifers to cold-season recharge was evident in hydrographs following heavy snows in the winter of 2009–10. Water levels in 33G 1 SOW 224 and 35H 1 SOW 226 rose approximately 2 ft and nearly 6 ft in 33 G2 SOW 225, the highest water levels recorded during the study period (fig. 11). The 2009–10 winter represents a weather extreme; however, the response of water levels to typical winter season precipitation have similar timing albeit lower magnitude of change than seen during the 2009–10 winter.

Replacement of 35H 1 SOW 226 with the nested piezometers 35H 3 SOW 226A (168-ft fracture zone) and 35H 4 SOW 226B (37-ft fracture zone) resulted in measurement of a 4.5- to 9.9-ft head differential between fractures at 37 ft and 168 ft below land surface. This range of measurements equates to a downward vertical gradient of 0.03 to 0.08. Water levels in the shallow fracture as measured in 35H 4 SOW 226B are more responsive to individual precipitation events than water levels in the deeper fracture measured in 35H 3 SOW 226A.

On August 23, 2011, a magnitude 5.8 earthquake was recorded near Mineral, Va., approximately 110 miles (mi) northwest of well 33G 2 SOW 225. This event was one of the largest earthquakes recorded in the eastern United States. Water levels in 33G 2 SOW 225 instantaneously lowered 0.74 ft in response to the propagation of seismic waves following the earthquake. Following the initial response to the earthquake, water levels in 33G 2 SOW 225 appeared to stabilize until September 26, 2011. During the next 4 days, through September 30, 2011, water levels rose 8.43 ft. The anomalous rise in water level followed the timing of the earthquake and daily rainfall totals of 4.8 in. (August 14, 2011) and 5.5 in. (September 6, 2011), the two highest daily rainfall totals observed during the study period.

Topography in mountainous settings has been related to differences in the range of water levels observed in valley and hilltop settings. Aquifer recharge and discharge are both largely driven by topography, such that valleys and hilltops serve as hydrologic boundaries. Proximity of wells to these boundaries is reflected in observed water levels. Seasonal water-level fluctuations tend to be greater in the elevated recharge areas and in areas underlain by low-permeability rocks; water levels in discharge areas near streams and springs and areas underlain by permeable rocks tend to fluctuate less (Nelms and Moberg, 2010). The water levels in Bedford County wells 33G 1 SOW 224 and 33G 2 SOW 225 exemplify these tendencies. Well 33G 2 SOW 225 is on a ridge at an elevation of 1,070 ft above mean sea level approximately 0.7 mi southeast of well 33G 1 SOW 224, which is in the adjacent valley at an elevation of 930 ft above mean sea level. Depth to water in the valley setting ranges from about 7 to 12 ft in 33G 1 SOW 224 and is much shallower than the measured range of depth to water of about 41 to 55 ft along the ridgeline in 33G 2 SOW 225 (fig. 11).

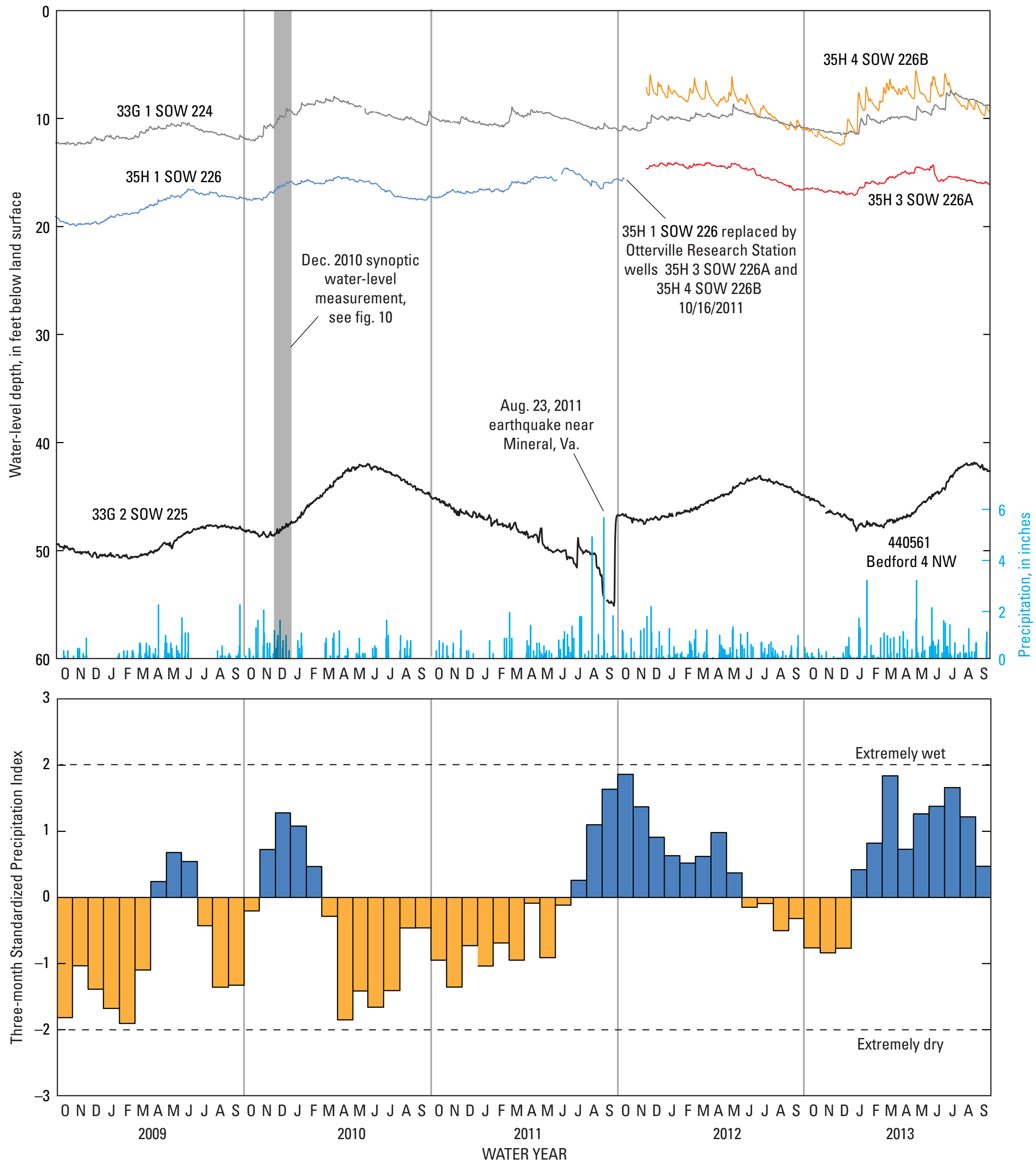


Figure 11. Water-level data from continuous monitoring wells and daily precipitation from National Oceanic and Atmposheric Administration site 440561 in Bedford County, Virginia, 2009–2013 water years.

Table 5. Water-level measurements from selected wells in Bedford County, Virginia.

[Site locations shown in figure 3. Abbreviations: USGS, U.S. Geological Survey; ft-blsd, feet below land-surface datum; —, no data reported]

Site number	USGS station number	December 2010		April 2011		December 2010–April 2011 change in water level (ft blsd)
		Date	Water level (ft-blsd)	Date	Water level (ft-blsd)	
32G 12	371621079482801	12/7/2010	46.48	4/18/2011	45.73	0.75
33F 2	371019079375301	12/6/2010	36.91	4/18/2011	37.46	–0.55
33F 3	371136079431901	12/6/2010	41.64	4/18/2011	41.22	0.42
33F 4	371138079431601	12/6/2010	79.7	4/18/2011	73.58	6.12
33F 5	371139079432001	12/6/2010	60.91	4/18/2011	48.05	12.86
33G 3	372106079374501	12/7/2010	0.57	4/18/2011	0.60	–0.03
33G 4	371538079435501	12/7/2010	33.51	4/18/2011	32.22	1.29
33G 5	371508079431001	12/7/2010	43.66	4/18/2011	39.81	3.85
33G 6	371505079431301	12/7/2010	6.85	4/18/2011	4.62	2.23
33G 7*	371503079431701	—	—	6/21/2011	7.42	—
34E 2	370645079341501	12/10/2010	45.3	4/18/2011	45.04	0.26
34E 3	370522079352601	12/10/2010	81.54	4/18/2011	80.27	1.27
33F 6	371252079434501	12/7/2010	167.51	4/18/2011	170.21	–2.70
34F 1	371330079333001	12/7/2010	28.78	4/18/2011	29.51	–0.73
34F 2	371237079321901	12/7/2010	47.36	4/18/2011	49.22	–1.86
34G 2	371634079333501	12/7/2010	45.34	4/19/2011	46.37	–1.03
34G 3	372100079345901	12/8/2010	5.08	4/18/2011	3.59	1.49
34H 4	372256079303001	12/8/2010	196.73	—	—	—
34H 5	372531079343701	12/6/2010	33.29	4/18/2011	29.43	3.86
34H 6	372514079320801	12/6/2010	60.09	4/18/2011	41.07	19.02
34H 7	372323079331301	12/6/2010	2.9	4/18/2011	2.58	0.32
34H 8	372324079331701	12/6/2010	1.05	4/18/2011	3.59	–2.54
34H 9	372326079332001	12/6/2010	7.07	4/18/2011	6.16	0.91
34H 10	372329079331901	12/6/2010	6.97	4/18/2011	6.26	0.71
34H 11	372332079331801	12/6/2010	11.88	4/18/2011	11.16	0.72
35F 1	371337079230701	12/9/2010	41.6	4/18/2011	42.49	–0.89
35F 2	370929079284801	12/10/2010	88.22	4/18/2011	89.15	–0.93
35G 1	371511079284901	—	—	4/18/2011	51.60	—
35G 2	371508079285501	12/10/2010	36.75	4/18/2011	37.04	–0.29
35G 3	372150079263101	12/8/2010	43.9	4/19/2011	45.05	–1.15
35G 4	372102079292801	12/8/2010	25.37	4/18/2011	28.23	–2.86
35H 2	372539079285901	12/8/2010	51.98	4/19/2011	51.03	0.95
35J 4	373108079233301	12/8/2010	67.29	4/19/2011	69.73	–2.44
35J 5	373138079250801	12/8/2010	41.07	4/19/2011	39.45	1.62
36G 1	371819079180501	12/9/2010	27.75	4/18/2011	29.88	–2.13
36G 2	372214079185501	12/9/2010	41.11	4/19/2011	42.40	–1.29
36G 3	372143079222501	12/9/2010	38.31	4/19/2011	37.33	0.98
36G 4	371625079202301	12/9/2010	42.28	4/18/2011	42.32	–0.04
36G 5	371627079202501	12/9/2010	18.47	4/18/2011	17.97	0.50
36H 1	372246079220901	12/9/2010	32.33	4/19/2011	33.24	–0.91
36H 2	372244079210401	12/9/2010	30.42	4/19/2011	29.80	0.62
36H 3	372508079221401	12/8/2010	82.38	4/18/2011	81.49	0.89
36J 7	373151079213001	12/8/2010	42.88	4/19/2011	42.12	0.76

* Single measurement made at well in June 2011.

Water Budget

A water budget is an estimate of water entering and leaving a basin plus or minus storage changes for a given time period. Water enters a basin as precipitation and leaves as streamflow, ET, and diversions, such as surface-water withdrawals and groundwater pumpage. The conceptualization of flow in Bedford County suggests that groundwater and surface-water divides are coincident and that groundwater does not enter or leave a surface-water basin as underflow. Based on that conceptualization, a simple water budget for a typical watershed in Bedford County can be described by the following equation:

$$PR = ET + SF + \Delta S, \tag{1}$$

where

- PR is the mean precipitation, in inches per year,
- ET is the mean evapotranspiration, in inches per year,
- SF is the mean streamflow, in inches per year, and
- ΔS is the change in groundwater storage, in inches per year.

Streamflow can be divided into two components as

$$SF = RO + BF, \tag{2}$$

where

- RO is runoff from land surface, in inches per year, and
- BF is base-flow discharge from aquifers, in inches per year.

For the purposes of this water-budget analysis, ΔS is assumed equal to zero because all recharge from precipitation is assumed to discharge to streams as base flow. All terms in the water-budget equation are known or can be estimated except *ET*, and the equation is solved for *ET*. Deviations from the assumptions of the equation, such as underflow between basins, and errors in other terms, are, therefore, included in *ET*.

Inflows to Aquifer

Precipitation that falls on Bedford County is the sole source of inflow to the groundwater system. Climate records of precipitation and hydrologic records of streamflow are the most available data with which to assess the variability in hydrologic budget components of Bedford County. The magnitude and distribution of groundwater supplies in Bedford County was quantified by using streamflow data collected at continuous and partial-record stations and estimates of return flow from domestic wastewater systems.

Precipitation

The PRISM (Parameter-elevation Regressions on Independent Slopes Model) Climate Group at Oregon State University provides yearly and monthly precipitation datasets (Daly and others, 2008). The average annual precipitation for Bedford County was extracted from the National PRISM data (fig. 12; PRISM Climate Group, Oregon State University, 2014). The gridded precipitation data are based on a model of the National Weather Service (NWS) station data for the normal climatological period 1981–2010. Mean annual precipitation rates listed in table 6 are the average of 1981–2010 PRISM precipitation normals (fig. 12) within respective locality or basin boundary defining polygon. PRISM data indicate that average precipitation ranges from 39.4 to 66.9 inches per year in Bedford County. The orographic influence of the Blue Ridge Mountains is shown by the higher values in the western portions of the Big Otter River Basin. The lowest values are found in the southeastern portion of Bedford County near the mouth of Goose Creek.

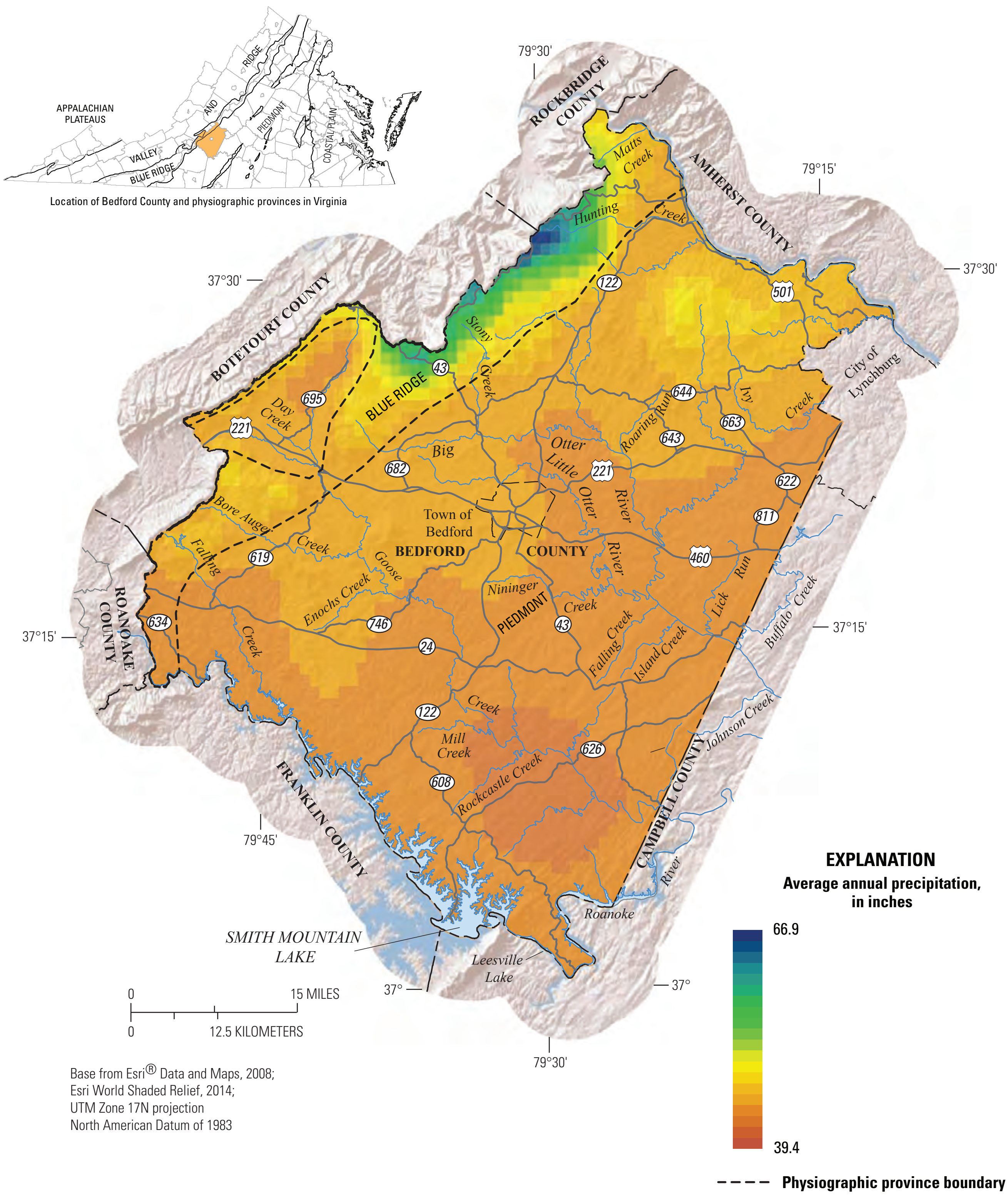


Figure 12. Average annual precipitation across Bedford County, Virginia, based on parameter-elevation regressions on independent slopes model (PRISM) (PRISM Climate Group, Oregon State University, 2014). The normal values are based on the National Weather Service's current normal climatological period from 1981 to 2000.

Table 6. Hydrologic budget components for select localities and basins in Bedford County, Virginia.

[—, not determined; na, not applicable; Rt, Route]

Locality or station name	Station number	Mean annual precipitation, in inches[1]	Evapotranspiration: Total, in inches per year	Evapotranspiration: Riparian evapotranspiration, in inches per year	Runoff, in inches per year	Effective recharge, in inches per year	Base-flow index, in percent	Period of record
Bedford County, Virginia[2]		45.6	29.2	1.2	3.8	13.7	—	na
City of Bedford, Virginia[2]		45	25.7	1.2	8.3	12.3	—	na
Goose Creek Basin								
Goose Creek near Huddleston, VA[3]	02059500	44.2	31.4	—	4.5	8.3	65	1931–2012
Goose Creek at Rt 747 near Bunker Hill, VA[3]	02059485	45.7	35.9	—	3.3	6.5	66	2007–2012
North Fork Goose Creek at Rt 460 near Montvale, VA[4]	02059420	46.4	38.7	—	2.7	5.0	65	2006–2009
South Fork Goose Creek at Rt 607 at Montvale, VA[4]	02059450	46.4	35.8	—	3.7	6.9	65	2006–2009
Goose Creek at Rt 726 near Irving, VA[4]	02059452	46.4	37.3	—	3.2	6.0	65	2006–2009
Bore Auger Creek at Rt 754 near Irving, VA[4]	02059459	46.2	35.6	—	3.7	6.9	65	2006–2009
Stony Fork at Rt 608 near Moneta, VA[4]	02059490	44.9	34.2	—	3.7	7.0	65	2006–2009
Big Otter River Basin								
Big Otter River near Evington, VA[3]	02061500	44.6	30.5	—	4.7	9.3	66	1938–2012
Big Otter River at Rt 221 near Bedford, VA[3]	02061000	47.8	35.1	—	3.8	8.9	70	2007–2012
Sheep Creek at Rt 688 near Thaxton, VA[5]	02060592	50.2	42.0	—	2.7	5.4	66	2006–2009
Stony Creek at Rt 43 near Peaks of Otter, VA[5]	02060692	50.2	36.7	—	4.5	8.9	66	2006–2009
Big Otter River at Rt 43 near Peaks of Otter, VA[5]	02060695	50.2	37.8	—	4.2	8.2	66	2006–2009
North Otter Creek at Rt 643 near Cifax, VA[5]	02060800	50.4	34.6	—	5.3	10.5	66	2006–2009
Oslin Creek at Rt 644 near Cifax, VA[5]	02060860	45.9	34.4	—	3.9	7.6	66	2006–2009
Elk Creek at Rt 668 near Goode, VA[5]	02061160	45.5	34.3	—	3.8	7.5	66	2006–2009
Little Otter River at Rt 715 near Otter Hill, VA[5]	02061320	44.8	34.9	—	3.3	6.5	66	2006–2009

[1]Data from 1981–2010 PRISM raster dataset (PRISM Climate Group, Oregon State University, 2014).

[2]Data from Sanford and others (2012) chemograph separation.

[3]Runoff, recharge, and BFI data computed using PART.

[4]Denotes partial record station with value determined by log-log dischage relation with 02059500 Goose Creek near Huddleston, VA.

[5]Denotes partial record station with value determined by log-log dischage relation with 02061500 Big Otter River near Evington, VA.

Estimates of Recharge from Streamflow

Precipitation that infiltrates into the soil and percolates to the water table recharges the groundwater system. The amount of recharge depends on many factors, including antecedent soil-moisture conditions, the timing, duration, and intensity of precipitation, depth to the water table, and soil, regolith, and bedrock characteristics. Generally, recharge areas coincide with topographic highs in an area, whereas topographic lows are commonly discharge areas. Because of climatic variability, the amount of recharge is expected to vary from year to year.

Because the water table is relatively shallow and streams in Bedford County are assumed to be gaining, or receiving groundwater discharge, base flow is used as an approximation of recharge. Annual recharge rates were estimated based on streamflow records and the hydrograph separation technique PART (Rutledge, 1998). PART was used to separate streamflow into its groundwater discharge (BF) and surface-runoff (RO) components and to estimate groundwater recharge under the previously mentioned assumptions. PART also generates annual estimates of base-flow index (BFI), which is the percentage of streamflow that is accounted for by BF. Streamflow data from the four continuous-record stream gages in the study area were analyzed for their respective entire periods of record (tables 3, 6). Graphical regression methods were used to relate measurements at the partial-record sites (table 3) to concurrent daily mean discharge at their respective downstream long-term continuous data stations (02061500 Big Otter River near Evington or 02059500 Goose Creek near Huddleston; fig. 13). A curve was visually fitted to the data points, and mean base-flow discharge was estimated by transferring the mean base-flow discharge from the continuous stations through the relation line to the partial-record station based on the methods of Harlow and others (2005).

Base-flow discharge is commonly assumed to be equivalent to effective recharge; however, it is not the total recharge for a basin. Total recharge is always larger than effective recharge and includes riparian evapotranspiration (RET), which is the quantity of water evaporated or transpired by plants in the riparian zone adjacent to streams. RET is also a component of total ET and is included in the ET component of the water-budget estimates presented in table 6.

Using PART, the average annual effective recharge for continuous stream gage 02059500, Goose Creek near Huddleston, Va., during the period 1931–2012 was 8.3 in/yr (table 6), with base-flow discharge composing 65 percent of mean streamflow. During the 2006–2009 period of partial-record station data collection, average annual effective recharge at continuous stream gage 02059500, Goose Creek near Huddleston, Va., was 5.4 in/yr, and BFI was 67 percent. The 2006–09 average annual effective recharge at continuous stream gage 02059500 is a decrease from the 1931–2012 average annual effective recharge at the gage by about 35 percent, which is equivalent to a decrease of approximately 26.9 million gallons per day (Mgal/d) over the 188-mi^2 drainage area.

Continuous stream gage 02059485, Goose Creek at Route 747 near Bunker Hill, Va., is upstream of 02059500 about 4 mi south of the town of Bedford and was constructed in December 2006 as part of the current investigation (fig. 1). The average annual effective recharge for this station for 2007–12 was 6.5 in/yr, with base-flow discharge composing 66 percent of mean streamflow. A log-log relation of base-flow discharge at five partial-record streamflow sites in the Goose Creek Basin (02059420, 02059450, 02059452, 02059459, and 02059490) and concurrent daily mean discharge at the continuous station 02059500 Goose Creek near Huddleston, Va., yielded base-flow values ranging from 5.0 to 6.9 in/yr during the period 2006–09 (table 6; figs. 14–15).

The average annual effective recharge for 02061500, Big Otter River near Evington, Va., during the period 1938–2012 was 9.3 inches per year, with base-flow discharge composing 66 percent of mean streamflow (table 6). During the 2006–09 period of partial-record station data collection, average annual effective recharge at 02061500, Big Otter River near Evington, Va., was 6.5 in/yr, and BFI was 68 percent. The 2006–09 average annual effective recharge at continuous stream gage 02061500 was a decrease from the 1938–2012 average annual effective recharge by about 30 percent, which is equivalent to a decrease of approximately 26.9 Mgal/d over the 315-mi^2 drainage area.

At 02061000, Big Otter River at Route 221 near Bedford, Va., the average annual effective recharge for 2007–12 was 8.9 in/yr, with base-flow discharge composing 70 percent of mean streamflow. Partial-record streamflow measurements conducted at seven sites in the Big Otter River Basin (02060592, 02060692, 02060695, 02060800, 02060860, 02061160, and 02061320) were evaluated with concurrent daily mean discharge values at station 02061500 Big Otter River near Evington, Va. The average annual effective recharge values for the seven partial-record stations ranged from 5.4 to 10.5 in/yr during the period (2006–09) (table 6; figs. 14 and 16).

Sanford and others (2012) computed water budgets at two gages in Bedford County by using chemograph separation techniques. They spatially extrapolated the computed water budgets across the entire State of Virginia through a regression analysis of ET and runoff with land use and basin characteristics. Sanford and others (2012) used political boundaries to summarize their water-budget results. Chemograph separation techniques by Sanford and others (2012) for the town of Bedford and Bedford County areas yielded approximately 30 percent higher effective recharge values than those generated from PART for the period of record at 02059500 and 02061500 (table 6). Hydrograph separation techniques are intuitive and follow the assumption that peak flows are dominated by surface-water runoff. Water in streams, however, will retain the chemical character of the groundwater and surface-runoff inputs throughout the peak event. Where groundwater chemistry is known or can be established, stream chemograph separation techniques have been shown to consistently yield higher base-flow values (Sanford and others, 2012).

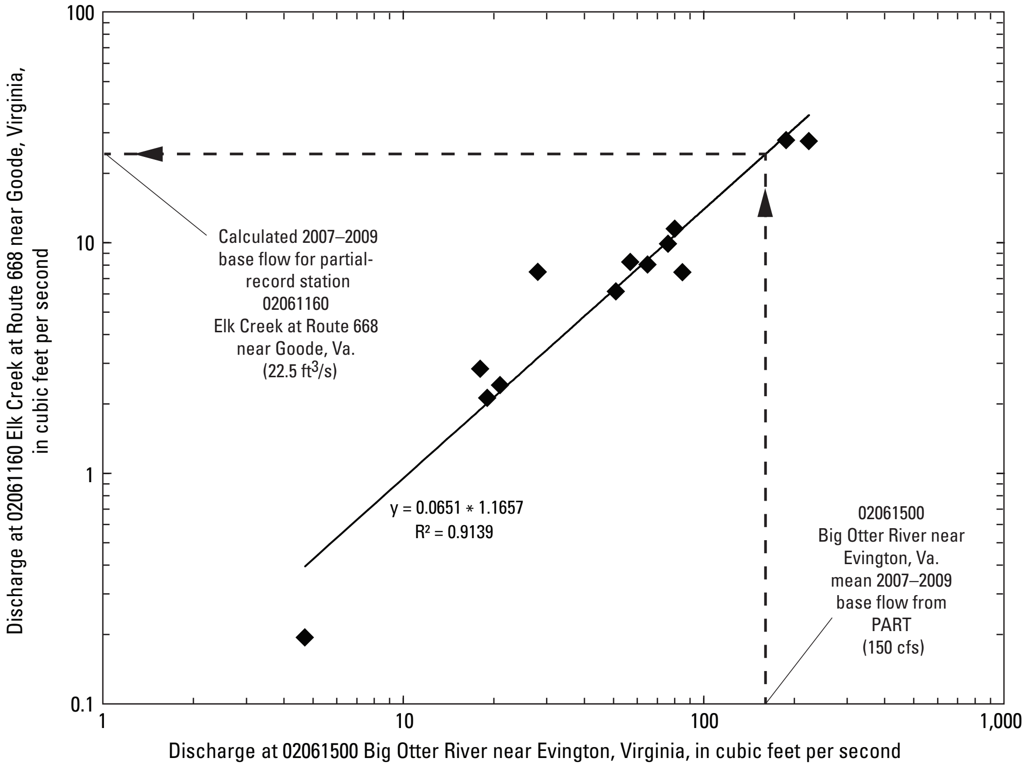


Figure 13. Relation of discharge measured at partial-record station 02011600 Elk Creek at Route 668 near Goode, Virginia, with concurrent mean daily discharge at long-term stream-gaging station 02061500 Big Otter River near Evington, Virginia.

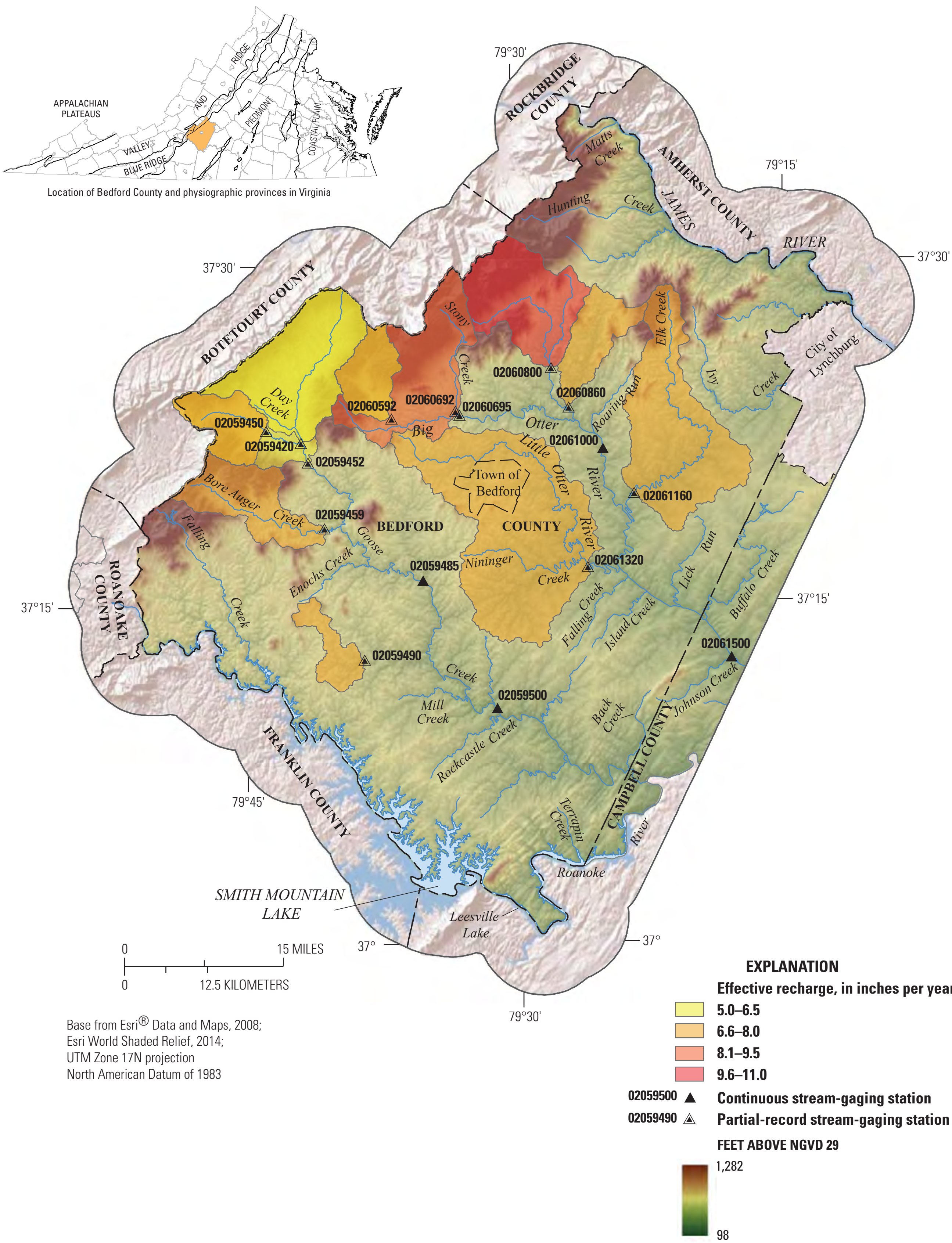


Figure 14. Effective recharge as base flow calculated at partial-record stream-gaging stations in Bedford County, Virginia, 2006–2009.

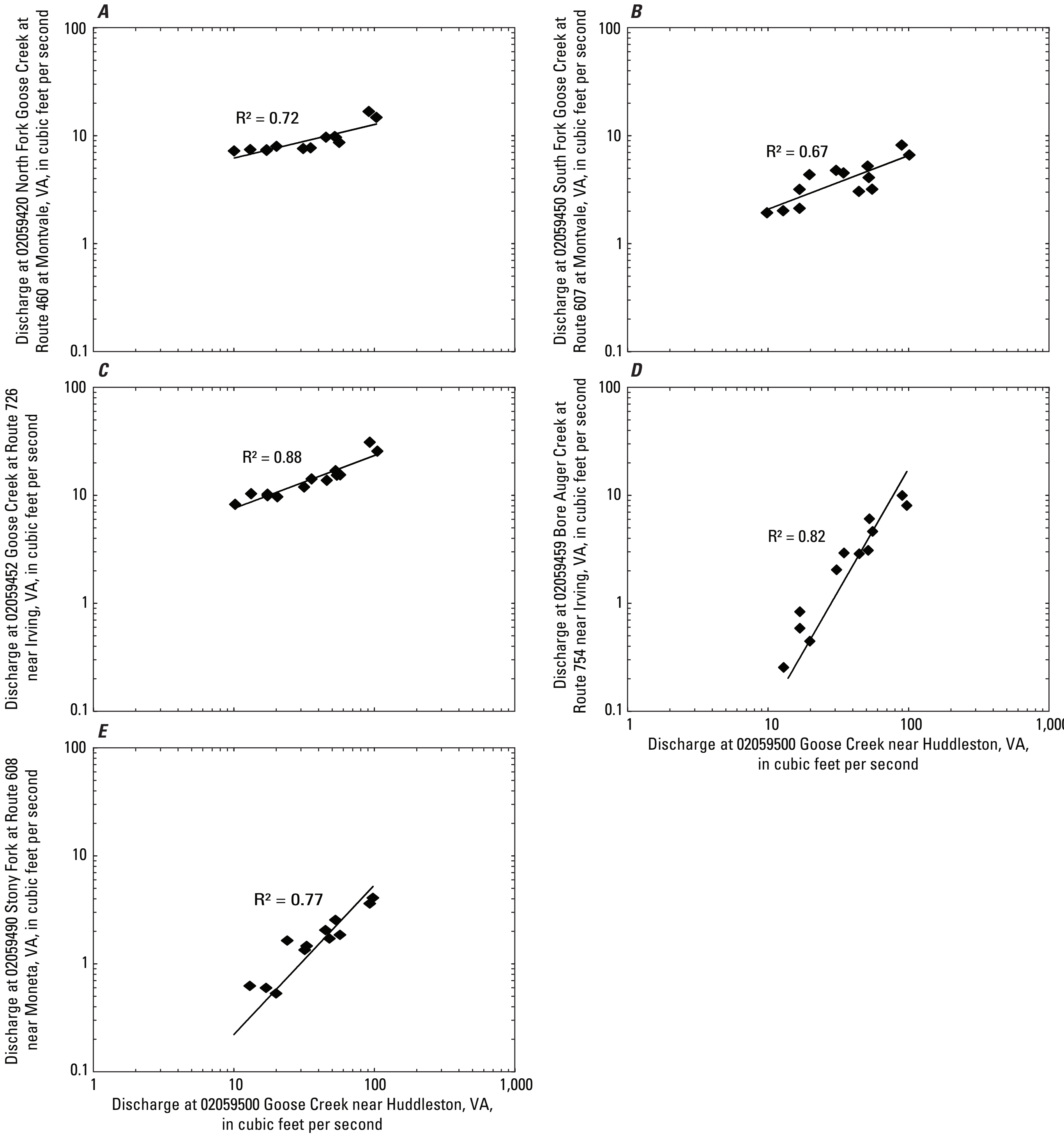


Figure 15. Relation of discharge measured at partial-record stations in the Goose Creek Basin to concurrent mean daily discharge at long-term stream-gaging station 02059500 Goose Creek near Huddleston, Virginia.

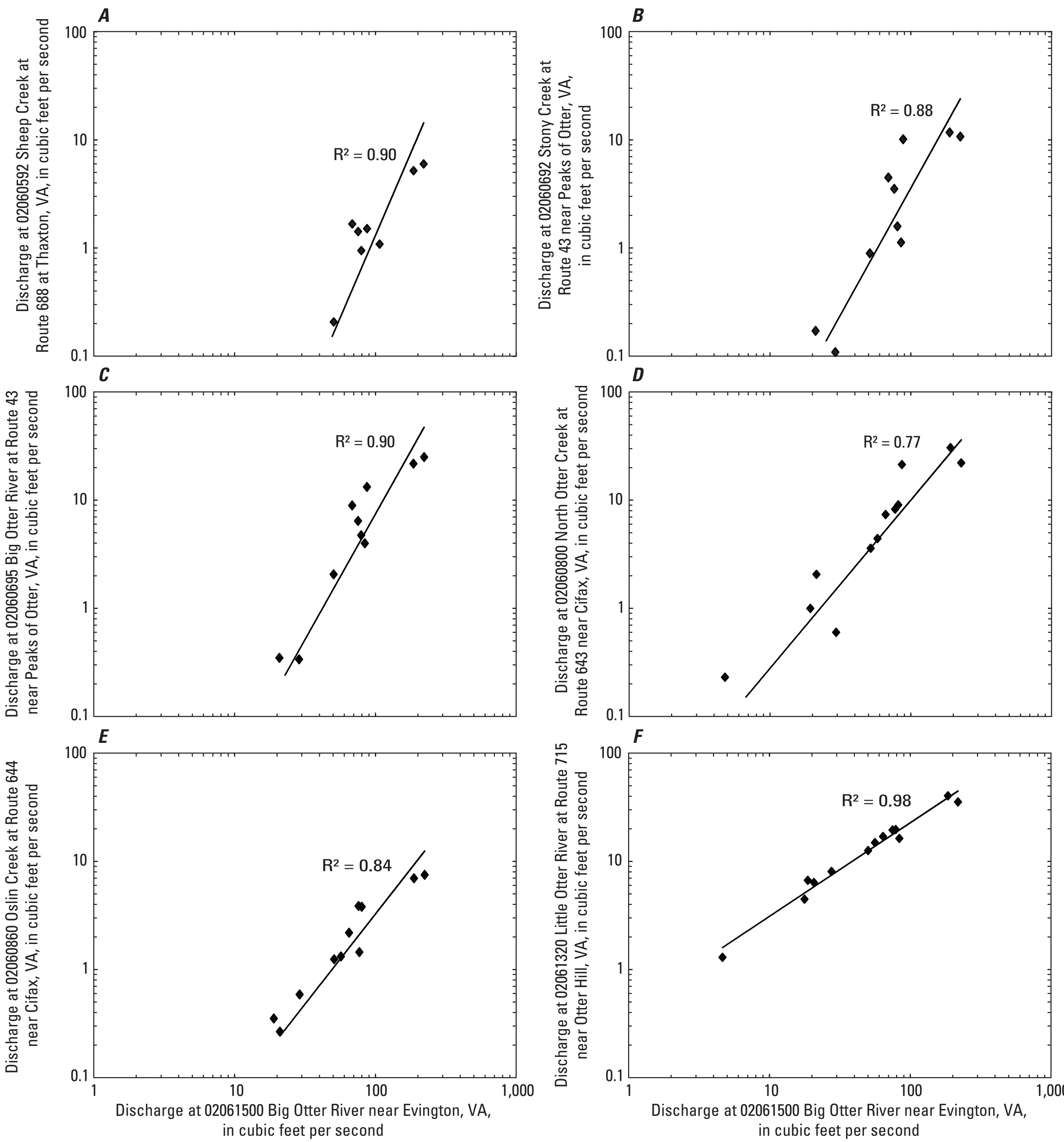


Figure 16. Relation of discharge measured at partial-record stations in the Big Otter River Basin to concurrent mean daily discharge at long-term stream-gaging station 02061500 Big Otter River near Evington, Virginia.

Effluent from Septic Systems

For a typical domestic household discharging to a septic system, 84 percent of the indoor water use is assumed to return to the aquifer (Horn and others, 2008). Approximately 43,200 residents who relied on domestic wells for water supplies in Bedford County in 2005 (Kenny and others, 2009) are assumed to have also discharged wastewater to a typical household septic system. If per-capita use from domestic wells is 75 gallons per day (gal/d) (Hutson and others, 2004), then 2.8 Mgal/d (0.8 in/yr) of water is returned to the aquifer by infiltration from septic systems. Return flow from septic systems can potentially mitigate adverse effects of aquifer pumping, but the fate of water discharged into the shallow regolith by septic systems and the rate and timing at which wastewater will recharge to deeper bedrock aquifers are unknown.

Outflows from Aquifer

The hydrologic budget for any basin must balance the quantity of water entering the basin with the quantity of water leaving the basin. Outflows from aquifers in Bedford County include ET and groundwater withdrawals. Groundwater is an important source of freshwater withdrawn from the Piedmont and Blue Ridge fractured-rock aquifers for several uses, including drinking water, and for commercial, industrial, mining, and agricultural purposes.

Water Use

Annual water-use data for Bedford County were compiled from the Virginia Water Use Data System (VWUDS), a State-mandated reporting system for surface- and groundwater withdrawals exceeding 10,000 gal/d in any single month. VWUDS provides data on specific water-use categories in Virginia that are summarized on a county level as part of the USGS national water-use estimates (Kenny and others, 2009). The USGS summary also includes computed daily estimated withdrawal rates attributed to private domestic use based on a published per-capita withdrawal coefficient of 75 gal/d (Hutson and others, 2004). All withdrawals are reported in million gallons per day.

Bedford County relies on surface-water sources to meet most of its residential and industrial demands (fig. 17). Surface-water withdrawals primarily support industry along the James River and public drinking water supplies for the town of Bedford (Virginia Department of Environmental Quality, oral commun., 2012). Between the years of 1985 and 2005, withdrawals from surface-water sources increased from 12.8 to 13.6 Mgal/d (Kenny and others, 2009). The proportion of surface-water withdrawals with respect to total withdrawals, however, has steadily declined from 82 to 74 percent during this period (fig. 17).

Total groundwater withdrawals from bedrock aquifers in Bedford County increased from about 2.80 Mgal/d in 1985 to about 4.85 Mgal/d in 2005 (fig. 17; Kenny and others, 2009). Domestic withdrawals of groundwater also increased between 1985 and 2005 from 2.37 Mgal/d to 3.24 Mgal/d, ranging from 67 to 94 percent of the total, likely as a result of the increase in county population by 37,000 residents during that period. The domestic withdrawal estimate includes small community systems that supply less than the 10,000-gal/d State-mandated reporting limit.

The town of Bedford operates five public-supply wells as a backup supply when surface-water sources are insufficient to meet public-supply demands. Groundwater used for public supply in Bedford County was less than 10 percent of the total public supply for 1985, 1995, and 2005 (fig. 17). Groundwater withdrawals used for public supply were about 0.17 Mgal/d, or 6 percent of the total groundwater withdrawals, in 1985, and were about 0.07 Mgal/d, or 1 percent of the total groundwater withdrawals, in 2005 (Kenny and others, 2009).

The proportion of groundwater use attributed to mining varied considerably between 1985 and 2005. Groundwater withdrawals for mining purposes from a single quarry dewatering operation accounted for 1.27 Mgal/d, or 26 percent of the total groundwater used, in 2005. Increases in groundwater withdrawals used in mining from 1985 to 2005 likely reflect inconsistencies in data reporting.

Evapotranspiration

For each basin, total ET was computed by subtracting the PART-derived mean streamflow at individual gages from the PRISM-derived mean value of precipitation (fig. 12) over the respective drainage areas for each gage (tables 3, 6). Long-term estimates of total ET accounted for 68 percent (30.5 in/yr) of precipitation falling in the Goose Creek Basin (station 02059500) and 71 percent (31.4 in/yr) of precipitation falling in the Big Otter River Basin (station 02061500). These estimates of total ET are within the previous range of ET values (23.7 to 31.5 in/yr) for Bedford County that were extrapolated from national datasets (Sanford and Selnick, 2013). Most of the water removed by ET is rerouted to the atmosphere by plant uptake prior to reaching the groundwater table. RET or direct discharge from the water table by plant uptake is included in estimates of total ET, but cannot be estimated from the PART data. Rutledge and Mesko (1996) noted that RET generally ranges between 1 and 2 in/yr in the Appalachian Valley and Ridge, the Piedmont, and the Blue Ridge from Alabama to New Jersey, and Sanford and others (2012) estimated an RET of 1.2 in/yr for Bedford County (table 6).

Discharge to Surface Waters and Subsurface Outflow

As previously discussed, groundwater discharges from the groundwater system in Bedford County as base flow to streams and rivers. In the previously described conceptual model, the rates of discharge to surface-water bodies are assumed to be equal to the rate at which the aquifer is recharged by precipitation, which ranges from 5.0 to 13.7 in/yr (table 6) depending on the method of analysis and the prevailing climatic conditions. Stream gages permit direct measurement of stream discharge and estimation of base flow for areas draining to Big Otter River and Goose Creek. Direct groundwater discharge to the James River or Smith Mountain Lake are assumed to be similar in magnitude to that measured in the Big Otter River and Goose Creek drainages, but are more fully explored in the following sections of this report.

Subsurface outflow, or groundwater that originates in Bedford County and discharges to surface-water bodies outside of the county, is assumed to be minimal. The northern and southern boundaries of Bedford County are major regional drains that likely function as controlling hydrologic boundaries on the outflow of adjacent fractured-rock aquifers.

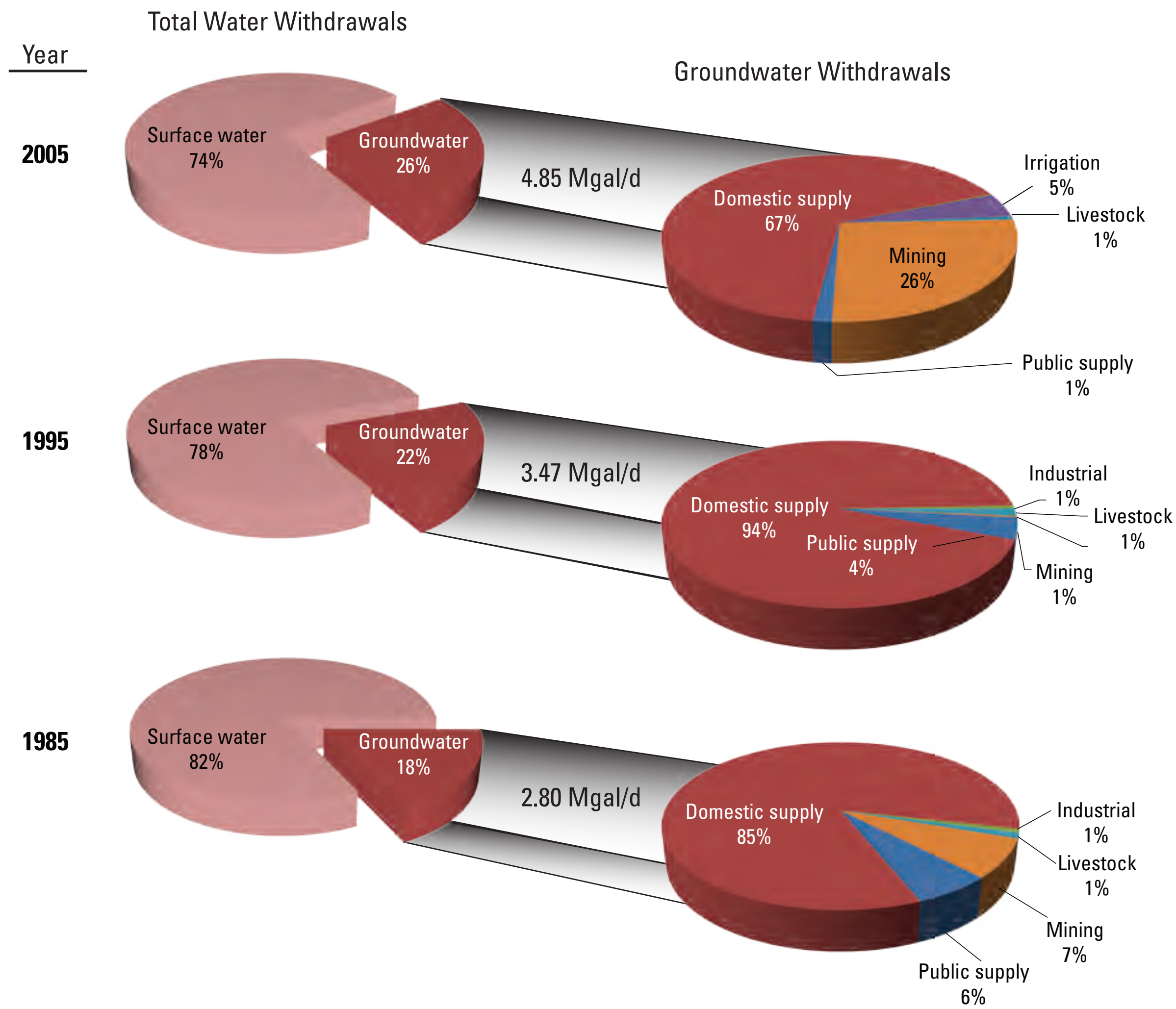


Figure 17. Water-use estimates for Bedford County, Virginia, 1985–2005 (from Kenny and others, 2009).

Simulation of Groundwater Flow

Groundwater flow through the fractured-rock aquifer underlying Bedford County was simulated by using MODFLOW (Harbaugh, 2005), a three-dimensional finite-difference model. The application of MODFLOW was based on the common assumption that flow through the fracture network at the scale of miles can be represented as flow through porous media. This approach has been used in other simulation studies of groundwater flow through crystalline rocks (Tiedeman and others, 1997). The groundwater-flow model used a steady-state simulation to represent equilibrium conditions for the aquifer system in December 2010 (fig. 10), a period when water levels were near the median levels measured in continuous observation wells (fig. 11). Model simulations were used to compute a groundwater budget including sources of recharge and groundwater discharge for the four major drainages within the study area. Previously discussed groundwater budgets are constrained by watershed boundaries and do not include estimates for all areas in Bedford County. Model estimates therefore provide additional coverage of areas in Bedford County outside of gaged basins.

Model Design

The model domain was divided into a uniform grid of 820-ft cells aligned along 229 rows and 280 columns. The grid was oriented at an azimuth of 320º to best align model boundaries with major surface-water features. The western model boundary coincided with the upper watershed divides of the major drainages: the James River, Big Otter River, and Goose Creek (fig. 18). The northern and southern model boundaries were delineated along the James River and the Roanoke River, respectively. The eastern model boundary coincided with local surface-water divides of the major drainages downstream of the stream gages that were used to measure streamflow within the study area. The resulting 975-mi^2 model domain contains 40,424 active cells (fig. 18).

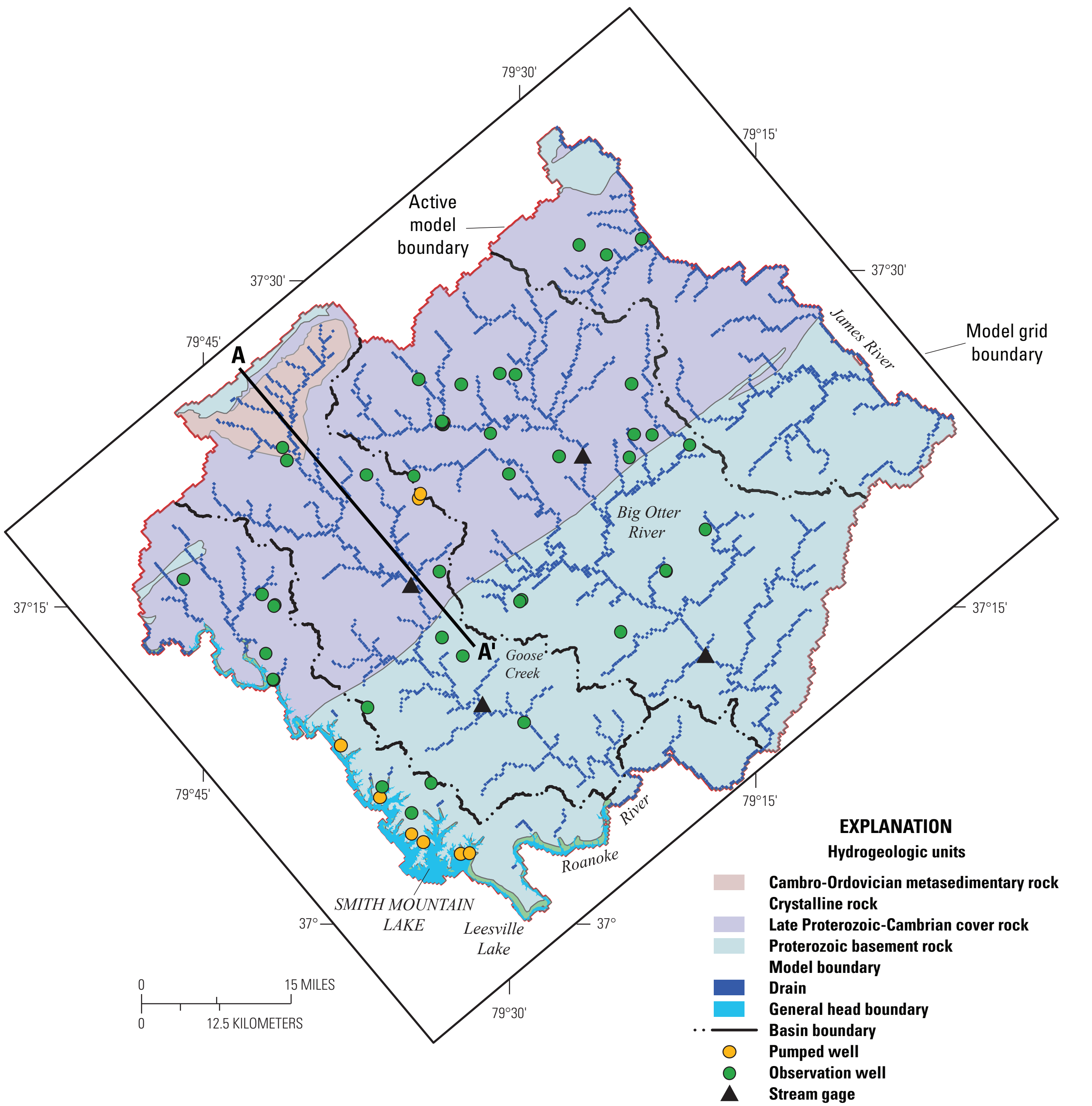


Figure 18. Model domain, boundary conditions, and hydrogeologic units delineated in the Bedford County, Virginia, groundwater model.

Boundary Conditions

Three model layers of increasing thickness with depth were used to represent the upper 655 ft of crystalline (Proterozoic basement and Late Proterozoic-Cambrian cover crystalline rocks) and Cambro-Ordovician metasedimentary rock aquifers. The model layers are parallel and were constructed to form a subdued reflection of the land surface. Model layer 1, the top layer, has minimum and maximum thicknesses of 130 and 425 ft, respectively, and represents the regolith and upper weathered portion of bedrock. In areas of high relief in the western portion of the model domain, the thickness of layer 1 reaches the maximum value of 425 ft and extends into unweathered bedrock so that the bottom of layer 1 is smoother than the land surface (fig. 19). Model layers 2 and 3 have uniform thicknesses of 200 and 330 ft, respectively, and represent the fractured bedrock aquifers.

The western, eastern, and bottom model boundaries were specified as no-flow (impervious to flow). Head-dependent (GHB) boundaries were specified in layer 2 to represent discharge to Smith Mountain Lake and Leesville Lake along the southern model boundary (fig. 18); the head elevation of the GHB boundaries were specified as the mean lake stages (795 and 615 ft, respectively). Drain boundaries were specified in layer 1 within the model domain and along the northern boundary to represent groundwater discharge to the major drainages. Head elevations of the drain boundaries were calculated from a 100-ft DEM. The conductance C of the GHB and drain boundaries is defined as

$$C = \frac{KA}{L}, \tag{3}$$

where

K is hydraulic conductivity,

A is the cross-sectional flow area of the boundary, and

L is the flow-path length across the boundary.

For the GHB boundary, the flow area A is WB where W is the cell width and B is the layer thickness. If the flow-path length L is assumed equal to W, then the conductance can be computed as KB, the transmissivity of the model cell.

For drain cells, the flow-path length L is the streambed thickness, which is assumed to be 3 ft. The flow area A was computed for each cell as the product of the length of the stream channel crossing the cell and an assumed stream width. Stream widths were based on the location of the channel within the drainage network. The widths of the channels of the James River and Roanoke River ranged from 16 to 330 ft. The widths of the channels of the Big Otter River and Goose Creek ranged from 16 to 65 ft. The hydraulic conductivities K for the GHB and drain boundaries were insensitive model parameters and were specified as 3 feet per day (ft/d) so that groundwater discharge through the boundaries was limited by the 1.3- to 1.7-ft/d hydraulic conductivities of the adjacent crystalline fractured-rock aquifers (table 7).

The top model boundary represents recharge as a specified flux. The recharge rate was spatially variable and assumed to be proportional to the available water, calculated as the difference between mean annual precipitation from PRISM (Daly and others, 2008) and actual ET. ET was estimated from a regression based on national climate and land-cover data by Sanford and Selnick (2013). The proportion of the available water that was diverted to recharge (*Rch*) was estimated through model calibration, discussed further on. Values of *Rch* were estimated for three areas within the model domain with different degrees of relief, ranging from low to steep. Grid cells were assigned to low, moderate, or steep slope classes (less than 770 feet per mile [ft/mi], 770 to 1,540 ft/mi, and more than 1,540 ft/mi, respectively) by using a 100-ft DEM to estimate the average surface slope within each cell. Pumping from 14 municipal wells was represented by constant flux boundaries in layer 2. Reported well depths from 7 of the 14 municipal wells averaged 366 ft. Pumping rates ranged from 1.3 to 12.6 Mgal/d and were estimated from average monthly groundwater withdrawals recorded in VWUDS for 2005.

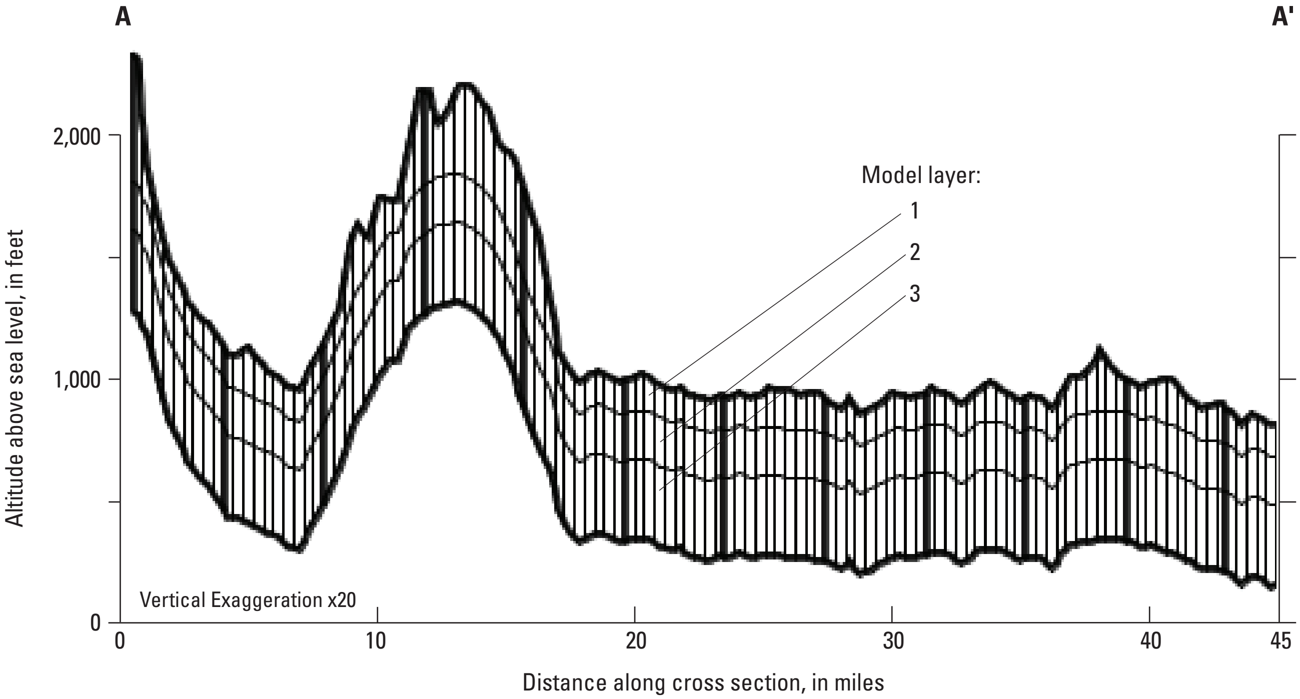


Figure 19. Section A-A′ showing thickness of model layers beneath areas with high relief, Bedford County, Virginia. Location of section shown in figure 18.

Table 7. Hydraulic properties estimated and specified in Bedford County groundwater-flow model.

[ft/d, foot per day; ft, foot; Shaded values were specified]

Parameter	Abbreviation	Value	Approximate 95-percent confidence interval	Coefficient of variation, percent
		Horizontal hydraulic conductivity, ft/d		
Cambro-Ordovician metasedimentary rocks	$K_{h\text{-}carb}$	15		
Proterozoic basement crystalline rocks	$K_{h\text{-}base}$	1.9	0.7 to 1.9	51
Late Proterozoic-Cambrian cover crystalline rocks	$K_{h\text{-}cover}$	1.1	0.3 to 2	32
Faults	$K_{h\text{-}fault}$	[1]2		
Streambed	K_{DRN}	3		
General head boundary	K_{GHB}	3		
		Vertical anisotropy		
	$K_{h/v\text{-}carb}$	5		
	$K_{h/v\text{-}base}$	78.5		26
	$K_{h/v\text{-}cover}$	15.8		56
	$K_{h/v\text{-}fault}$	2		
		Decay factor, ft^{-1}		
	λ_{carb}	7.6E-05		
	$\lambda_{crystal}$	1.5E-4		
		Recharge factor, percent		
Shallow slope	Rch_{low}	0.645		40
Moderate slope	Rch_{mod}	0.418		60
Steep slope	Rch_{steep}	0.125		73
		Standard error of weighted residuals, ft		
Water level		42.6		
Flow		49.9		

[1] Hydraulic conductivity perpendicular to fault specified as 0.02 ft/d.

Hydraulic Properties

Transmissivity was estimated from specific-capacity data from 22 wells drilled into Proterozoic basement crystalline rocks and Late Proterozoic-Cambrian cover crystalline rocks in Bedford County by using an equation from Todd and Mays (2005). Transmissivity values (T) ranged from 7 to 800 feet squared per day (ft^2/d), with a mean value of 73 ft^2/d. Corresponding values of horizontal hydraulic conductivity (K_h) were computed by dividing the T values by the aquifer-saturated thickness intercepted by each well. The K_h values ranged from 0.04 to 8.6 ft/d, with a mean value of 1.7 ft/d, and occupy the lower end of the range of mean values reported by Daniel and others (1997) for 1,153 wells in three topographic settings in fractured crystalline rocks in North Carolina (fig. 20).

The three hydrogeologic units presented earlier (Cambro-Ordovician metasedimentary rocks, Proterozoic basement crystalline rocks, and Late Proterozoic-Cambrian cover crystalline rocks) were each delineated as separate zones within the model domain and assumed to extend vertically through all model layers (figs. 18 and 21). Values of horizontal and vertical hydraulic conductivity (K_h and K_v) were estimated for each unit through model calibration. Horizontal hydraulic conductivity is assumed to be isotropic even though some borehole data indicate that the fracture networks within these units are aligned regionally along preferential directions (fig. 6). Available hydraulic conductivity data are insufficient, however, to characterize the differences in horizontal tensors. Hydraulic conductivity is assumed to be vertically anisotropic; values of vertical anisotropy (K_h/K_v) were estimated through model calibration. Assigned hydraulic conductivity values are intended only to characterize bulk aquifer properties at a regional scale. It is important to note that at the regional scale, data are also insufficient to define discrete fracture heterogeneities which may supply water at depth.

In addition, a power function similar to that applied in the groundwater-flow model of the Shenandoah Valley by Yager and others (2008) was used to relate the decrease in hydraulic conductivity K_h in each hydrogeologic unit below a threshold depth D below the bedrock surface (fig. 21):

$$K_{depth} = K10^{-\lambda d} \tag{4}$$

where

K_{depth} is the hydraulic conductivity [length/time] at depth D [length] below a threshold depth D, and

λ is a decay factor [length^{-1}].

The threshold depth for each hydrogeologic unit was set as the depth to the bottom of layer 1. Two depth-decay factors λ were estimated through model calibration—one for the Cambro-Ordovician metasedimentary rock unit and one for the combined Proterozoic basement crystalline-rock and Late Proterozoic-Cambrian cover crystalline-rock units.

The specified-thickness approximation (Sheets and others, 2015) was used to compute transmissivity in model layer 1, which is assumed to be unconfined. With this approximation model, layer 1 is treated as confined with a specified thickness to facilitate the numerical convergence of the groundwater-flow equation. The specified thickness was obtained by multiplying the model layer 1 thickness by a factor (*Satfactor*) to account for partial saturation:

$$Satfactor = \frac{head_{sim} - bottom}{thick}, \tag{5}$$

where

$head_{sim}$ is the head computed by a prior simulation,

$bottom$ is the elevation of the bottom of model layer 1, and

$thick$ is the thickness of model layer 1.

Satfactor was set to 0.05 × *thick* in areas where the simulated head was below the bottom of model layer 1. The value of *Satfactor* was computed iteratively through a sequence of simulations until the computed $head_{sim}$ approached stable values. Model layers 2 and 3 were also treated as confined, but the full layer thicknesses were used to compute the transmissivity of each layer. The specified thickness assumption for model layer 1 is appropriate for the steady-state simulation because the effect of pumping on the saturated thickness is negligible.

The location of the public-supply wells along mapped faults in Bedford County (City of Bedford, Well System Drilling Report, unpub. data, 1981) suggests the potential for hydraulic conductivity along faults to be larger than that of the surrounding fracture network, thereby channeling groundwater flow parallel to the fault. Alternatively, faults could be sealed, thereby creating low-permeability barriers to flow perpendicular to the fault. Faults cutting Blue Ridge and Piedmont rocks are extremely complex and not currently well understood. Because the faults have experienced multiple phases of deformation, structural markers indicating the direction of fault displacement are often variable within individual fault systems. Regional fault systems such as the Rockfish Valley Fault system in the northeastern portion of Bedford County contain both extensional and compressional (younger) features (Simpson and Kalaghan, 1989; Bailey and Simpson, 1993; Tollo and others, 2004), while the Bowens Creek Fault system to the southeast has a primarily dextral sense of shear with a younger compressional overprint (Edelman and others, 1987; Conley, 1989). The Blue Ridge and Peaks of Otter faults are thought to be primarily thrust related (Henika, 1981, 1997), while two unnamed faults in the central and northeastern portions of the county are shown to have a dextral sense of shear (Henika, 1997).

The hydraulic properties of faults are insensitive model parameters and could not be estimated through model calibration because the faults mapped at a 1:100,000 scale are widely spaced (Henika, 1997) and appear to have little influence on regional groundwater flow. The hydraulic effects of faults, however, are known to be important locally (Seaton and Burbey, 2005; White and Burbey, 2007), so hydraulic properties of faults were specified in the model to reflect larger horizontal and vertical hydraulic conductivities than those of the surrounding fracture network. Faults were also specified as horizontally anisotropic with the larger hydraulic conductivity value parallel to the fault and the smaller value perpendicular to the fault.

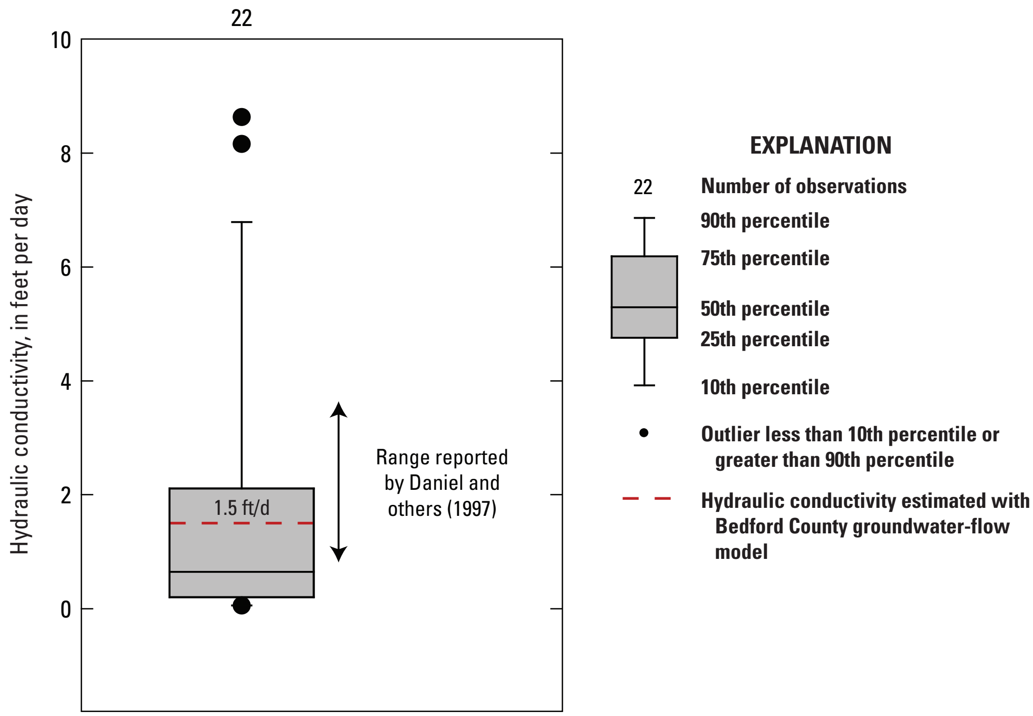


Figure 20. Distribution of hydraulic conductivity values computed from specific capacity of crystalline bedrock wells in Bedford County, Virginia.

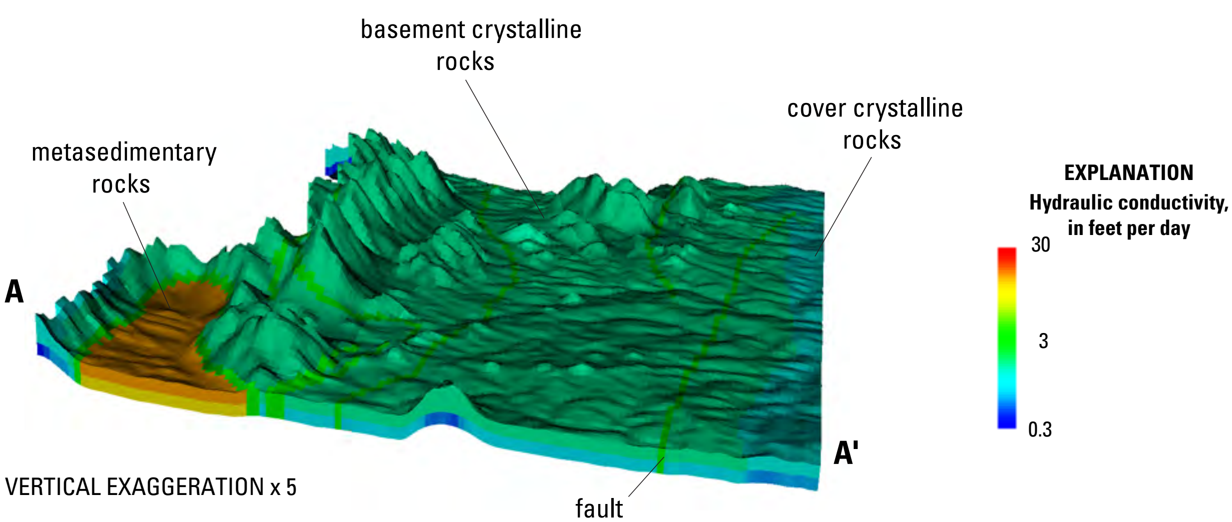


Figure 21. Model layers and hydraulic conductivity zones within model domain, Bedford County, Virginia. Location of section A'–A' shown in figure 18.

Model Calibration

The Bedford County groundwater-flow model was calibrated by adjusting hydraulic conductivity and recharge values to minimize the difference between simulated groundwater levels and discharges, and those observed in wells and at stream gages. Optimum values for model parameters were obtained through nonlinear regression by using UCODE 2005 (Poeter and others, 2005).

Observations

The nonlinear regression compared simulated water levels with the water levels measured in 47 bedrock wells in December 2010. The open intervals of many of these wells are not known (table 2), but drilled wells in the study area are typically cased below bedrock (table 4; fig. 7). For wells where the open interval was unknown, the assumed measurement depth was computed as midway between land surface and the bottom of layer 1. The measurement depths of two wells (33G 2 SOW 225 and 34H 5), located in areas where the simulated water table was below the bottom of layer 1, were specified as midway between the top and bottom of layer 2. The weights assigned to all water-level measurements were equivalent to a 3-ft standard deviation in measurement error.

Simulated groundwater discharges to drain boundaries representing Big Otter River and Goose Creek were compared with average base flows computed by PART from streamflow measurements obtained from four continuous record streamflow gages (02059485, 02059500, 02061000, and 02061500) during the period 2007 through 2012 (table 6). Each stream was divided into an upstream and downstream reach based on the locations of the four stream gages (fig. 18). The flow observations were assigned weights based on an arbitrary coefficient of variation in measurement error of 1 percent. Although the accuracy of the flow measurements is less than this value, the resulting weights in flow observations resulted in weighted residuals (observed minus simulated values) that were comparable in magnitude with weighted residuals in heads, so that both observation groups influenced the regression equally.

Model Fit

Residual plots for heads and flows (observed minus simulated values) indicate that the model simulates the groundwater system reasonably well. Residual plots for heads (fig. 22) show little bias in model error with the mean weighted residual near zero. The standard error (SE) of the weighted residuals for water level is 42.6 ft, which is less than 7 percent of the 685-ft range in head (table 7). Nearly all the simulated heads are within 100 ft of the measured value.

Simulated groundwater discharge to Big Otter River and Goose Creek closely match base flow estimated by PART for the period 2007 through 2012 with little bias (fig. 23).

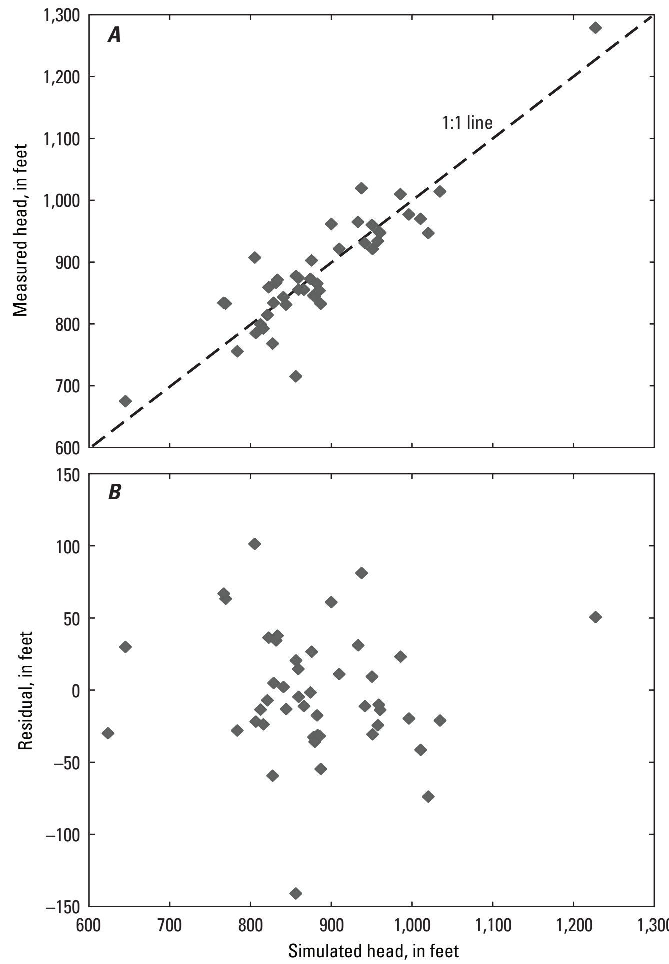


Figure 22. Residual plots for heads in Bedford County groundwater-flow model: (*A*) relation between simulated and observed values; (*B*) relation between simulated values and weighted residuals.

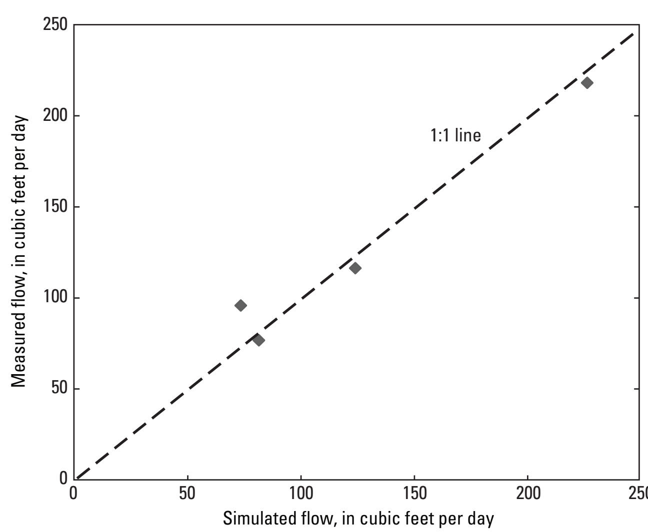


Figure 23. Residual plot for flows in Bedford County groundwater-flow model showing relation between simulated and measured values.

Simulated flows are within 10 percent of the measured flow at three of the stream gages. The largest error (23 percent) is for the upstream gage on Big Otter River. At a scale of hundreds of square miles, the error in head and flow is acceptable for an exercise in conceptual understanding; however, extrapolation of model results to smaller-scale domains would require additional hydrogeologic detail than is currently (2014) available.

Parameters

Fifteen parameter values were specified in the model, 8 of which were optimized through nonlinear regression (table 7). Coefficients of variation (cv) for the hydraulic conductivity parameters ranged from 26 to 56 percent. The cv for the recharge parameters ranged from 40 to 73 percent. These cv values indicate that the regression is relatively sensitive to most of the estimated parameters, but that the values are not particularly well estimated. Values of K for Proterozoic basement and Late Proterozoic-Cambrian cover crystalline rocks are similar (table 7), indicating little difference in permeability between these two units. Daniel and others (1997) reached a similar conclusion with respect to five hydrogeologic units delineated within crystalline rocks in North Carolina. The estimated depth-decay factor for crystalline rocks (5.5×10^{-4} ft^{-1}) resulted in an 80 percent reduction in K_h at a depth of 400 ft in model layer 3, which compares favorably with the range of 60 to 90 percent reduction in K_h at a depth of 300 ft estimated by Daniel and others (1997) for crystalline bedrock wells in North Carolina.

Estimated recharge factors ranged from about 42 percent of available water (PR–ET) for moderate to steep slopes, to 64 percent of available water for shallow slopes. Recharge values applied to model cells ranged from 4 in/yr near streams and in the southeastern corner of the model domain, to 28 in/yr in isolated flat areas in uplands near the northwestern corner (fig. 24). The mean recharge value within the model domain was 9.4 in/yr, which is similar to the mean long-term base flow computed using hydrograph separation for Goose Creek (8.3 in/yr at continuous station 02059500) and Big Otter River (9.3 in/yr at continuous station 02061500). The spatial distribution of recharge reflects the grid resolution of the PRISM data used to estimate precipitation within the model domain (fig. 24).

Model Application

The hydraulic head distribution computed by using the Bedford County model indicates that groundwater flows towards the major drainages within the model domain and that the water table reflects the pattern of the stream network (fig. 25). The highest water levels are in the uplands along watershed divides and along the northwest model boundary. The distribution of head residuals indicates no spatial bias in the model (fig. 25).

The water budget computed by the Bedford County model indicates that nearly all the recharge within each major watershed discharges to the drainage network within that watershed (table 8). The distribution of recharge and discharge is generally proportionate to the drainage area of the four major basins, although the Goose Creek and James River Basins derive about 7 percent of their discharge as underflow from adjacent watersheds. The Big Otter River receives 40.8 percent of the total daily groundwater outflow from fractured-rock aquifers in Bedford County; Goose Creek receives 25.8 percent, the James River receives 18.2 percent, and Smith Mountain and Leesville Lakes receive 15.2 percent (fig. 26). The volume of public-supply pumping within the model domain (not indicated in table 8) is a small part of the water budget and represents 0.2 percent of the total groundwater outflow.

Model Limitations

The Bedford County groundwater-flow model was constructed to represent recent (December 2010) steady-state conditions when recharge and water levels are near their long-term averages, and not extreme conditions, such as drought. The model could be modified to simulate transient conditions with varying recharge, but this would necessitate the specification of values for specific yield in the model. Continuous water-level monitoring for periods of 3 or more years at multiple wells could supply the required information to estimate this model parameter. Simulation of transient conditions could have the additional benefit of supporting an estimate of the volume of groundwater stored within the model domain.

The volume of pumping within the model domain is small and exerts little stress on the aquifer system. The values of hydraulic parameters estimated through model calibration are based on limited data, including base-flow volumes and water-level observations at 47 wells. The uncertainty in the parameter values is reflected by the relatively large coefficients of variation associated with the estimates. In addition, although the hydraulic conductivity of the fracture network varies over 4 orders of magnitude, uniform values of K_h are specified within the model. As a result, computed water levels likely will differ from those measured locally in wells, as indicated by the standard error (45.6 ft) and scatter in the head residuals (fig. 22). Other approaches could include anisotropy in K_h to account for the preferential alignment of fractures noted in borehole logs (fig. 6). Finally, computed water levels are likely underpredicted in upland areas as a result of the specified-thickness approximation for model layer 1. More accurate results could be obtained by treating model layers 1 and 2 as convertible, that is, either confined or unconfined. This representation would require the application of the Newton solver (Niswonger and others, 2011), however, which does not support the depiction of directional permeability along faults that was used in the current model. Despite these limitations, the model provides accurate estimates of the rate and direction of regional groundwater flow and serves as a solid basis for future modifications to simulate the aquifer system in greater detail.

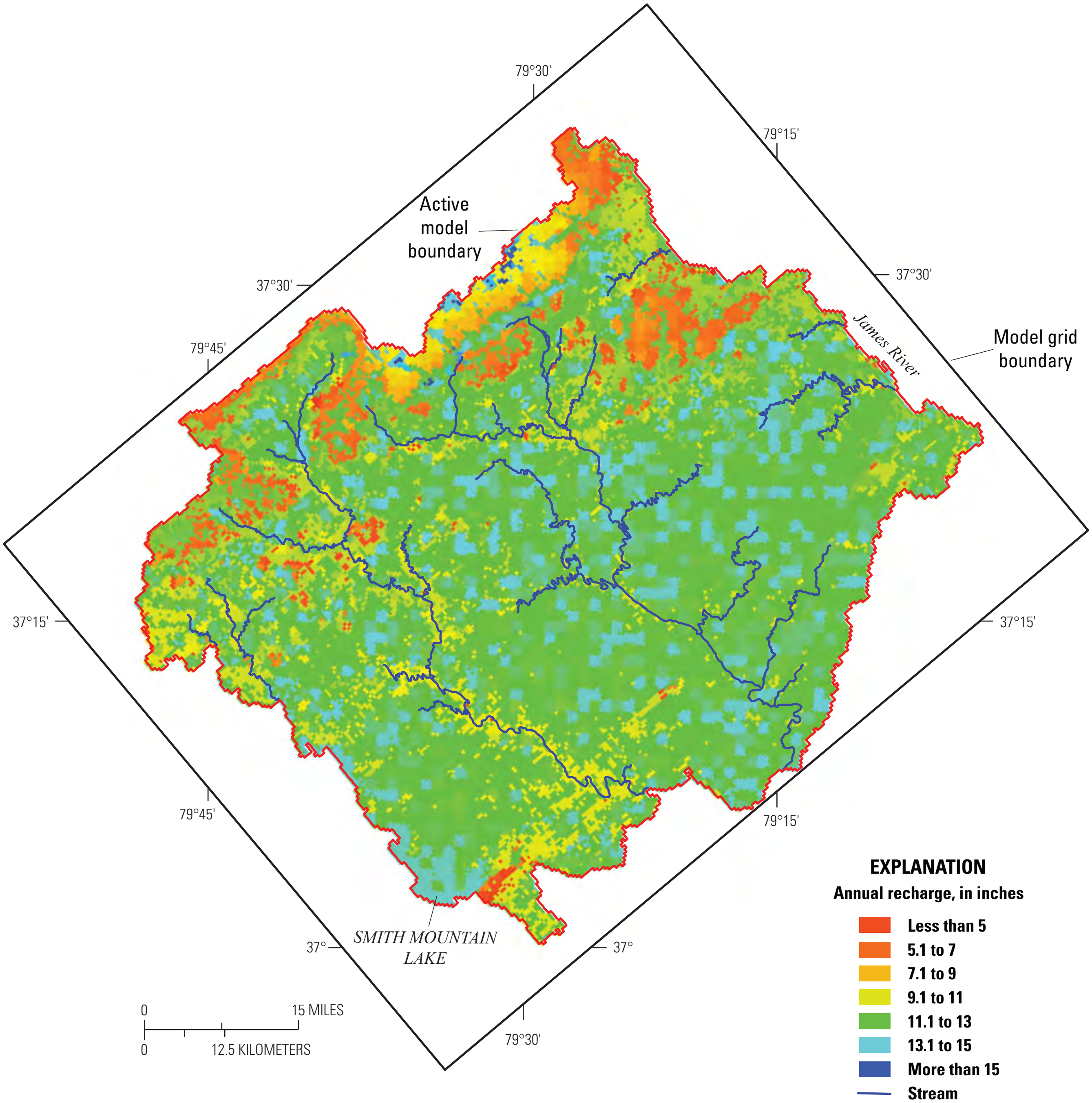


Figure 24. Distribution of annual recharge for steady-state conditions in the Bedford County groundwater-flow model.

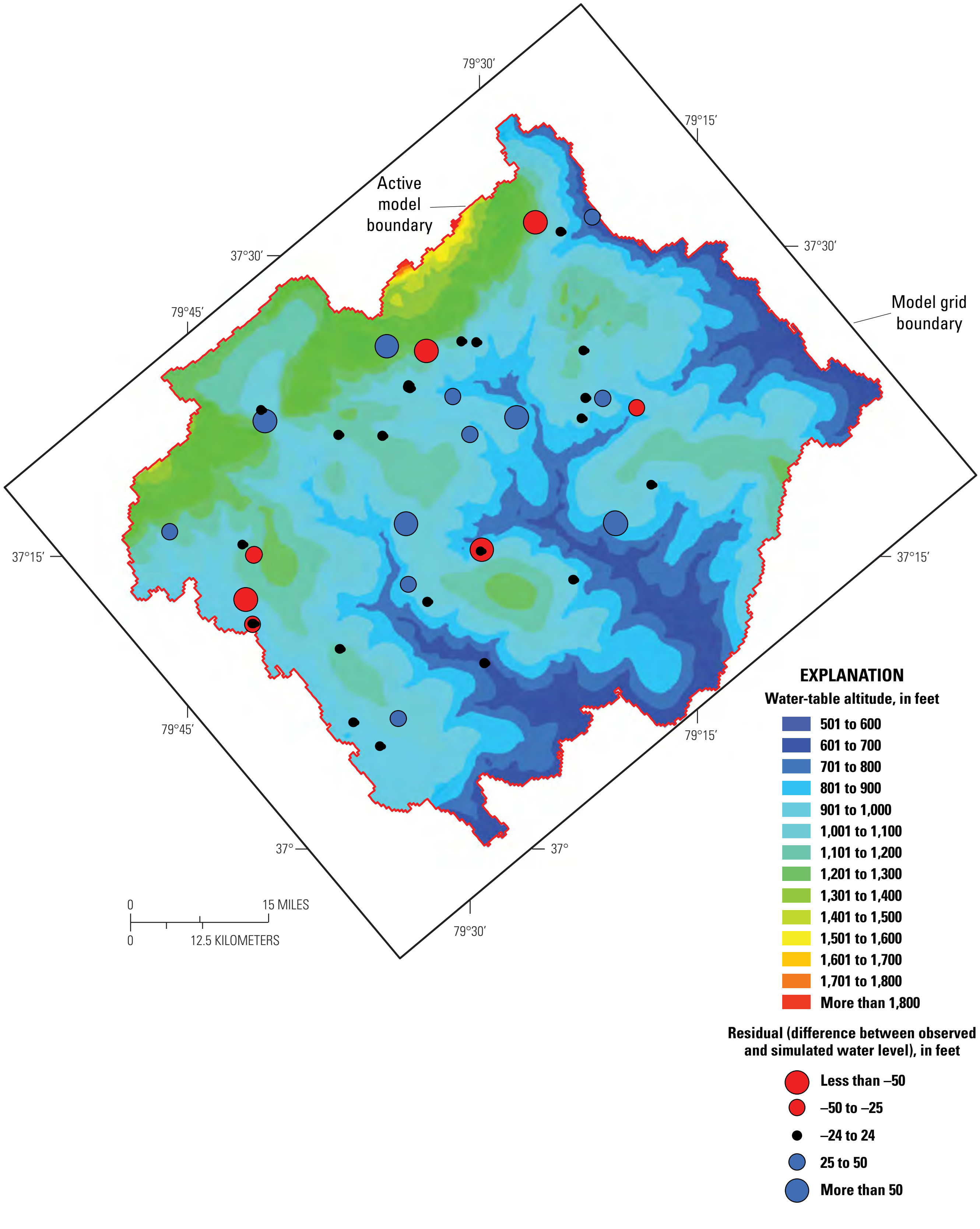
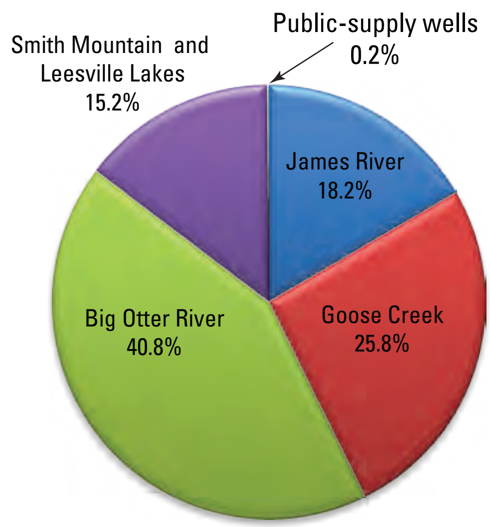


Figure 25. Water table simulated for steady-state conditions by the Bedford County groundwater-flow model.

Table 8. Simulated water budget for Bedford County groundwater-flow model.

[ft^3/d, cubic foot per day]

Source	Inflow volume (10^5 ft^3/d)	Percent	Discharge	Outflow volume (10^5 ft^3/d)	Percent
Recharge			**Base flow**		
James River	9.8	16.9	James River	10.5	18.2
Big Otter River	23.5	40.5	Big Otter River	23.5	40.8
Goose Creek	14.1	24.3	Goose Creek	14.9	25.8
Smith Mountain & Leesville Lakes	10.6	18.3	Smith Mountain & Leesville Lakes	8.8	15.2
Total	58.0	100.0		57.7	100.0



Note: Numbers may not add up to 100 percent because of independent rounding.

Figure 26. Proportion of groundwater flow through major basins, Bedford County, Virginia.

Summary and Conclusions

Groundwater resources in Bedford County, Virginia, include water that is stored in and flows through Piedmont and Blue Ridge regolith and fractured-rock aquifers. The fractured-rock aquifers in Bedford County were subdivided into three hydrogeologic units based on rock type: Cambro-Ordovician metasedimentary rocks, Proterozoic basement crystalline rocks, and Late Proterozoic-Cambrian cover crystalline rocks. Cambro-Ordovician metasedimentary rocks in the northwest part of the county have the highest driller-reported well yields, largest depth to bedrock, and highest model calibrated hydraulic conductivity of the three hydrogeologic units. Mean reported drillers' well yields and model-calibrated hydraulic conductivity in Proterozoic basement and Late Proterozoic-Cambrian cover crystalline-rock units were similar. Well characteristics and model results indicate little difference between the crystalline-rock units in Bedford County, a conclusion that is supported by findings in geologically similar areas of North Carolina. Reported seasonally dry wells or low-yielding wells in the southeastern portion of the county cannot be explained by rock type but likely reflect local heterogeneities that are beyond the scope of this study.

Groundwater levels for the period of study (2009–2013) were found to vary only a few feet in response to seasonal changes in climate conditions. Because the water level mimics land surface, groundwater flow is topographically driven. Groundwater flows from upland areas to local streams and rivers such that surface-water divides are coincident with groundwater divides.

A water budget for the county shows model-calibrated recharge rates that mimic precipitation patterns and range from 4 inches per year (in/yr) near streams to 28 in/yr in upland areas of the Blue Ridge Physiographic Province. The mean rate of recharge from model results is 9.4 in/yr, a value equivalent to the long-term mean from hydrograph separation of streamflow in the Big Otter River. An expanding rural population has resulted in an increase in the groundwater use in Bedford County from 2.80 million gallons per day (Mgal/d) to 4.85 Mgal/d between 1985 and 2005. Evapotranspiration rates ranged from 30.5 to 31.4 in/yr based on streamflow and are within the higher end of the 23.7- to 31.5-in/yr range of values from national datasets. The percent volume of groundwater discharging from Piedmont and Blue Ridge fractured-rock aquifers of Bedford County is proportional to the drainage area of the four major basins: Big Otter River (40.8 percent), Goose Creek (25.8 percent), James River (18.2 percent), and Smith Mountain and Leesville Lakes (15.2 percent).

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Appendix 1. Borehole Geophysical Logs

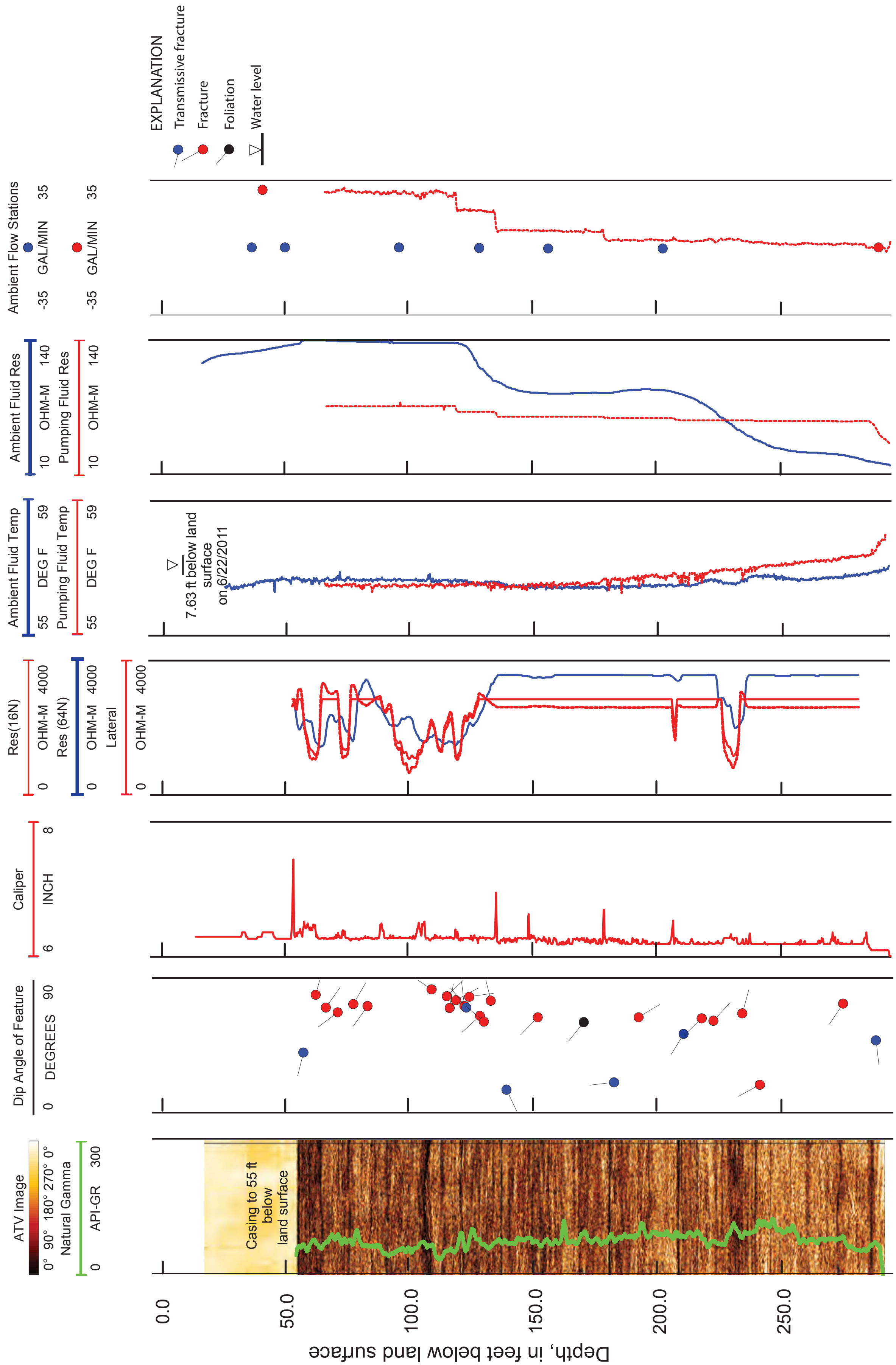

Figure 1–1. Borehole geophysical logs from well 33G 7 in Bedford County, Virginia.

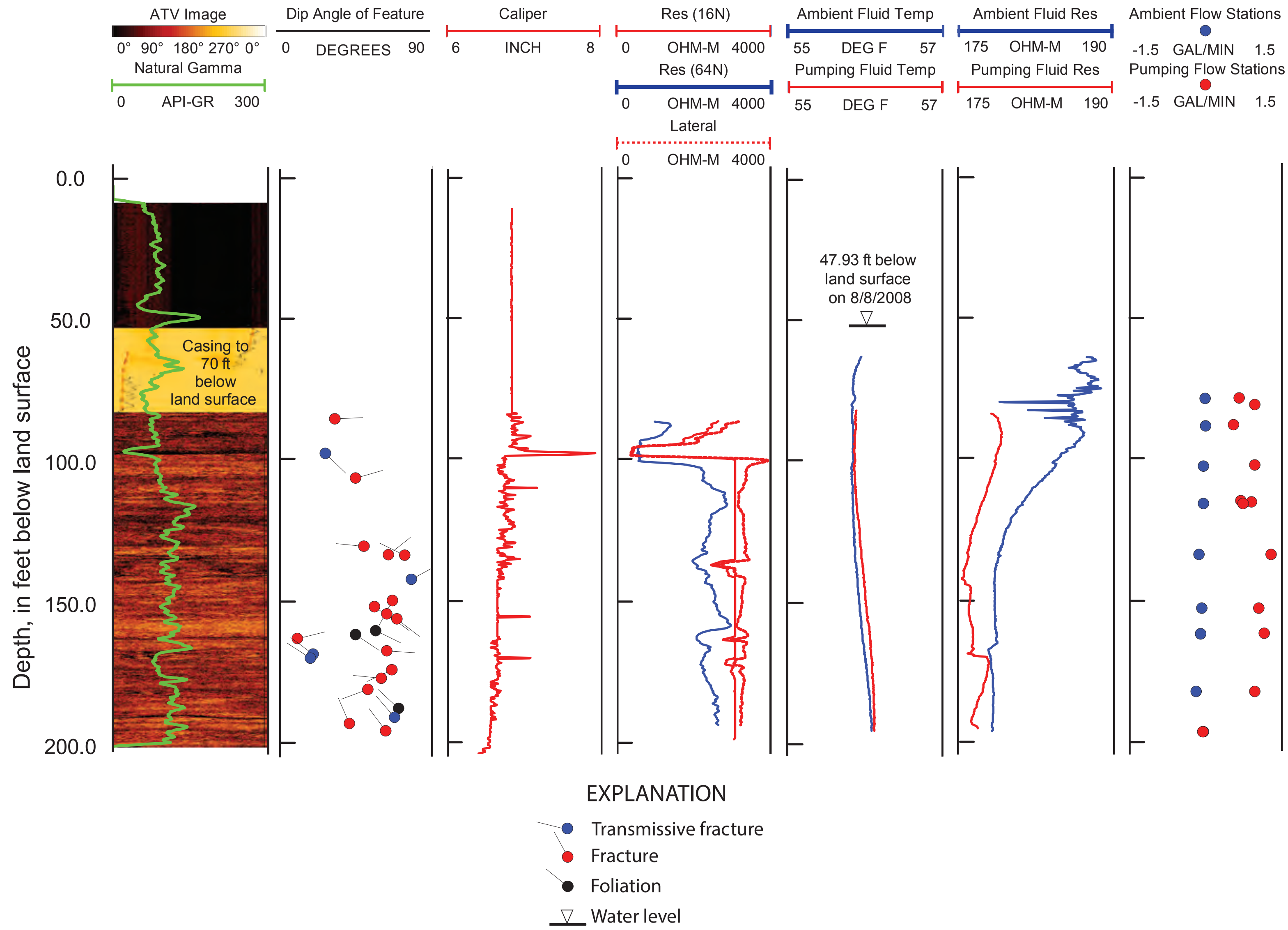

Figure 1–2. Borehole geophsyical logs from well 33G 2 SOW 225 in Bedford County, Virginia.

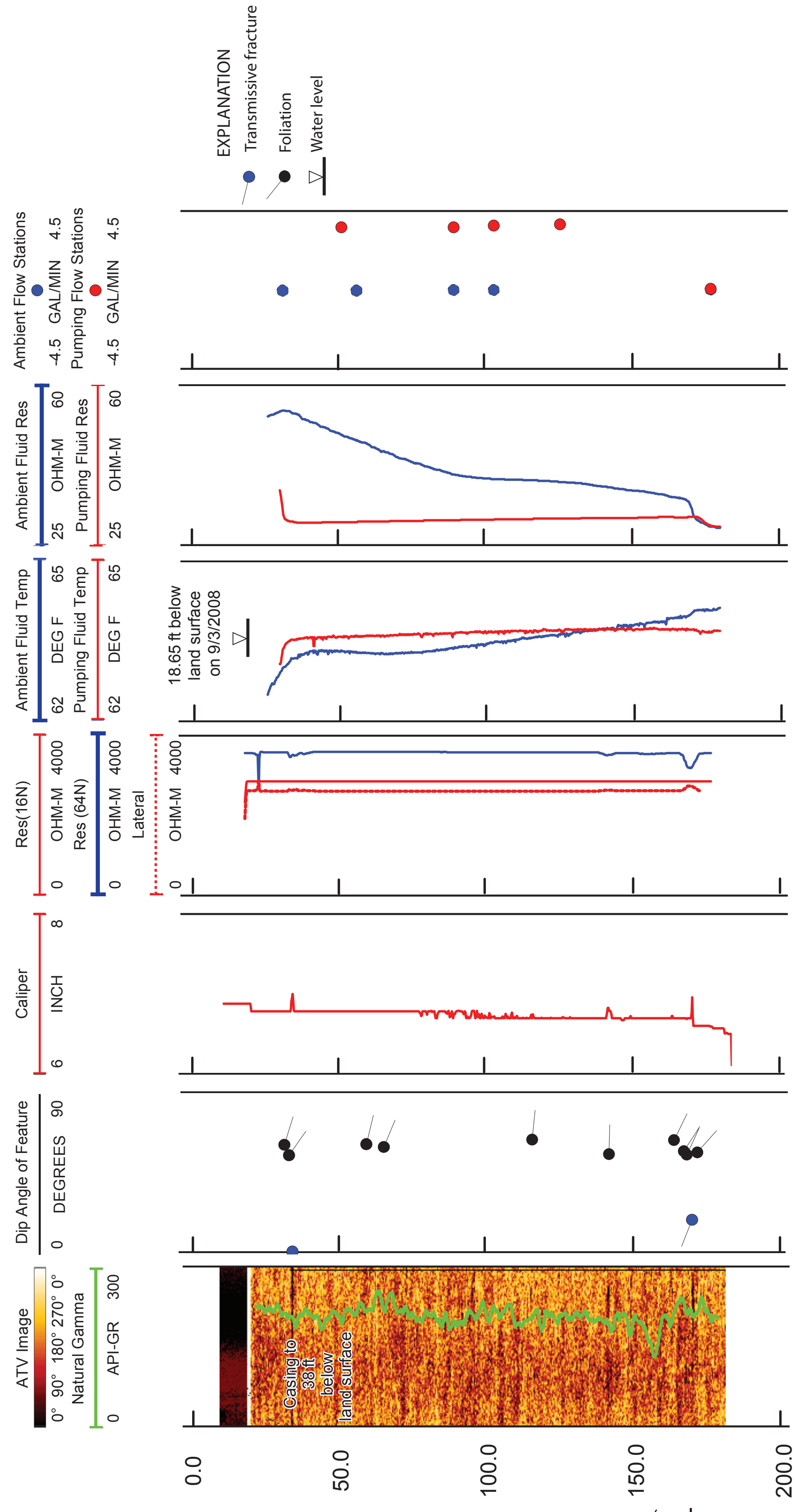

Figure 1–3. Borehole geophsyical logs from well 35H 1 SOW 226 in Bedford County, Virginia.

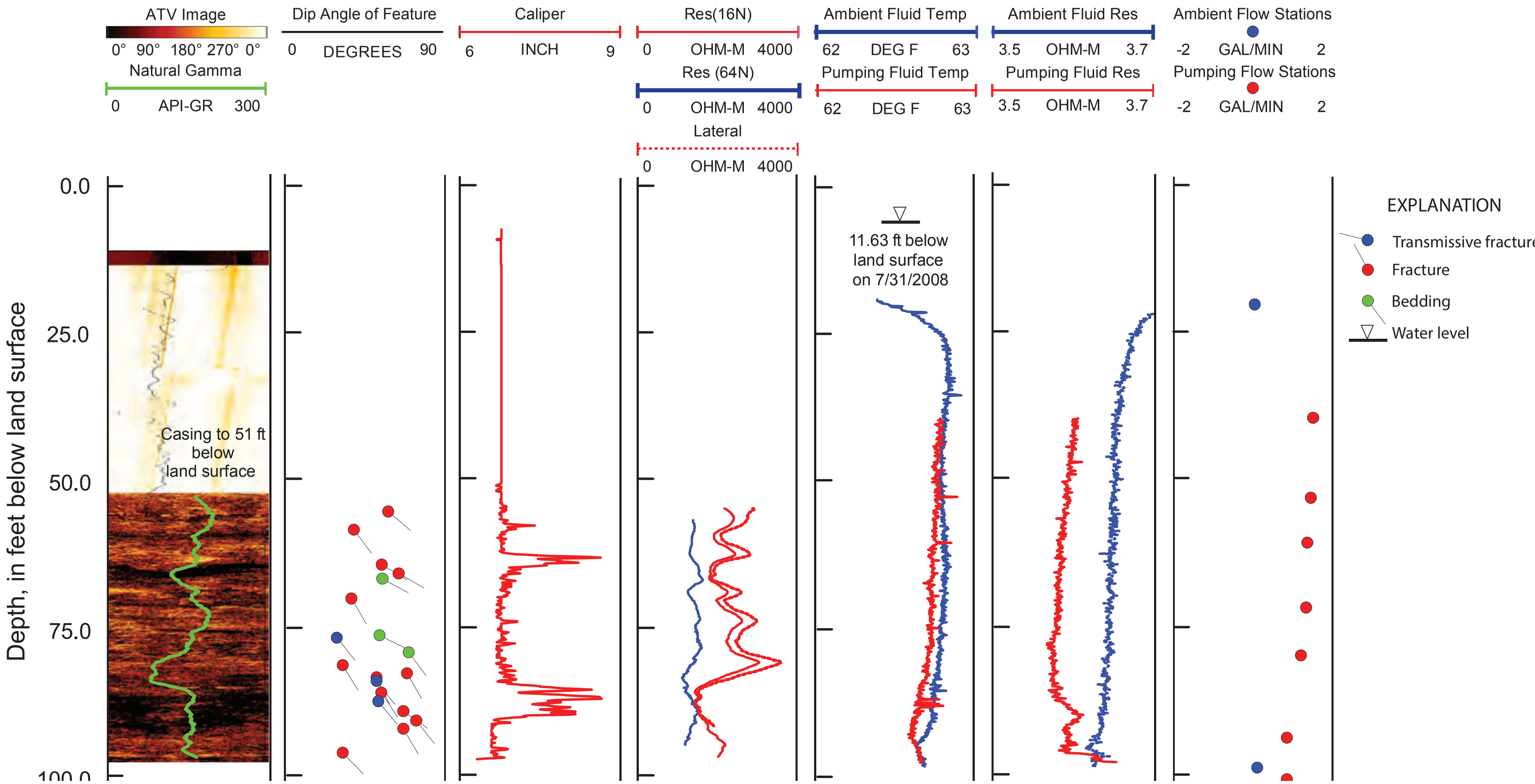

Figure 1–4. Borehole geophsyical logs from well 33G 1 SOW 224 in Bedford County, Virginia.

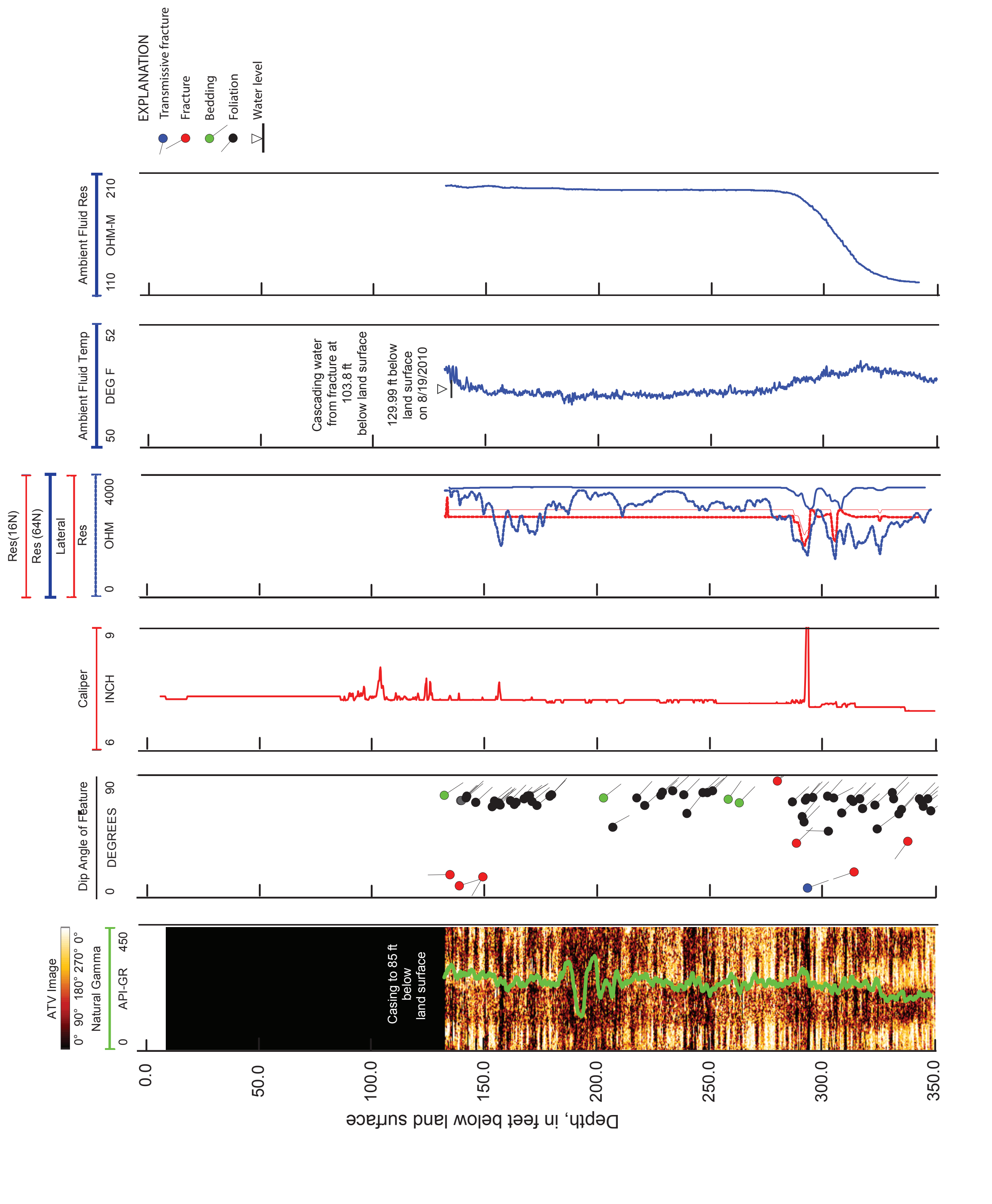

Figure 1–5. Borehole geophsyical logs from well 34E 1 in Bedford County, Virginia.

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