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Groundwater Resources Program

Hydrologic Budget and Conditions of Permian, Pennsylvanian, and Mississippian Aquifers in the Appalachian Plateaus Physiographic Province

Scientific Investigations Report 2015–5106
Version 1.1, October 2015

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

Front Cover. Big South Fork National River and Recreation Area, Kentucky/Tennessee. Photograph from National Park Service.
Back cover. Map of principal aquifers of the United States showing the Appalachian Plateaus study area.

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SALLY JEWELL, Secretary

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Conversion Factors

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square foot (ft^2)	929.0	square centimeter (cm^2)
square foot (ft^2)	0.09290	square meter (m^2)
square mile (mi^2)	259.0	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
million gallons (Mgal)	3,785	cubic meter (m^3)
	Flow rate	
foot per day (ft/d)	0.3048	meter per day (m/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per square mile [(ft^3/s)/mi^2]	0.01093	cubic meter per second per square kilometer [(m^3/s)/km^2]
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
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)
	Mass	
ton, short (2,000 lb)	0.9072	megagram (Mg)

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

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.

Abbreviations

AMO	Atlantic Multidecadal Oscillation
AWC	available water capacity
BFI	base-flow index
DEM	digital elevation model
ET	evapotranspiration
ETo	reference evapotranspiration
LOWESS	locally weighted scatterplot smoothing
LST	land surface temperature
NAO	North Atlantic Oscillation
NRCS	Natural Resources Conservation Service
PDSI	Palmer Drought Severity Index
PET	potential evapotranspiration
SSEB	simplified surface energy balance
SSEBop	Operational Simplified Surface Energy Balance (model)
SSURGO	Soil Surface Geographic Database
SWB	Soil-Water-Balance (model)
USGS	U.S. Geological Survey

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Abstract

In response to challenges to groundwater availability posed by historic land-use practices, expanding development of hydrocarbon resources, and drought, the U.S. Geological Survey Groundwater Resources Program began a regional assessment of the Appalachian Plateaus aquifers in 2013 that incorporated a hydrologic landscape approach to estimate all components of the hydrologic system: surface runoff, base flow from groundwater, and interaction with atmospheric water (precipitation and evapotranspiration). This assessment was intended to complement other Federal and State investigations and provide foundational groundwater-related datasets in the Appalachian Plateaus.

A regional Soil-Water-Balance model was constructed for a 160,000-square-mile study area that extended to the topographic divide of all streams originating outside but flowing into areas underlain by Appalachian Plateaus aquifers. The model incorporated soil, landscape, and climate variables to estimate an annual water budget for the 32-year period from 1980 to 2011 and was calibrated using base-flow data estimated by hydrograph separation techniques from 20 streamflow gaging stations across the study area. Over this period, an average of 47 inches per year (in/yr) of precipitation fell on Appalachian Plateaus aquifers. Simulations from the regional Soil-Water-Balance model indicate that only 19 percent of the precipitation or an average 9 in/yr recharged aquifers, and 19 percent resulted in surface runoff to streams. The remaining 62 percent, an average of 27 in/yr of water, was returned to the atmosphere via evapotranspiration. Because withdrawals from aquifers due to pumping equated to less than 1 percent of the water budget, differences in predevelopment and postdevelopment regional water budgets of the Appalachian Plateaus were minimal. Storage changes caused by filling of abandoned coal-mine aquifers and long-term differences in aquifer storage resulting from climate fluctuations constitute a small portion of the overall water budget.

The percentage of precipitation that results in recharge, runoff, or evapotranspiration from the landscape varies annually by up to a factor of two depending on temporal changes in prevailing climate conditions and spatial changes in basin characteristics, precipitation patterns, and sources of atmospheric moisture over a large study area. A comparison of water-budget estimates from the regional Soil-Water-Balance model for a dry year (1988) and wet year (2004) showed that evapotranspiration accounts for most of the annual differences in precipitation. As a portion of annual precipitation, evapotranspiration ranged from 69 percent (dry year) to 52 percent (wet year), a range four times greater than the 15 percent (dry year) to 18 percent (wet year) range estimated for recharge. Evapotranspiration as a percentage of precipitation peaks during dry periods, whereas base flow and runoff tend to reach minimum values. During wet periods, this relationship is reversed and base flow and runoff as a percentage of precipitation generally peak while evapotranspiration percentages reach minimum values. Annual recharge in the Appalachian Plateaus reaches a maximum at near 20 percent of annual precipitation, regardless of the severity of wet conditions.

Hydrograph separation data from 849 streamflow gaging stations in the study area were used to assess trends in streamflow, base flow, surface runoff, and base-flow index, or ratio of base flow to streamflow, in the Appalachian Plateaus for the period from 1930 to 2011. Annual data anomalies for each of the four variables were individually defined as the annual standard deviation from the mean at all 849 streamflow gaging stations. Annual data anomalies confirm the close relation of annual precipitation to both base flow and runoff components of streamflow, and both components increased during the period of analysis. Around 1970, conditions shifted streamflow from values generally below to above long-term means. At a regional scale, increases in base flow account for most of these observed increases in mean annual streamflow. The independence of the base-flow index to annual climate trends indicate that changes in the components of streamflow of the Appalachian Plateaus are probably in response to shifts in seasonal precipitation or widespread land-use practices.

A subset of 77 index streamgages, defined as having 60 or more years of complete record between the years 1930 and 2011 with no more than 20 percent missing data, was selected to show spatial patterns of change in the water budget. Data from the index streamgages showed that the overall trends in base flow are dependent upon the period of evaluation. Long-term (1930–2011) increases in base flow were observed throughout the study area. For two shorter periods (1930–1969 and 1970–2011) trends in base flow were largely negative. In general, spatial patterns of change in streamflow, base flow, and runoff were mixed but generally consistent with prevailing climate patterns and land-use changes.

Introduction

Groundwater is essential for domestic supplies, development of energy resources, and sustaining aquatic ecosystems in the Permian, Pennsylvanian, and Mississippian fractured-rock aquifers of the Appalachian Plateaus Region of the eastern United States. The Appalachian Plateaus region occupies approximately 86,000 square miles (mi^2) in portions of Alabama, Georgia, Kentucky, Maryland, Ohio, Pennsylvania, Tennessee, Virginia, and West Virginia and is home to approximately 13.7 million residents (fig. 1) (Kenny and others, 2009). Aquifers in the Appalachian Plateaus consist of alternating sequences of fractured sandstone, siltstone, shale, limestone, and coal in Permian-, Pennsylvanian-, and Mississippian-age rock formations. Erosion of the rock formations has produced the characteristic steep hills, rugged topography, and deeply incised valleys. The region is home to approximately 2.4 million rural residents that withdraw a total of 163 million gallons per day (Mgal/d) of potable water from domestic wells and 3.6 million residents that receive 286 Mgal/d of groundwater from public supplies (Kenny and others, 2009). The remaining 7.7 million residents rely on 1,500 Mgal/d of surface water for potable water supply, which in many cases depends on groundwater contributions as base flow to streams.

Groundwater withdrawals from Appalachian Plateaus aquifers are small compared to the billions of gallons withdrawn daily from more prolific aquifers of the United States (Reilly and others, 2008). As a result, few studies have focused on groundwater availability in Appalachian Plateaus aquifers at regional scales consistent with (1) Appalachian Basin energy-resource assessments (for example, Kirschbaum and others, 2012), or (2) State and Federal regulatory frameworks. As a vital resource of the Appalachian Plateaus, supplies of good-quality groundwater require adequate characterization to support informed water-management planning. The availability of this vital resource is critical to the development of other natural resources in the region, such as coal, oil, natural gas, iron ore, limestone, dolomite, and timber. Base flow from groundwater also sustains aquatic resources, including Appalachian streams notable for their highly diverse ecosystems (Amey, 2011; Constantz, 2004; Weidensaul, 2000).

The primary concern limiting the availability of groundwater resources in Appalachian Plateaus has been, and will continue to be, the quality of water that for decades has been affected by several factors, including coal mining (fig. 2), drilling for oil and gas, and industrial and agricultural practices (Appel, 1985). Rapid development of natural-gas resources in the Appalachian Plateaus since 2005 represents an important recent addition to the Nation's energy portfolio (fig. 3). Energy resource extraction, however, presents a conflict between economic development in the region, the Nation's energy demands, and potential environmental risks to water resources of the Appalachian Plateaus supporting human and ecosystem needs (Vidic and others, 2014; Ingraffea and others, 2014).

Sustainable water supplies may also be limited by periods of drought. Climate-model scenarios suggest seasonal recharge in winter and drawdown of groundwater in summer will shift to earlier months of the year (Neff and others, 2000). This is especially important because surface-water resources are stressed during periods of low flow. Surface-water resources in the Appalachian Plateaus are highly dependent on base-flow groundwater discharge, which accounts for 50 to 65 percent of annual streamflow (Zurawski, 1978; Bloyd, 1974). Water levels are generally controlled by topography (Gleeson and others, 2011), and the annual magnitude of groundwater discharge as base flow to streams is highly correlated with mean annual precipitation (Risser and others, 2008). Because water levels rise and fall in response to prevailing climate conditions (fig. 4) (Bolton and others, 2009; Buckwalter and Moore, 2007; Kipp and Dinger, 1991) and the fractured rocks of the region have relatively limited capacity to store water (Mooty, 1990), water levels may decline rapidly during extended periods of drought. Declining water levels can result in local water shortages for users dependent on small streams that may dry as the source of stream base flow falls to critical levels (Bolton and others, 2009; Bettandorff and Sholar, 1985).

In 2013, the U.S. Geological Survey (USGS) Groundwater Resources Program began a regional assessment of groundwater availability in the Appalachian Plateaus aquifers as part of an ongoing assessment of principal aquifers across the Nation. These regional assessments are conducted to gain a better understanding of the status of groundwater resources and how changes in land use, water use, and climate may affect those resources. The primary goal of these regional assessments is to improve our ability to forecast water availability for future economic and environmental uses (Reilly and others, 2008). The availability and sustainability of groundwater resources in the Appalachian Plateaus can be categorized by two broad themes: (1) existing and potential for future water-quality degradation on a regional scale, and (2) localized water shortages during periods of extended drought. The Appalachian Plateaus Groundwater Availability study is intended to provide the foundational groundwater-related datasets for other Federal and State water-resource investigations to assess these broad themes within the context of drinking-water resources, aquatic ecosystems, and continued energy resource development in the region. An improved understanding of groundwater availability in the Appalachian Plateaus thus plays a central role in sustained economic development of the region.

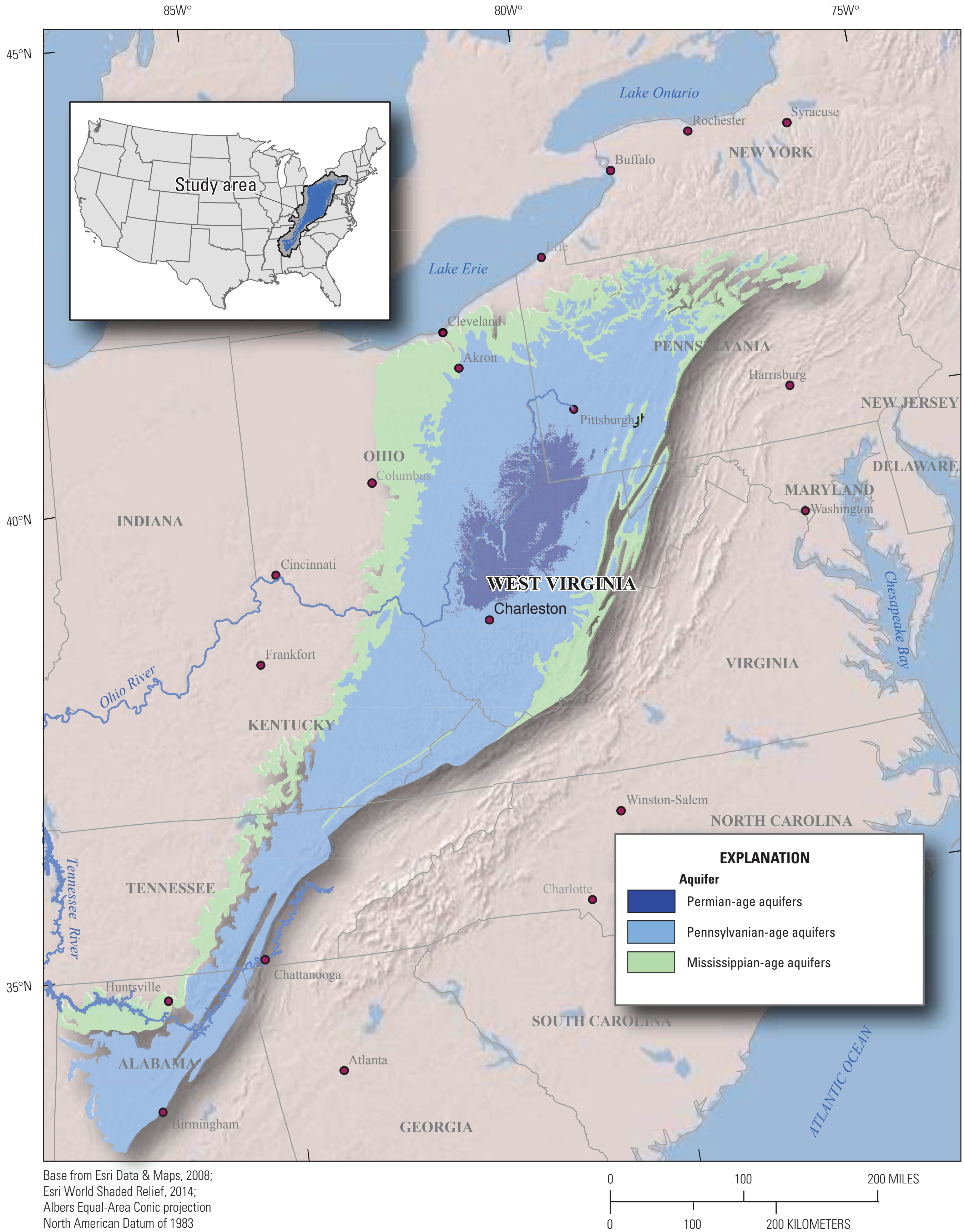


Figure 1. Location of Permian-, Pennsylvanian-, and Mississippian-age aquifers of the Appalachian Plateaus, eastern United States.

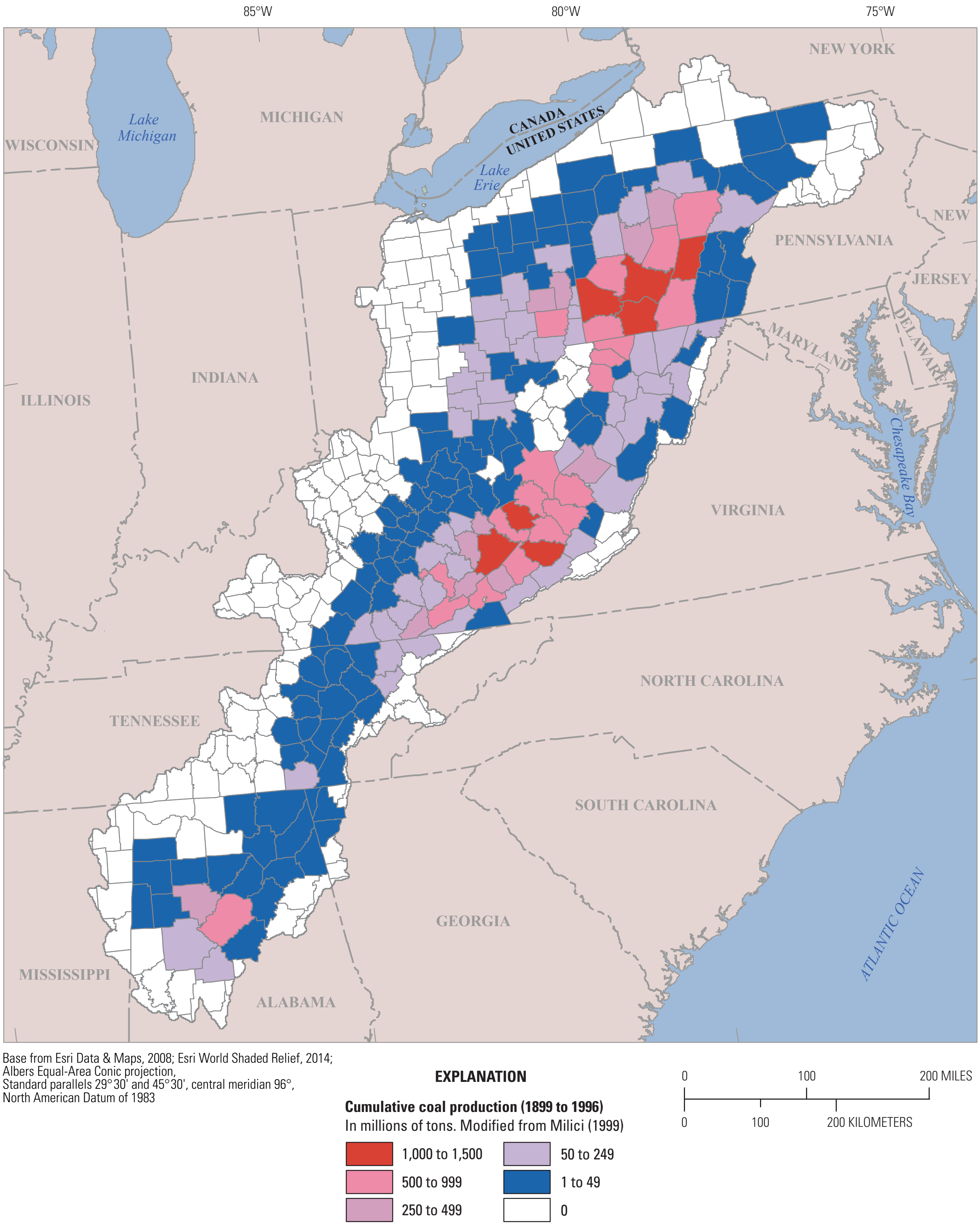


Figure 2. Cumulative bituminous coal production in the Appalachian Basin from 1899 to 1996 (modified from Milici, 1999).

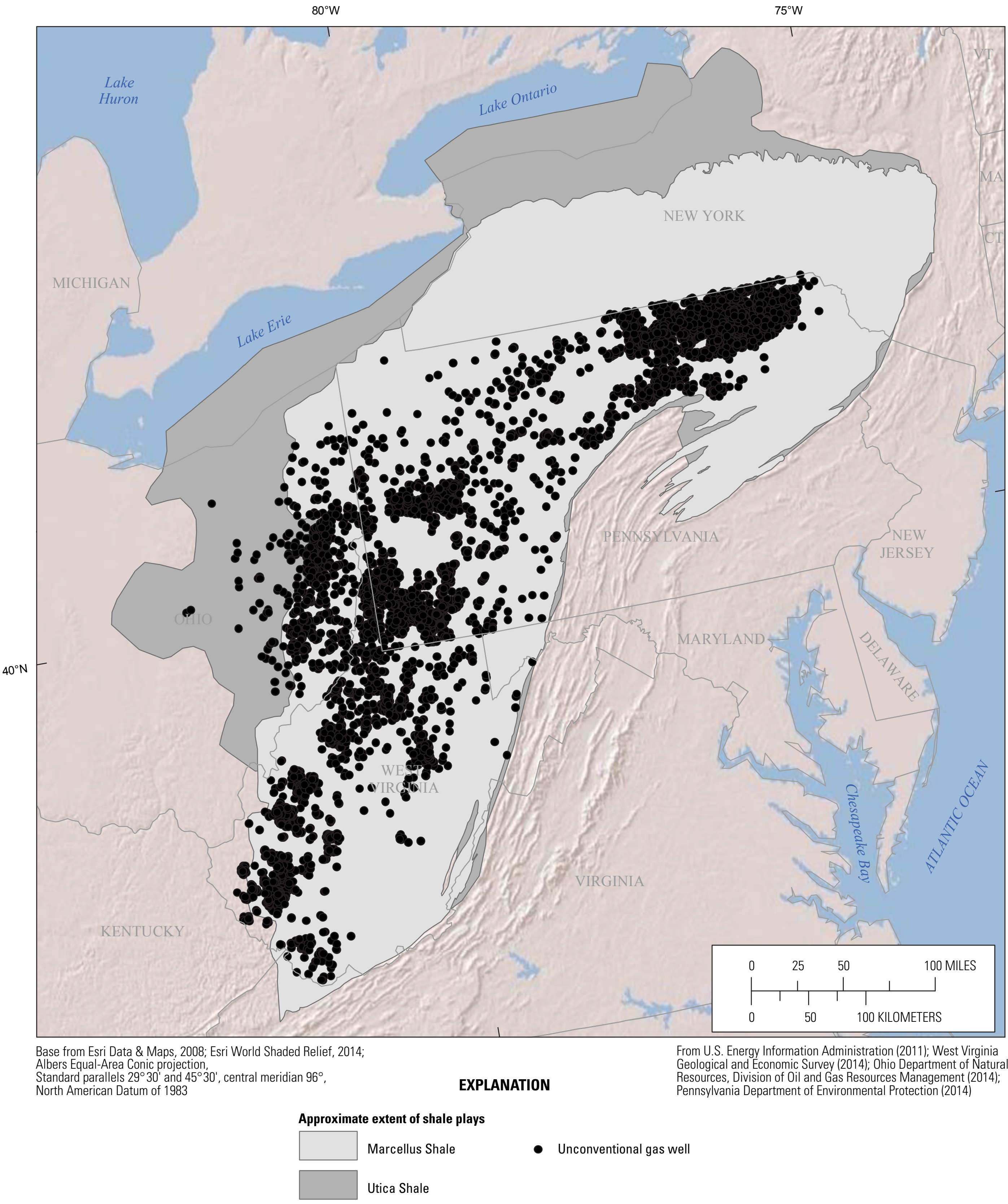


Figure 3. Distribution of unconventional gas wells completed in the Marcellus and Utica shales from 2005 to 2013.

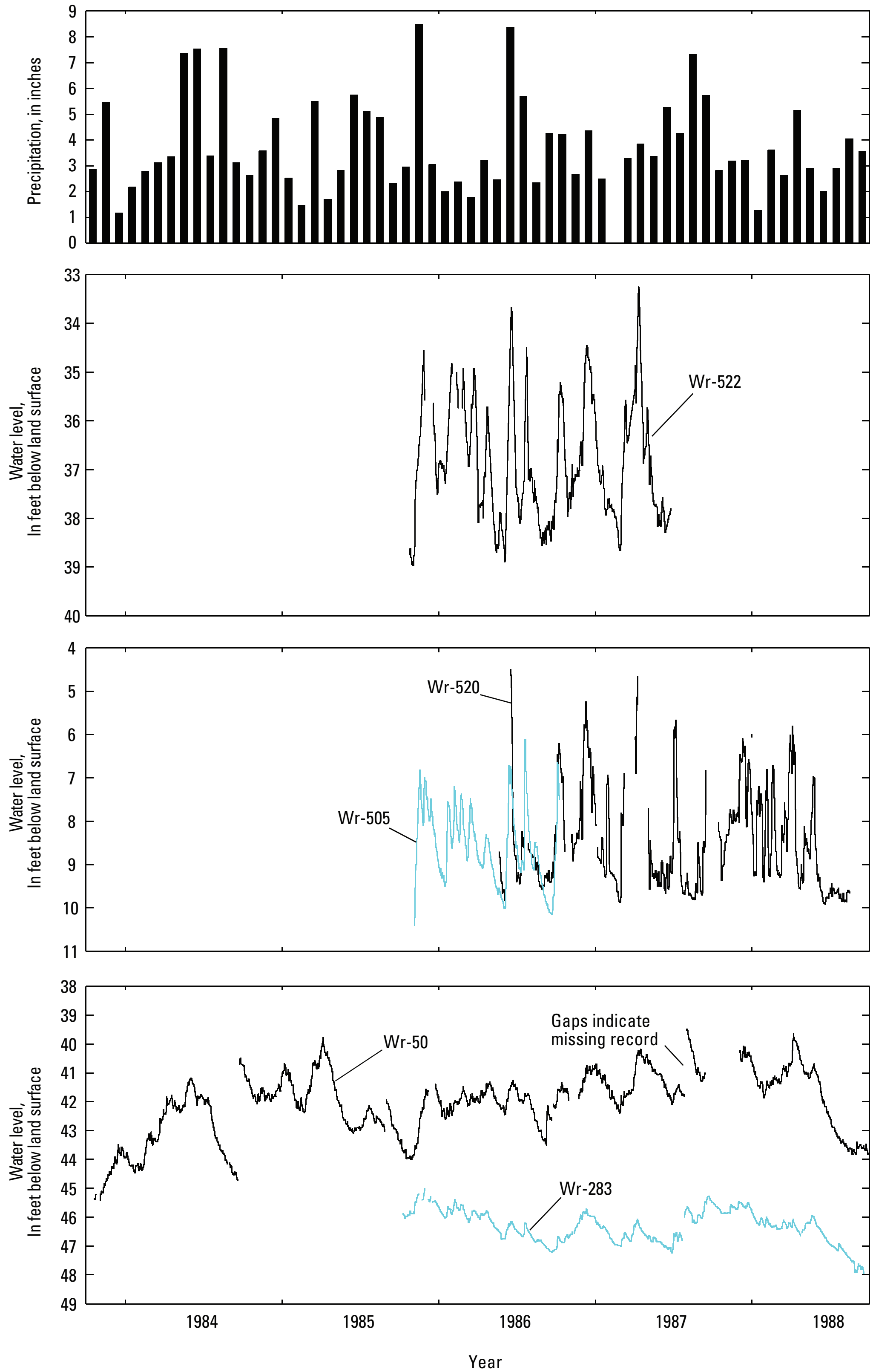


Figure 4. Monthly precipitation and water-level fluctuations in bedrock hillside wells Wr-50 and Wr-283 and valley bottom wells Wr-505, Wr-520, and Wr-522, Warren County, Pennsylvania (modified from Buckwalter and Moore, 2007).

Purpose and Scope

The purpose of this report is to describe the inflows, outflows, and system-wide water-budget changes that can be used to assess groundwater availability in Permian, Pennsylvanian, and Mississippian aquifers of the Appalachian Plateaus. Specific objectives of this report on the hydrologic budget of the Appalachian Plateaus are to: (1) present a mean annual water budget for the period from 1980 to 2011, (2) evaluate previous regionwide estimates of various water-budget components, (3) assess the range of hydrologic conditions observed during periods of wet and dry climate conditions, and (4) define the spatial and temporal variability in long-term base-flow discharge to streams. Understanding the water-budget trends are useful for defining potential water-limiting conditions associated with long-term climate change, seasonal droughts, changes in land- and water-use patterns, expanding urban populations, and various aspects of continued energy development. Two key components of this report are to document the Soil-Water-Balance (SWB) model (Westenbroek and others, 2010) constructed for the Appalachian Plateaus and surrounding areas, and to summarize the annual base-flow separation of data that were collected from 849 USGS streamflow gaging stations and span the period from 1900 to 2011 (Nelms and others, 2015).

Hydrogeologic Setting

Appalachian Plateaus aquifers underlie portions of the Appalachian Plateaus Physiographic Province as defined by Fenneman and Johnson (1946) and include the physiographic sections of Southern New York, Kanawha, Allegheny Mountain, Cumberland Plateau, and Cumberland Mountain, collectively henceforth referred to as the Plateaus (fig. 5). At the surface, Plateaus aquifers cap a broad 200-mile (mi)-wide, highly dissected landscape in New York, Ohio, Pennsylvania, West Virginia, Maryland, and Virginia that thins to form a narrow, 40-mile-wide elevated highland in Kentucky, Tennessee, and portions of Alabama and Georgia. Altitudes generally range from 1,000 to 1,800 feet (ft), with lower altitudes along the margins in Alabama and Ohio. Higher-altitude mountainous regions exceed 3,000 ft, with local relief of up to 1,000 ft.

The outcropping sedimentary rocks that compose the Plateaus aquifers range in thickness from 700 to 6,000 ft (Ryder and others, 2008; 2009; 2012) atop a regional trough-like structure (fig. 6). Regional lithofacies and individual formation thicknesses in the Plateaus aquifers vary considerably in Permian-age (McKee and Oriel, 1967), Pennsylvanian-age (McKee and Crosby, 1975), and Mississippian-age (Craig and others, 1979) units. The Allegheny structural front separates the gently folded and flat-lying rocks of the Appalachian Plateaus from the more structurally deformed rocks of the Valley and Ridge Province to the east. Folding in the Appalachian Plateaus rocks generally parallels the northeast-southwest trend of the Allegheny structural front in Pennsylvania and West Virginia, but dissipates to the west and south, coincident with the extent of the underlying Upper Silurian Salina Group (Mount, 2014).

The U.S. Geological Survey (2003) defined the areas of Pennsylvanian and Mississippian rocks in the Appalachian Plateaus as principal aquifer unit of the United States. These rocks have been subdivided by age into five units having similar hydraulic properties: (1) surficial deposits, (2) Permian- and Upper Pennsylvanian-age aquifers, (3) middle and Lower Pennsylvanian-age aquifers, (4) the Lower Pennsylvanian- to Upper Mississippian-age Pennington Formation confining unit (where present), and (5) Mississippian-age carbonate and clastic aquifers (Trapp and Horn, 1997; Lloyd and Lyke, 1995; and Miller, 1990). A thick sequence of Devonian-age shales marks the base of the geologic formations that are defined as principal aquifers in the Appalachian Plateaus (fig. 7).

Primary porosity is low in Plateaus aquifers and probably restricted to sandstone units (Abate, 1993). Diagenetic processes, such as cementation and compaction, have further limited the primary porosity of Plateaus aquifers (Abate, 1993), thereby limiting aquifer storage. Although tectonic processes have enhanced porosity through the development of joints and fractures, groundwater storage capacity in Plateaus aquifers is low and storage coefficients are less than 10^{-4} (Peffer, 1991).

Groundwater flow in Plateaus aquifers is locally controlled by stress-relief fracturing (Wyrick and Borchers, 1981) and by the orientation and permeability of gently dipping and moderately folded sedimentary rocks (Olson and others, 1992; Seaber and others, 1988). In the model developed by Wyrick and Borchers (1981), groundwater flow in Plateaus aquifers occurs primarily in bedding-plane separations beneath valley floors and in horizontal and nearly vertical stress-relief fractures along valley walls (fig. 8). Near-surface flow in valley and upland settings is the result of a network of fractures formed by the unloading of compressional stresses due to erosion of overlying sedimentary rocks. Expanding on the work of Wyrick and Borchers (1981), Sheets and Kozar (2000) and McCoy and Kozar (2006) found a progression of apparent groundwater ages derived from age-dating tracers to support the concept of topographically-driven groundwater flow at shallow (<300 ft) depths. The depth of freshwater flow thins in valley settings despite fracture permeability that is greater than that in surrounding uplands (Heisig and Scott, 2013; Williams, 2010). Using packer tests, Harlow and LeCain (1993) found that horizontal groundwater flow resulting from higher hydraulic conductivity along stratigraphic horizons, particularly coal seams, was more substantial than flow resulting from vertical connections within adjacent sandstone, siltstone, or shale layers. Horizontal permeability is dominated by development of secondary porosity through stress-relief fractures, and stratigraphic-related bedding-plane separations, coal cleats, and solution openings in limestone or dolostone (Ferrell, 1988). Although the rocks are relatively undeformed, stratification of aquifers and confining units forms complex vertical boundaries (Peffer, 1991). Where vertical and horizontal permeability are variable, groundwater flows in a stairstep pattern, alternating among vertical joints, faults, and fractures and horizontal bedding-plane separations (Trapp and Horn, 1997).

Climate

Climate in the Appalachian Plateaus is humid continental to humid subtropical, with 1981 to 2010 mean annual temperatures ranging from about 46 °F in northwestern Pennsylvania to 61 °F in Alabama (PRISM Climate Group, 2012). Annual precipitation in the area ranges from less than about 35 to greater than 60 in/yr, generally increasing with altitude and from north to south (fig. 9). Precipitation patterns are primarily affected by proximity to tropical maritime airmasses, which form in the Gulf of Mexico and western Atlantic Ocean, and by orographic effects of the Appalachian Mountains. Pacific, land-recycled, tropical continental, and polar airmasses also influence climate in the Appalachian Plateaus (Paulson and others, 1991). Most of the precipitation results from frontal systems and convective thunderstorms and is distributed fairly evenly throughout the year. The wettest months of the year tend to range from December to May in Alabama and from March to August in Ohio, Pennsylvania, and West Virginia. Lake-effect moisture from Lake Erie, particularly in winter, increases the quantity of water in northeastern Ohio and western Pennsylvania. Some high-altitude areas in West Virginia receive more than 100 inches of annual snowfall (Paulson and others, 1991).

The Palmer Drought Severity Index (PDSI) can be used to identify periods of relative wet and dry conditions in long-term climate records (Alley, 1984). The PDSI measures the cumulative departure in surface-water balance using temperature and precipitation data to calculate water supply and demand. PDSI data from 1895 to 2010 were obtained from the National Oceanic and Atmospheric Administration for three climatological divisions: Alabama Appalachian Mountains, Kentucky Eastern, and Pennsylvania Southwest Plateaus (National Climatic Data Center, 2013). A graph of the 1930–2010 PDSI from the three climatological divisions in the study area provides the context to evaluate north-to-south climate variation across several historical periods of wet, dry, or variable conditions (fig. 10). Wet periods are defined as positive departures in the annualized PDSI for all three climatological divisions. Negative PDSI departures for all three climatological divisions indicate dry periods. Between 1930 and 2010, five predominantly wet periods and seven predominantly dry periods were identified. Dry periods of variable severity and duration occurred once about every 10 years. Dry periods that extended across the study area were severe from 1930 to 1932 and less severe from 1932 to 1934, 1938 to 1942, 1952 to 1955, 1963 to 1966, 1968 to 1970, 1985 to 1988, and 1998 to 2002 (fig. 10). Wet periods extended across the study area from 1945 to 1952, 1970 to 1976, 1978 to 1980, 1988 to 1998, and 2002 to 2005.

There are some notable differences in the PDSI data between the climatological divisions. Dry conditions appear to have only affected Alabama and Kentucky in the early- to mid-1940s, whereas extended dry conditions only affected Pennsylvania from 1963 to 1970. On a regional scale, variable conditions from 2006 to 2009 are indicated by positive PDSI values in Pennsylvania, but negative values in Kentucky and Alabama.

The PDSI from all three climatological divisions show a general shift from prevailing negative values, or dry conditions, before 1970, to positive values, or wet conditions, during and after 1970 (fig. 10)—a common trend amongst hydrologic datasets in the eastern United States (Patterson and others, 2012; McCabe and Wolock, 2002). The change in climate conditions around 1970 has been correlated to trends in the Atlantic Multidecadal Oscillation (AMO) (McCabe and Wolock, 2002; Enfield and others, 2001) and the North Atlantic Oscillation (NAO) (Lins and Slack, 1999). Even though there are many exceptions, the wet and dry climate conditions defined using the PDSI are generally aligned with the variations in the cycles of the AMO index. Well-known climate indices such as the AMO, NAO, and others are, however, only weakly correlated to long-term variability observed in minimum, mean, and maximum streamflow across the United States (McCabe and Wolock, 2014).

Methods of Investigation

Quantifying the magnitude and distribution of fresh groundwater supplies in the Appalachian Plateaus requires that estimates of hydrologic budget components be developed based on compilation of historic stream-flow data, and available basin characteristics, climate, and water-use data sources. A regional hydrologic budget analysis provides for an essential understanding of water uses in context of the overall fluxes in the system.

Soil-Water-Balance Model

The magnitude and distribution of water-budget components in the Appalachian Plateaus was computed using regionally available spatial datasets and the SWB model (Westenbroek and others, 2010). The SWB model calculates spatial and temporal variations in groundwater recharge based on climatological data, and soil and landscape properties. SWB is a deterministic model that uses gridded data and physically based parameters to apportion water derived from daily precipitation and snowmelt into surface runoff, evapotranspiration, recharge, and water storage in the soil column. Model output consists of gridded distributions of water-budget components, such as surface flow leaving grid cells, actual evapotranspiration (ET), soil moisture, and recharge at a specified cell size within the study area. Computation of water-budget components relies on relations between surface runoff, land cover, and hydrologic soil group (Cronshey and others, 1986) and estimated values of ET and temperature (Hargreaves and Samani, 1985). Water storage in the soil column is estimated using a modified Thornwaite-Mather accounting method on a daily basis (Westenbroek and others, 2010).

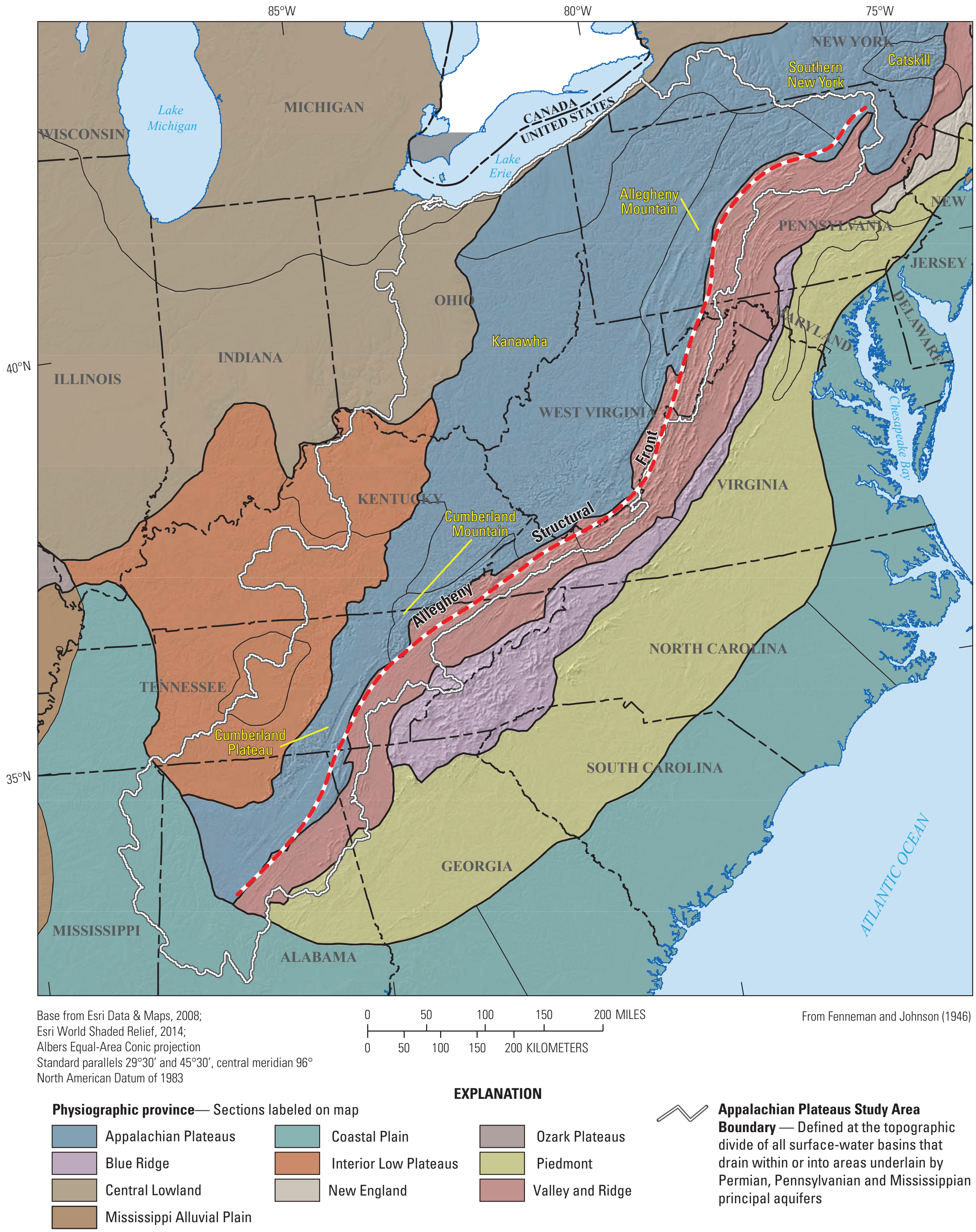


Figure 5. Physiography and the Allegheny Structural Front in the Appalachian Plateaus.

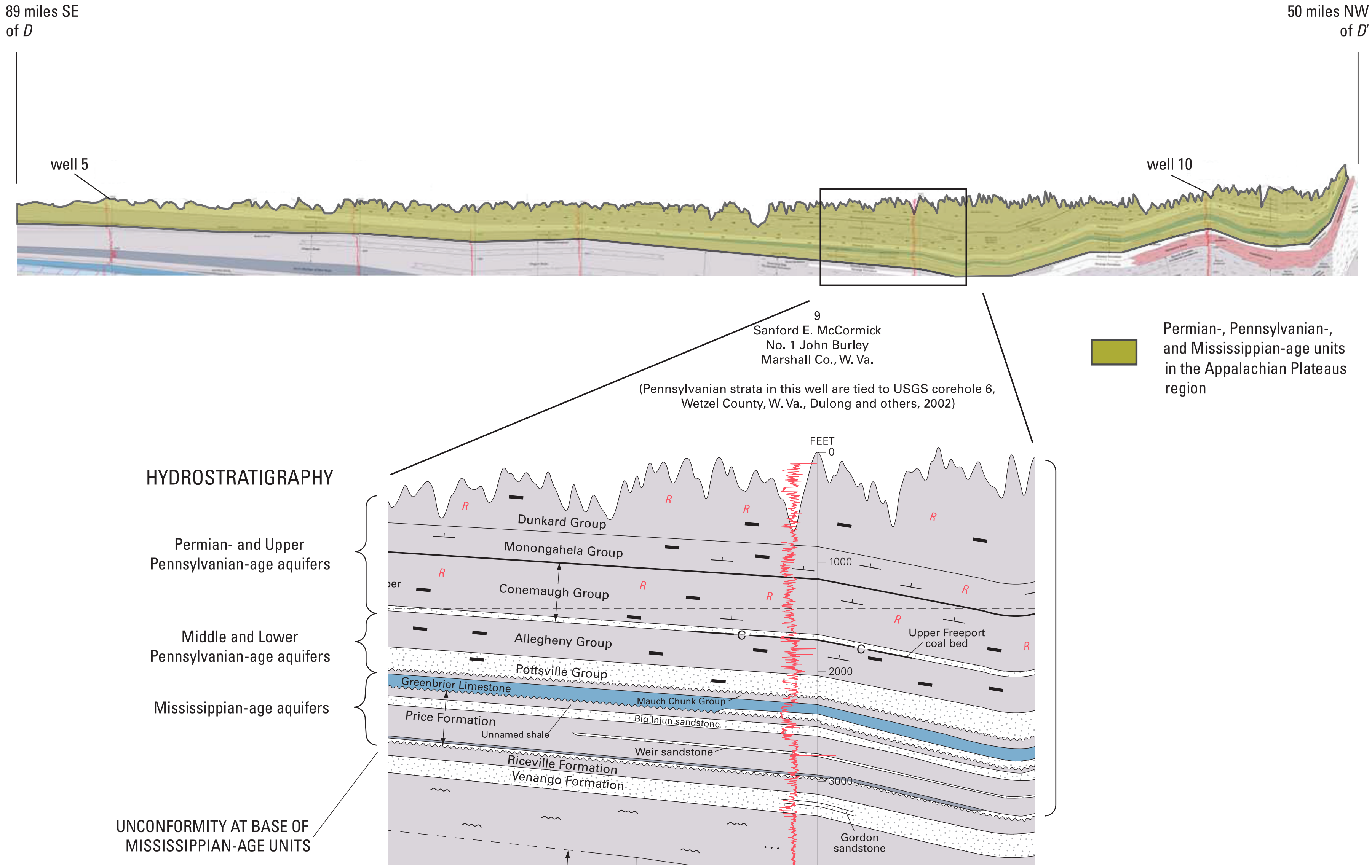


Figure 6. Modified excerpts from cross section *D–D′* of Ryder and others (2009) showing the Permian-, Pennsylvanian-, and Mississippian-age geologic units in the Appalachian Plateaus study area.

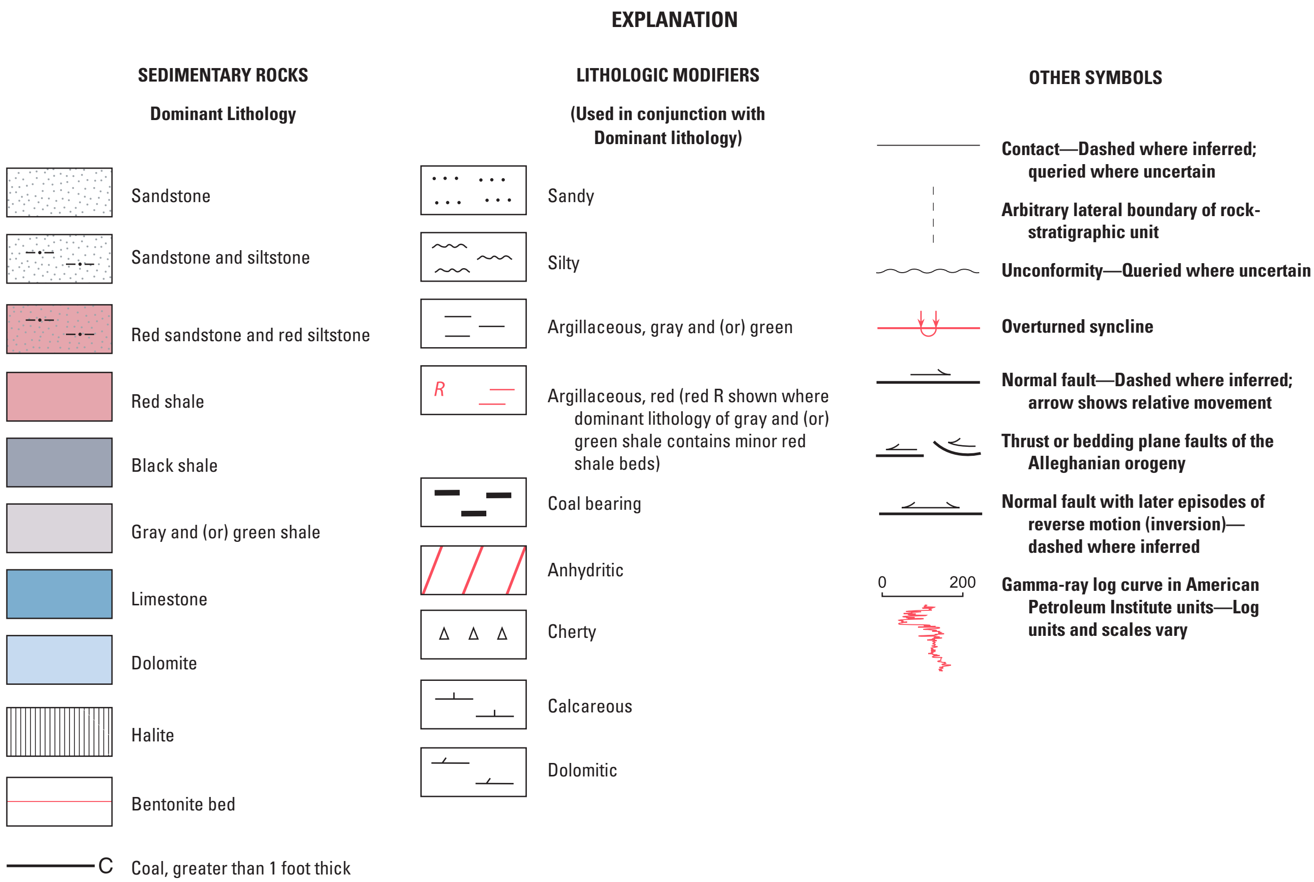


Figure 6. Modified excerpts from cross section *D–D′* of Ryder and others (2009) showing the Permian-, Pennsylvanian-, and Mississippian-age geologic units in the Appalachian Plateaus study area.—Continued

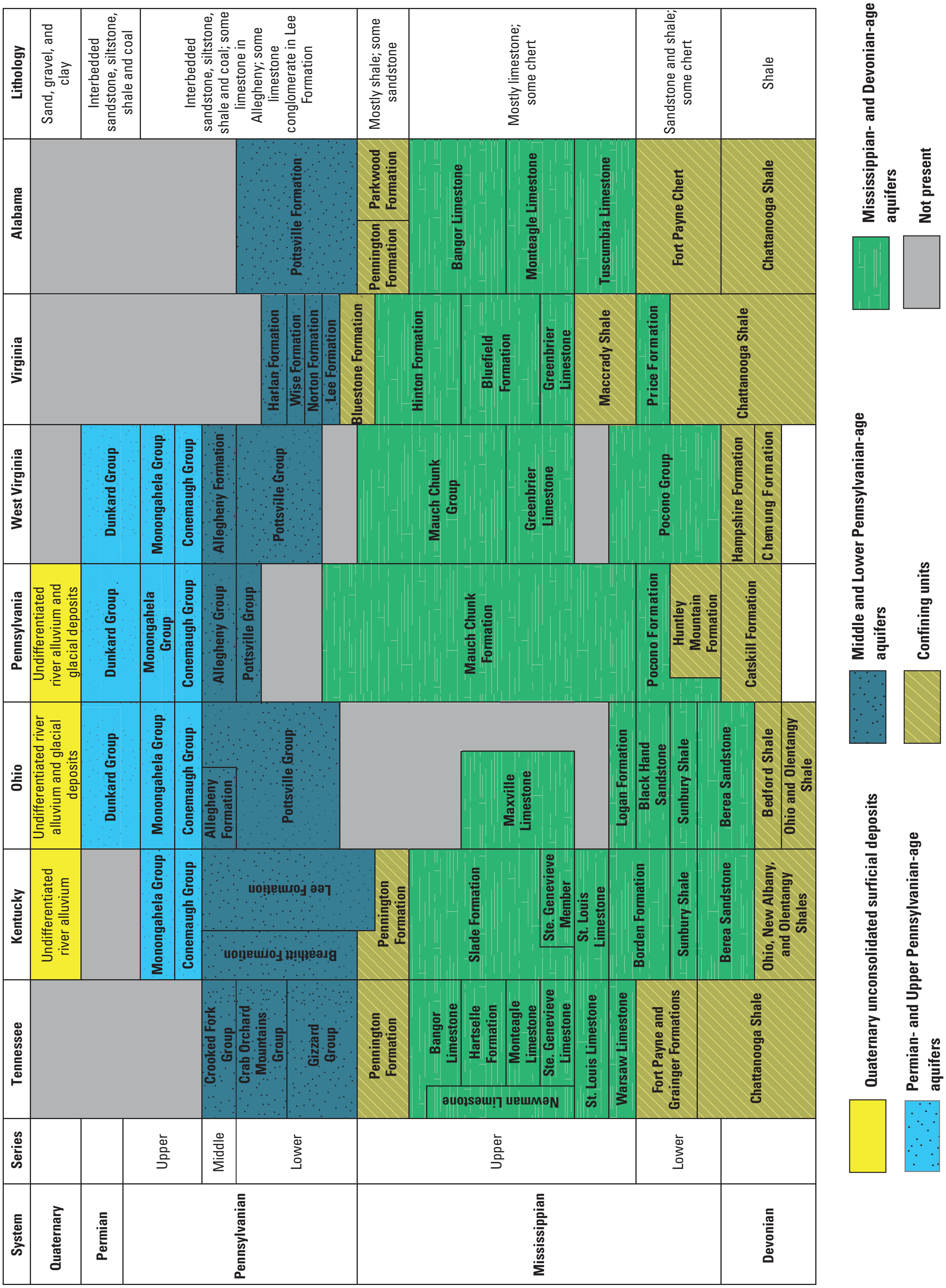


Figure 7. Principal aquifers and hydrostratigraphic units in the Permian-, Pennsylvanian-, and Mississippian-age strata of the Appalachian Plateaus (modified from Trapp and Horn, 1997; Lloyd and Lyke, 1995; Miller, 1990).

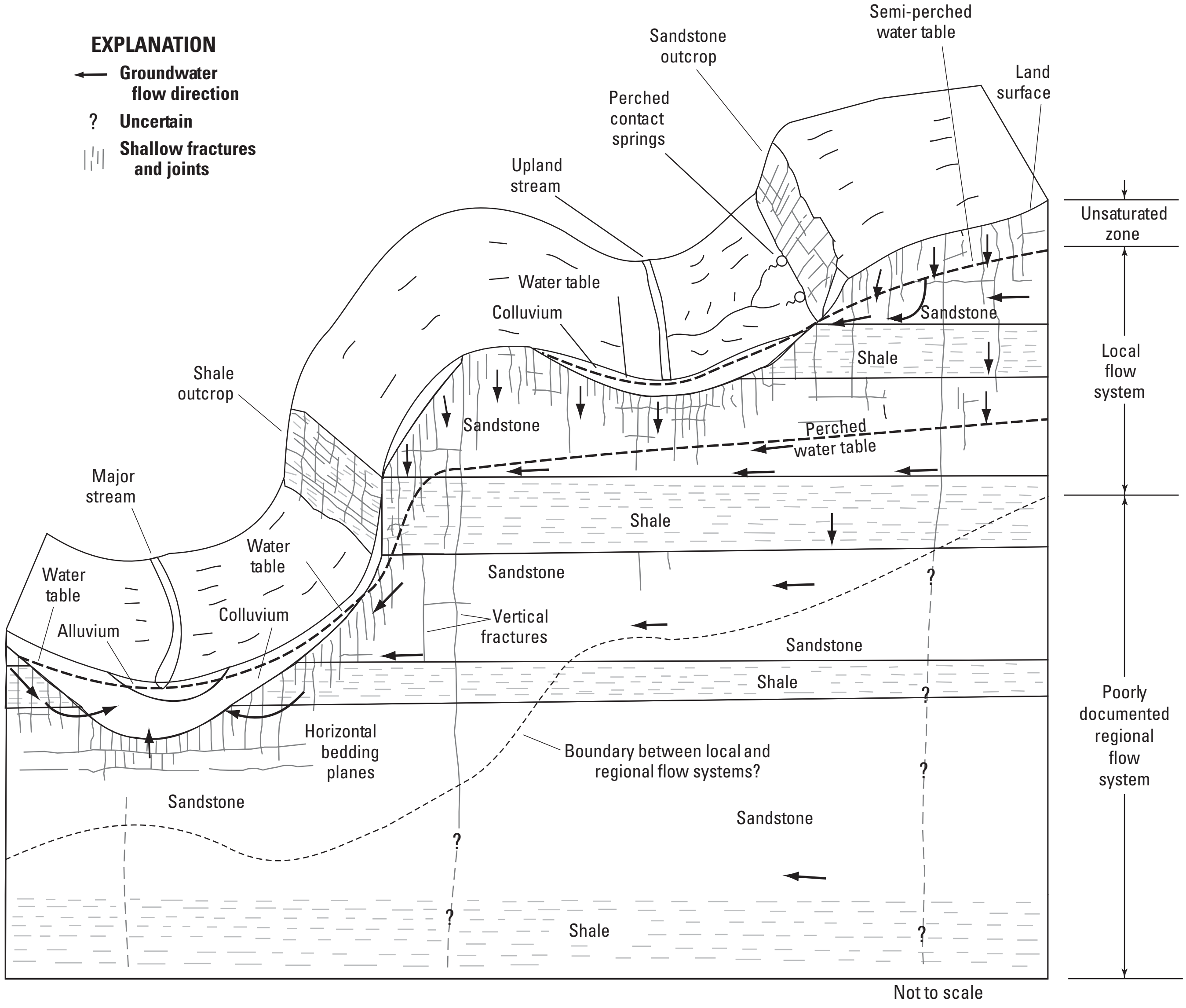
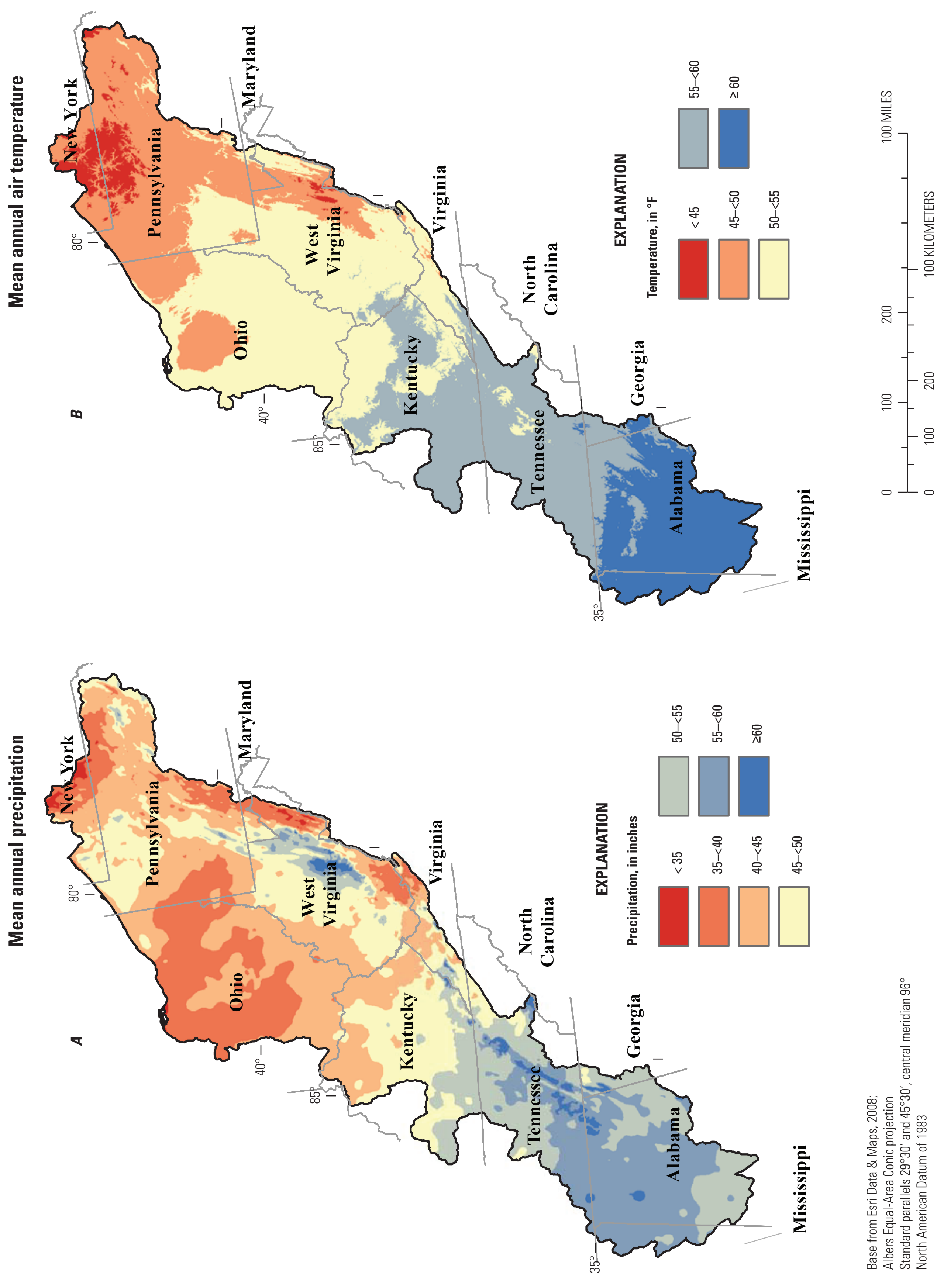


Figure 8. Groundwater flow in the Appalachian Plateaus (modified from Buckwalter and Moore, 2007).



Base from Esri Data & Maps, 2008;
Albers Equal-Area Conic projection
Standard parallels 29°30′ and 45°30′, central meridian 96°
North American Datum of 1983

Figure 9. PRISM maps of *A*, mean annual precipitation, and *B*, mean annual air temperature in the Appalachian Plateaus study area, 1981–2010.

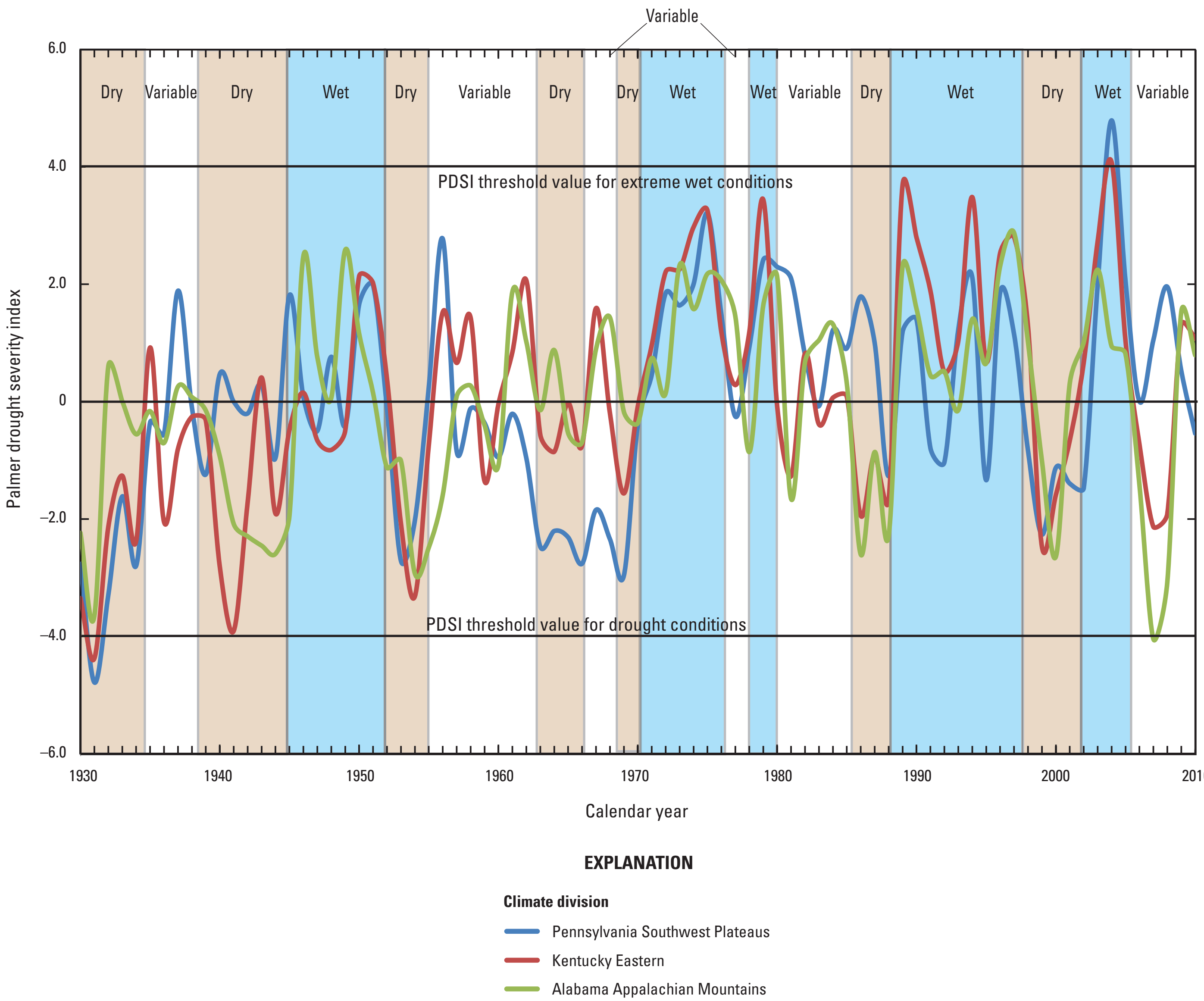


Figure 10. 1930 to 2010 Palmer Drought Severity Index (PDSI) for three climate divisions in the Appalachian Plateaus study area: Pennsylvania Southwest Plateaus (climate division 3609), Kentucky Eastern (climate division 1504), and Alabama Appalachian Mountains (climate division 102). Data from National Oceanic and Atmospheric Administration (2014). Wet periods represent positive departures in the annualized PDSI and the dry periods represent negative PDSI departures for all three climatological divisions.

Water from precipitation and snowmelt is either diverted to surface runoff, intercepted by the plant canopy, consumed by ET, or allowed to infiltrate the soil column. Surface runoff either infiltrates the soil column in downslope cells, discharges to open water bodies, or accumulates in closed surface depressions. Daily accounting of the volume of water stored in the soil column in each grid cell is calculated from the estimated ET for distinct combinations of soil group and land cover. Recharge is computed as surplus water in excess of the maximum soil-water capacity, a product of available water capacity (AWC) and root depth in the soil column. Surplus water in excess of a specified maximum recharge rate for each hydrologic soil group is rejected as recharge and passed to downslope cells.

The SWB model of the Appalachian Plateaus consists of 419,724 grid cells (0.6-mi spacing) across a study area that includes the boundary based on the topographic divide of all surface-water basins that drain within, or into, areas underlain by Pennsylvanian and Mississippian principal aquifers (fig. 11). Climatological data specified for each grid cell, including daily values of precipitation and maximum and minimum daily temperature, were obtained from the Daymet database (Thornton and others, 1997 and 2012) for the period 1980 through 2011.

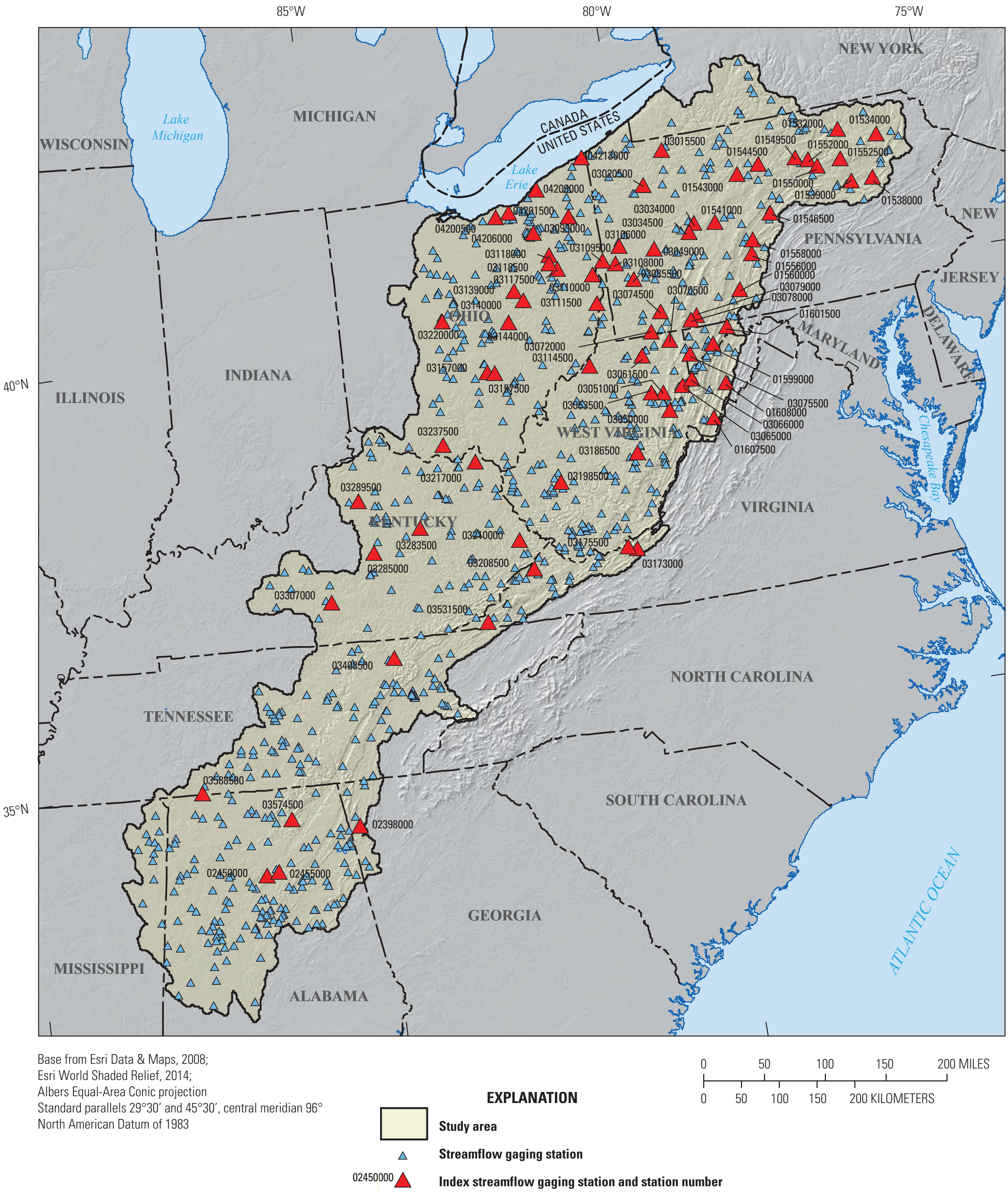


Figure 11. Location of Appalachian Plateaus study-area boundary and streamflow gaging stations where hydrograph separation was conducted (Nelms and others, 2015). Index streamgages were defined as having 60 or more years of complete record between the years 1930 and 2011, no more than 20 percent missing data, and previously defined basin boundaries in the GAGES-II dataset (Falcone, 2011).

Hydrograph Separation

Variability in annual recharge rates was evaluated using annual base-flow yield estimates from PART (Nelms and others, 2015). PART is a hydrograph-separation technique that divides streamflow into its groundwater discharge (Q^{bf}) and surface-runoff (*RO*) components (Rutledge, 1998). PART estimates a daily record of base flow from streamflow records by using a form of streamflow partitioning based on antecedent streamflow recession (Rutledge, 1998). Base-flow discharge is commonly assumed to be equivalent to effective recharge in the Appalachian Plateaus; 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. Nelms and others (2015) used GW Toolbox (Barlow and others, 2015) to estimate annual base-flow yield at 849 continuous-record streamflow gaging stations within the Appalachian Plateaus study area (fig. 11). Gage data covered calendar years 1900 to 2011, although, individual periods of record varied considerably among streamflow gaging stations within that time period. Nelms and others (2015) present the results from additional hydrograph-separation techniques HYSEP and base-flow index (BFI). The study described herein focuses on the results from PART because the program is not affected by watershed scale (Delin and Risser, 2007), thereby allowing comparison among basins of various sizes. All of the hydrograph separation methods have a number of simplifying assumptions:

1. The streamflow hydrographs reflect contributions from two sources: surface runoff in response to a precipitation event and groundwater discharge to streams.
2. Diffuse areal groundwater recharge is uniformly distributed over a watershed, as opposed to focused groundwater recharge such as occurs from losing stream reaches.
3. All groundwater recharge within the basin discharges to the receiving stream network, except that amount that is evapotranspired directly from the groundwater system (sometimes referred to as riparian evapotranspiration).
4. Groundwater discharge to streams is a continuous process (Healy, 2010).
5. Surface-water and groundwater drainage areas to the streamflow gaging station must be coincident.
6. Groundwater is not lost to underlying regional groundwater-flow systems or to groundwater withdrawals.
7. Streams are unregulated and not influenced by reservoirs, streamflow diversions, or wastewater return flows.

A requirement of hydrograph separation is that antecedent recession exceeds the time increment of the data (1 day). To meet this requirement, base-flow yields were only computed for basins with drainage areas larger than 1 mi^2 (Nelms and others, 2015). The upper drainage area limit for a basin to be evaluated with hydrograph separation techniques is somewhat ambiguous and dependent upon the degree of nonuniformity of weather systems (Rutledge, 1998). Nelms and others (2015) selected an upper limit of 500 mi^2 for drainage area, which follows the recommendation established by Rutledge (1998).

Normalized annual anomalies were calculated to synthesize long-term trends (1930–2011) for all hydrograph separation data from Nelms and others (2015) within the Appalachian Plateaus study area. Normalized, or standardized anomalies are defined as

$$N = \frac{(X - \mu)}{\sigma} \quad , \qquad (1)$$

where

- N is the magnitude of the annual anomaly,
- X is the individual yearly value,
- μ is the annual mean over the period of record, and
- σ is the standard deviation of the period of record.

Long-term data trends are highlighted with a 10-year moving average windowing technique, similar to that presented by Weider and Boutt (2010).

To evaluate spatial patterns in the direction and magnitude of change in the water budget at individual gaging stations, a subset of 77 index streamgages were selected from the 849 continuous-record streamflow gaging stations shown on figure 11 (table 1). Index streamgages were defined as having 60 or more years of complete record between the years 1930 and 2011, no more than 20 percent missing data, and previously defined basin boundaries in the GAGES–II dataset (Falcone, 2011). Base-flow estimates from the 77 index streamgages were initially evaluated for the period from 1930 to 2011. Initial evaluation of the dataset showed an abrupt change in base flow consistent with that noted in climate and streamflow conditions around 1970 (McCabe and Wolock, 2002) and 1995 (Enfield and others, 2001). To evaluate shorter-term trends before and after the 1970 shift, the approach of Patterson and others (2012) was followed, whereby the long-term dataset was separated into pre-1970 (1930 to 1969) and post-1970 (1970 to 2011) time periods. For each of the three time periods, the magnitude of change was calculated using the nonparametric Kendall-Theil robust line, also known as the Sen slope. The Sen slope is calculated as the median of all possible pairwise slopes in temporal datasets (Helsel and Hirsch, 2002) and is used to obtain the total magnitude of change over the specified time periods by multiplying its value by the number of years of data being evaluated (Patterson and others, 2012; Hodgkins and others, 2007). This approach has the advantage of being relatively insensitive to outliers and can be more accurate than simple linear regression for skewed data. In the current study, as in Hodgkins and Dudley (2011), the magnitude and spatial patterns of trend slopes are discussed without consideration of statistical significance.

Table 1. List of index streamgages in the Appalachian Plateaus region.

[USGS, U.S. Geological Survey; HCDN-2009, Hydro-Climatic Data Network 2009 from Lins (2012). Latitude and longitude are given in decimal degrees, referenced to North American Datum of 1983; Y/N, yes or no]

USGS station identification number	USGS station name	Drainage area (square miles)	Type of regulation[a]	Period of record	Years of complete record	HCDN-2009 gage (Y/N)	Latitude	Longitude
01532000	Towanda Creek near Monroeton, PA	215	Unregulated	1915–2011	97	Y	41.7070	–76.4847
01534000	Tunkhannock Creek near Tunkhannock, PA	383	Unregulated	1915–2011	97	N	41.5584	–75.8946
01538000	Wapwallopen Creek near Wapwallopen, PA	43.8	Unregulated	1920–2011	92	N	41.0593	–76.0935
01539000	Fishing Creek near Bloomsburg, PA	274	Unregulated	1939–2011	73	Y	41.0781	–76.4311
01541000	West Branch Susquehanna River at Bower, PA	315	Unregulated	1914–2011	98	N	40.8970	–78.6770
01543000	Driftwood Br Sinnemahoning Cr at Sterling Run, PA	272	Unregulated	1914–2011	98	Y	41.4134	–78.1970
01544500	Kettle Creek at Cross Fork, PA	136	Unregulated	1941–2011	71	Y	41.4759	–77.8258
01546500	Spring Creek near Axemann, PA	87.2	Unregulated	1941–2011	71	N	40.8898	–77.7942
01549500	Blockhouse Creek near English Center, PA	37.7	Unregulated	1941–2011	71	Y	41.4737	–77.2308
01550000	Lycoming Creek near Trout Run, PA	173	Unregulated	1914–2011	98	Y	41.4184	–77.0327
01552000	Loyalsock Creek at Loyalsockville, PA	435	Unregulated	1926–73, 1976–2011	84	Y	41.3251	–76.9125
01552500	Muncy Creek near Sonestown, PA	23.8	Unregulated	1941–2011	71	Y	41.3570	–76.5347
01556000	Frankstown Br Juniata River at Williamsburg, PA	291	Partially regulated-Mill	1917–2011	95	N	40.4631	–78.1997

Table 1. List of index streamgages in the Appalachian Plateaus region.—Continued

[USGS, U.S. Geological Survey; HCDN-2009, Hydro-Climatic Data Network 2009 from Lins (2012). Latitude and longitude are given in decimal degrees, referenced to North American Datum of 1983; Y/N, yes or no]

USGS station identification number	USGS station name	Drainage area (square miles)	Type of regulation[a]	Period of record	Years of complete record	HCDN-2009 gage (Y/N)	Latitude	Longitude
01558000	Little Juniata River at Spruce Creek, PA	220	Unregulated	1939–2011	73	N	40.6126	–78.1406
01560000	Dunning Creek at Belden, PA	172	Unregulated	1940–2011	72	N	40.0717	–78.4925
01599000	Georges Creek at Franklin, MD	72.4	Unregulated	1930–2011	82	N	39.4939	–79.0447
01601500	Wills Creek near Cumberland, MD	247	Partially regulated-Mining	1930–2011	82	N	39.6696	–78.7880
01607500	S F South Branch Potomac River at Brandywine, WV	103	Unregulated	1944–2010	67	N	38.6315	–79.2436
01608000	S F South Branch Potomac River nr Moorefield, WV	277	Unregulated	1929–34, 1939–2010	78	N	39.0123	–78.9561
02398000	Chattooga River at Summerville, GA	192	Unregulated	1938–2011	74	N	34.4664	–85.3361
02450000	Mulberry Fork near Garden City, AL	365	Unregulated	1929–2011	83	N	33.9951	–86.7489
02455000	Locust Fork near Cleveland, AL	303	Unregulated	1937–85, 1993–2011	68	N	34.0245	–86.5742
03015500	Brokenstraw Creek at Youngsville, PA	321	Unregulated	1910–2011	102	Y	41.8526	–79.3173
03020500	Oil Creek at Rouseville, PA	283	Unregulated	1933–2011	79	N	41.4817	–79.6953
03034000	Mahoning Creek at Punxsutawney, PA	158	Partially regulated-Mining	1939–2011	73	N	40.9392	–79.0084
03034500	Little Mahoning Creek at McCormick, PA	87.4	Unregulated	1940–2008, 2010–11	71	N	40.8362	–79.1100
03049000	Buffalo Creek near Freeport, PA	137	Unregulated	1941–2011	71	Y	40.7164	–79.6999

Table 1. List of index streamgages in the Appalachian Plateaus region.—Continued

[USGS, U.S. Geological Survey; HCDN-2009, Hydro-Climatic Data Network 2009 from Lins (2012). Latitude and longitude are given in decimal degrees, referenced to North American Datum of 1983; Y/N, yes or no]

USGS station identification number	USGS station name	Drainage area (square miles)	Type of regulation[a]	Period of record	Years of complete record	HCDN-2009 gage (Y/N)	Latitude	Longitude
03050000	Tygart Valley River near Dailey, WV	185	Unregulated	1916–74, 1989–2011	82	Y	38.8093	–79.8817
03051000	Tygart Valley River at Belington, WV	406	Partially regulated-Mining	1908–2010	103	N	39.0293	–79.9359
03053500	Buckhannon River at Hall, WV	277	Partially regulated-Mining	1916–2011	96	N	39.0512	–80.1145
03061500	Buffalo Creek at Barrackville, WV	116	Partially regulated-Mining	1908, 1916–23, 1933–2011	88	N	39.5040	–80.1720
03065000	Dry Fork at Hendricks, WV	349	Unregulated	1941–93, 1996–2010	68	N	39.0723	–79.6228
03066000	Blackwater River at Davis, WV	85.9	Unregulated	1922–90, 1993–2011	88	Y	39.1271	–79.4684
03070500	Big Sandy Creek at Rockville, WV	200	Partially regulated-Mining	1910–17, 1922–2011	98	Y	39.6218	–79.7046
03072000	Dunkard Creek at Shannopin, PA	229	Partially regulated-Mining	1941–2011	71	N	39.7592	–79.9706
03074500	Redstone Creek at Waltersburg, PA	73.7	Partially regulated-Mining	1943–2011	69	N	39.9801	–79.7642
03075500	Youghiogheny River near Oakland, MD	134	Unregulated	1942–2011	70	N	39.4216	–79.4236
03079000	Casselman River at Markleton, PA	382	Unregulated	1921–2011	91	N	39.8599	–79.2288
03080000	Laurel Hill Creek at Ursina, PA	121	Unregulated	1919–2011	93	N	39.8204	–79.3214
03085500	Chartiers Creek at Carnegie, PA	257	Partially regulated-Mining	1920–32, 1941–2011	84	N	40.4006	–80.0964
03093000	Eagle Creek at Phalanx Station OH	97.6	Unregulated	1927–33, 1938–2011	81	N	41.2612	–80.9543
03106000	Connoquenessing Creek near Zelienople, PA	356	Partially regulated-Mill	1920–2011	92	N	40.8170	–80.2423

Table 1. List of index streamgages in the Appalachian Plateaus region.—Continued

[USGS, U.S. Geological Survey; HCDN-2009, Hydro-Climatic Data Network 2009 from Lins (2012). Latitude and longitude are given in decimal degrees, referenced to North American Datum of 1983; Y/N, yes or no]

USGS station identification number	USGS station name	Drainage area (square miles)	Type of regulation[a]	Period of record	Years of complete record	HCDN-2009 gage (Y/N)	Latitude	Longitude
03108000	Raccoon Creek at Moffatts Mill, PA	178	Unregulated	1942–2011	70	N	40.6278	–80.3376
03109500	Little Beaver Creek near East Liverpool, OH	496	Unregulated	1916–2011	96	N	40.6759	–80.5406
03110000	Yellow Creek near Hammondsville, OH	147	Unregulated	1941–2011	71	N	40.5378	–80.7251
03111500	Short Creek near Dillonvale, OH	123	Unregulated	1942–2011	70	N	40.1934	–80.7342
03114500	Middle Island Creek at Little, WV	458	Unregulated	1929–94, 2010	67	N	39.4751	–80.9971
03117500	Sandy Creek at Waynesburg OH	253	Unregulated	1939–2011	73	N	40.6726	–81.2598
03118000	Middle Branch Nimishillen Creek at Canton OH	43.1	Unregulated	1942–2010	69	N	40.8414	–81.3537
03118500	Nimishillen Creek at North Industry OH	172	Partially regulated	1922–2011	90	N	40.7497	–81.3694
03139000	Killbuck Creek at Killbuck OH	464	Unregulated	1931–2011	81	N	40.4815	–81.9860
03140000	Mill Creek near Coshocton OH	27.2	Unregulated	1937–2011	75	Y	40.3628	–81.8624
03144000	Wakatomika Creek near Frazeysburg OH	140	Unregulated	1937–2011	75	Y	40.1326	–82.1479
03157000	Clear Creek near Rockbridge OH	89	Unregulated	1940–2011	72	N	39.5884	–82.5785
03157500	Hocking River at Enterprise OH	459	Unregulated	1932–2011	80	N	39.5651	–82.4746
03173000	Walker Creek at Bane, VA	299	Unregulated	1939–2011	73	Y	37.2682	–80.7095

Table 1. List of index streamgages in the Appalachian Plateaus region.—Continued

[USGS, U.S. Geological Survey; HCDN-2009, Hydro-Climatic Data Network 2009 from Lins (2012). Latitude and longitude are given in decimal degrees, referenced to North American Datum of 1983; Y/N, yes or no]

USGS station identification number	USGS station name	Drainage area (square miles)	Type of regulation[a]	Period of record	Years of complete record	HCDN-2009 gage (Y/N)	Latitude	Longitude
03175500	Wolf Creek near Narrows, VA	223	Unregulated	1909–15, 1939–94, 1997–2011	78	N	37.3057	–80.8498
03186500	Williams River at Dyer, WV	128	Unregulated	1930–10	81	Y	38.3790	–80.4840
03198500	Big Coal River at Ashford, WV	391	Partially regulated-Mining	1909–15, 1931–2011	88	N	38.1798	–81.7115
03208500	Russell Fork at Haysi, VA	286	Unregulated	1927–2011	85	N	37.2071	–82.2957
03210000	Johns Creek near Meta, KY	56.3	Unregulated	1942–92, 1995–2010	67	N	37.5670	–82.4579
03217000	Tygarts Creek near Greenup, KY	242	Unregulated	1941–2010	70	N	38.5642	–82.9521
03220000	Mill Creek near Bellepoint OH	178	Partially regulated-Mining	1944–2011	68	N	40.2484	–83.1738
03237500	Ohio Brush Creek near West Union OH	387	Unregulated	1927–34, 1941–2011	79	Y	38.8037	–83.4210
03283500	Red River at Clay City, KY	362	Partially regulated	1931, 1939–2003, 2007–11	71	N	37.8648	–83.9335
03285000	Dix River near Danville, KY	318	Unregulated	1943–2010	68	Y	37.6420	–84.6608
03289500	Elkhorn Creek near Frank-fort, KY	473	Partially regulated	1916–17, 1940–83, 1988–2011	70	N	38.2687	–84.8147
03307000	Russell Creek near Colum-bia, KY	188	Unregulated	1940–2011	72	N	37.1192	–85.3939
03408500	New River at New River, TN	382	Unregulated	1935–91, 1999–2011	70	N	36.3855	–84.5547
03531500	Powell River near Jones-ville, VA	319	Unregulated	1932–2011	80	N	36.6620	–83.0949
03574500	Paint Rock River near Woodville AL	320	Unregulated	1936–2011	76	Y	34.6243	–86.3064
03588500	Shoal Creek at Iron City, TN	348	Unregulated	1926–93, 2001–11	79	N	35.0241	–87.5790
04200500	Black River at Elyria OH	396	Unregulated	1945–2011	67	N	41.3803	–82.1046

Table 1. List of index streamgages in the Appalachian Plateaus region.—Continued

[USGS, U.S. Geological Survey; HCDN-2009, Hydro-Climatic Data Network 2009 from Lins (2012). Latitude and longitude are given in decimal degrees, referenced to North American Datum of 1983; Y/N, yes or no]

USGS station identification number	USGS station name	Drainage area (square miles)	Type of regulation[a]	Period of record	Years of complete record	HCDN-2009 gage (Y/N)	Latitude	Longitude
04201500	Rocky River near Berea OH	267	Unregulated	1924–34, 1944–2011	79	N	41.4067	–81.8871
04206000	Cuyahoga River at Old Portage OH	404	Regulated	1922–35, 1940–2011	86	N	41.1356	–81.5471
04209000	Chagrin River at Willoughby OH	246	Unregulated	1926–34, 1940–83, 1989–93, 1996–98, 2002–10	70	N	41.6309	–81.4034
04213000	Conneaut Creek at Conneaut OH	175	Unregulated	1923–35, 1951–2011	74	Y	41.9270	–80.6040

[a]Type of regulation:

Partially regulated: Streamflow partially regulated.

Partially regulated-Mill: Streamflow partially regulated by mill.

Partially regulated-Mining: Streamflow partially regulated by mining.

Regulated: Streamflow regulated (only periods of unregulated flow or regulation was minor were used).

Unregulated: Streamflow unregulated.

Previous Regional Hydrologic Investigations

Regional assessments of groundwater resources, including those in the Tennessee River (Zurawski, 1978) and Ohio River Basins (Bloyd, 1974) and entire Appalachian region (Wyrick, 1968; Schneider, 1965), were undertaken to show the need for management of groundwater resources in regional water-supply planning. These assessments described groundwater occurrence, water budgets, and development potential of aquifers in conceptual models of the Plateaus and adjacent physiographic provinces. Seaber and others (1988) described the dominant geologic controls on groundwater flow in the Plateaus, including Zurawski's (1978) conceptual model of the Cumberland Plateau in Tennessee and Wyrick and Borchers (1981) model of stress-relief fracturing in southern West Virginia. The USGS Coal Hydrology Program published many reports in the late 1970s and early 1980s that provided a broad characterization of the hydrology and energy production in the Plateaus as it related to coal extraction in 23 areas of the Eastern Coal Province.

In the following two decades, numerous studies focused on local-scale issues related to stress-relief fracture and coal-mining-related hydrology (that is, Donovan and Fletcher, 1999; Hawkins and others, 1996; Sasowsky and White, 1994). The development of hydrograph separation techniques in the late 1990s led to recharge studies in Alabama (Cook and others, 2009), Pennsylvania (Risser and others, 2008), Ohio (Dumouchelle and Schiefer, 2002), and West Virginia (Kozar and Mathes, 2001). Regional geologic constraints, conceptual models, and hydrostratigraphy of the Plateaus by physiographic province were synthesized by Trapp and Horn (1997) and Lloyd and Lyke (1995).

Hydrologic Budget

A water budget is a quantitative expression of the various components of the hydrologic cycle. The hydrologic budget for any basin must balance the quantity of water entering the basin with the quantity of water leaving the basin. A hydrologic landscape approach was taken in the Appalachian Plateaus to estimate all components of the hydrologic system: surface runoff, base flow from groundwater, and interaction with atmospheric water (precipitation and evapotranspiration). Accounting for all the inputs, outputs, and changes in storage in a basin, a basin water balance can be expressed as follows (Healy and others, 2007):

$$P + Q_{in} = ET + \Delta S + Q_{out}, \qquad (2)$$

In the simplified landscape water budget described by equation 2, precipitation (P) and water flow from adjacent basins (Q_{in}) represents the system input, and flow out of the basin (Q_{out}) and evapotranspiration (ET) represent system outputs. Differences in system inputs and outputs are represented by changes in storage (ΔS). ET is not typically measured directly, but computed by difference from more commonly measured knowns P, Q_{in}, and Q_{out}, assuming ΔS is small. In the Appalachian Plateaus, Q_{out} represents streamflow that is considered to consist of two components under natural conditions:

$$Q_{out} = RO + Q^{bf} , \quad (3)$$

where

RO is surface runoff, and

Q^{bf} is groundwater discharge to streams (base flow).

Equation 3 assumes that all streams are gaining at the regional scale. The Q^{bf} term in equation 3 only represents a portion of the groundwater system. The groundwater system as a separate hydrologic compartment can be expanded to

$$Q_{in}^{gw} = Q^{bf} + RET + \Delta S , \quad (4)$$

where

Q_{in}^{gw} is groundwater recharge, and

RET is riparian evapotranspiration.

At the regional scale of the Appalachian Plateaus study area, it is assumed that (1) groundwater recharge is equal to base-flow discharge to streams ($Q_{in}^{gw} = Q^{bf}$); (2) long-term net changes in storage caused by annual climate fluctuations, coal mining, or other water uses are negligible ($\Delta S = 0$); (3) any loss of streamflow to aquifers is returned to the streams in the study area; (4) RET is included in ET; (5) groundwater inflows originate entirely inside the boundaries of the study area ($Q_{in} = 0$); and (6) all groundwater discharges to streams in the study area. Using the landscape approach, equations 3 and 4 can be substituted in equation 2 and an annual hydrologic balance for the Appalachian Plateaus study area can be written as

$$P = RO + Q^{bf} + ET . \quad (5)$$

Annual hydrologic balances within the study area were developed using SWB for the 1980–2011 calendar years, the entire period for which Daymet input datasets were available at the time of analysis (Oak Ridge National Laboratory Distributed Active Archive Center for Biogeochemical Dynamics, 2013).

Soil-Water-Balance Model Analysis

For the SWB model input, landscape data, including land cover (fig. 12) and surface slope, were obtained from the National Land Cover Database (Multi-Resolution Land Characteristics Consortium, 2001) and 30-meter digital elevation model (DEM), respectively. The DEM was processed to eliminate all closed depressions in the elevation data to prevent internal drainage. No closed surface depressions with the potential to serve as an internal drain at the scale of the SWB model (0.4-m^2 grid cells) were noted in the DEM prior to processing. An additional land cover class was created for the study based on the geographic distribution of mined areas. Deciduous forest covers nearly half of the model area, and about 10 percent of the deciduous forest is underlain by either surface or underground coal mines (table 2). Pasture and hay or cultivated crops cover about a quarter of the study area.

Table 2. Distribution of 2001 land cover in the Appalachian Plateaus study area.

[AP, Appalachian Plateaus; 2001 land cover data from Multi-Resolution Land Characteristics Consortium (2001)]

Land cover class	Code	Percentage of total AP study area
Open water	11	1.4
Developed, open Space	21	5.7
Developed, low intensity	22	2.3
Developed, medium and high intensity	23/24	1.1
Mined areas of deciduous forest	40	5.6
Deciduous forest	41	44.3
Evergreen forest	42	4.5
Mixed forest	43	4.5
Shrub/scrub	52	2.1
Grassland/herbaceous	71	2.8
Pasture/hay	81	15.1
Cultivated crops	82	8.9
Woody wetlands	90	1.3

Hydrologic soil groupings (fig. 13) and AWC (fig. 14) were obtained from the Soil Surface Geographic Database (SSURGO) (U.S. Department of Agriculture, Natural Resources Conservation Service, Soil Survey Staff, variously dated). SSURGO data defines four hydrologic soil groups, A through D, that range from low to high runoff potential, high to low infiltration capacity, and less than 10 percent to greater than 40 percent clay content, respectively (U.S. Department of Agriculture, 2007). Hydrologic soil groups B and C underlie nearly three-quarters of the Appalachian Plateaus SWB model area (table 3). The remainder of the area is mostly underlain by soil group D or is composed of dual soil groups in which the water table is within 24 inches of relatively permeable

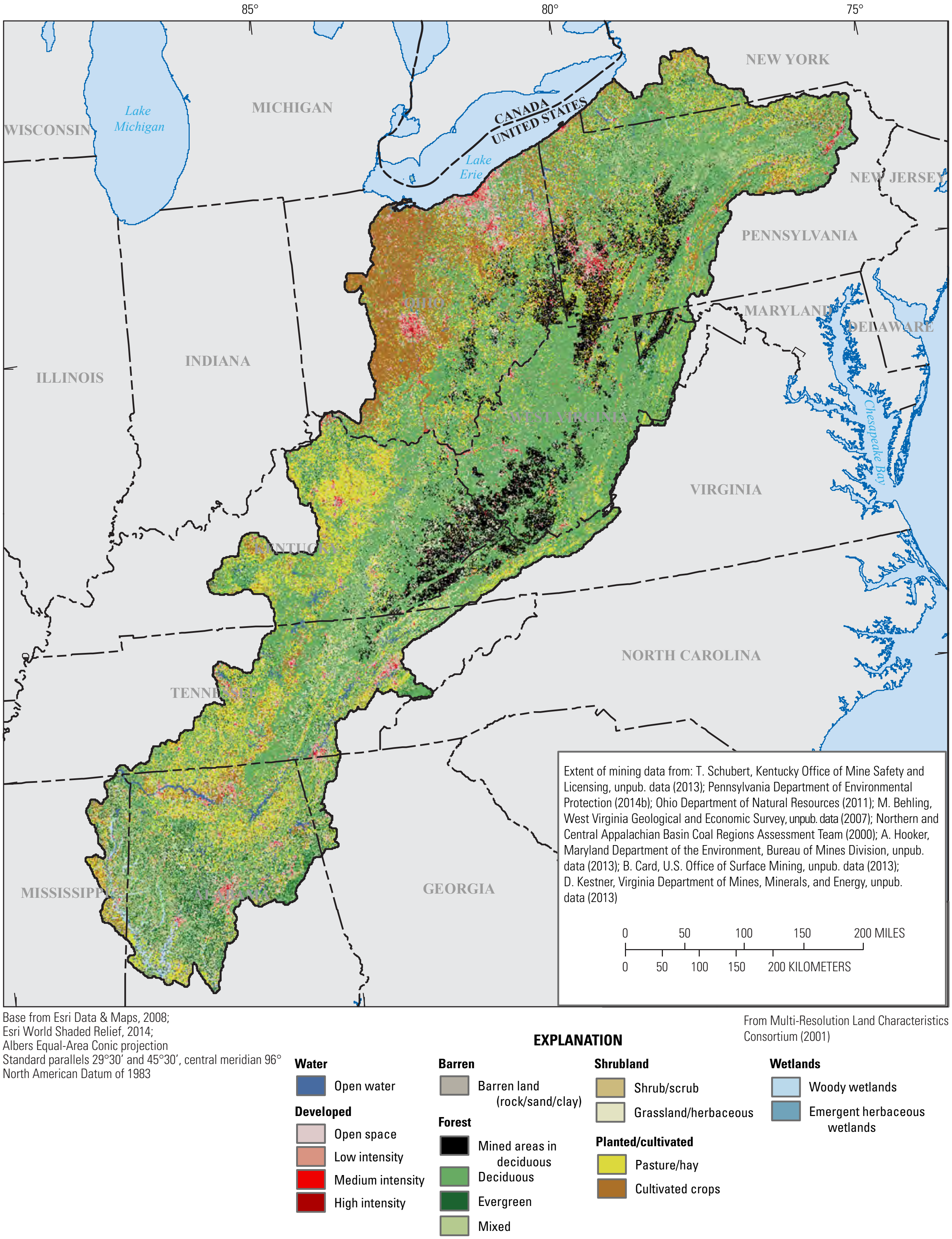


Figure 12. Land-use classes within the Appalachian Plateaus Soil-Water-Balance (SWB) model area.

Table 3. Distribution of soils in the Soil-Water-Balance (SWB) model of the Appalachian Plateaus study area.

[AP, Appalachian Plateaus]

Hydrologic soil group	Percentage of total AP study area
A	4.7
B	32.8
C	40.3
D	12.6
A,B,C/D[a]	9.6

[a]Saturated soils underlain by group D.

soils in groups A, B, or C. Soil properties in these dual soil-group areas were assumed to reflect those of soil group D, because the shallow water table increases the runoff potential of otherwise permeable soils.

Model Sensitivity Analysis

Annual recharge values computed by the SWB model where compared with values of annual base flow computed by the hydrograph separation program PART for 20 watersheds within the Appalachian Plateaus study area (fig. 15, table 4). All 20 watersheds have at least a 20-year record of streamflow data for the model period (1980 through 2011). The sensitivity analysis assumed that base flow estimated by PART reflects long-term groundwater discharge that is equal to recharge. This assumption is reasonable because PART underestimates groundwater discharge during runoff events, a volume of water that can be calculated more accurately by chemical hydrograph separation (Sanford and others, 2012).

The calibration watersheds range in size from 7.8 to 391 mi^2, with average base flows ranging from 7.0 to 18.9 in/yr. The median base flow is 10.7 in/yr. The median unit base flow for the calibration watersheds, computed by dividing the median base flow (10.7 in/yr) by the median drainage area (125 mi^2) and converting units, is 0.79 cubic foot per second per square mile ($[ft^3/s]/mi^2$) (cfsm). Four of the watersheds located at higher altitudes near the eastern boundary of the Appalachian Plateaus with Valley and Ridge Province have unit base flows that exceed 1.3 cfsm, much greater than that of other watersheds (table 4); 3 are in Pennsylvania (watersheds 1, 2, and 6) and 1 is in West Virginia (watershed 10). Watershed 9 in Ohio has a unit base flow value of 0.34 cfsm and generates much less base flow than the other watersheds. The predominant vegetation in 13 of the 20 watersheds is deciduous forest that overlies either hydrologic soil group B or C (tables 4, 5). The soils that underlie most of the other watersheds are generally mixtures of groups B and C. Three of the watersheds have been altered by extensive coal mining.

The SWB model was calibrated by adjusting three model parameters that reflect soil properties: root-zone depth, curve number, and plant-canopy interception values associated with soil groups B and C in areas of deciduous forest. Root-zone depth is used with the AWC grid to define the maximum soil water capacity for each SWB grid cell. Maximum root-zone depths for soils in the Appalachian Plateaus SWB model area range from 1.9 to 13 ft (Norby and others, 2004; Canadell and others, 1996). Curve numbers define the runoff potential for each of the soil groups using the Natural Resources Conservation Service (NRCS) runoff curve number method, a relation between the largest annual storm runoff and the associated rainfall (Woodward and others, 2002). Plant-canopy interception is an abstraction term that can reduce runoff by accounting for the amount of rainfall that is trapped on plant surfaces.

Ranges of values for the parameters were assessed for the 13 calibration watersheds mainly underlain by either soil group B or C (table 4) by comparing the annual recharge computed by SWB with the annual base flow in these watersheds as computed by PART. The values of soil parameters specified for soil groups A and D and for land cover classes other than deciduous forest had little effect on recharge computed by the SWB model for the Appalachian Plateaus because the areas associated with them were relatively small. Values for these insensitive parameters were derived from Westenbroek and others (2010). The maximum daily recharge associated with soil groups B and C was a sensitive parameter in the SWB model. The effect of this parameter was negatively correlated with that of root-zone depth; that is, recharge computed by SWB was directly proportional to maximum daily recharge and inversely proportional to root-zone depth. As a result, it was not possible to estimate values for these two parameters jointly, so maximum daily-recharge values from Westenbroek and others (2010) were specified for all soil groups in the SWB model.

Differences between PART base flow and SWB recharge on an annual basis (termed residuals) were squared and summed for each watershed to yield the sum-of-squared errors (*SSE*). The standard error for each watershed (*SE*) was then computed as

$$SE = \sqrt{\frac{SSE}{n}}\ , \qquad (1)$$

where

n is the number of years of record (table 4).

SE and SSE are strictly used as measures of relative model error to compare model sensitivity to different parameter values. Surface plots of SSE for the 13 calibration watersheds assigned to either soil group B or C indicate the relative sensitivity of the SWB model to different pairs of specified parameter values (fig. 16). The curve numbers of soil groups B and C have little effect on the SSE and, therefore, little effect on recharge computed by the SWB model for the calibration watersheds (fig. 16*A*, *B*, *E*, *F*). In contrast, the SWB model is sensitive to values of both the root-zone depth and the plant-canopy interception (fig. 16*A*–*F*).

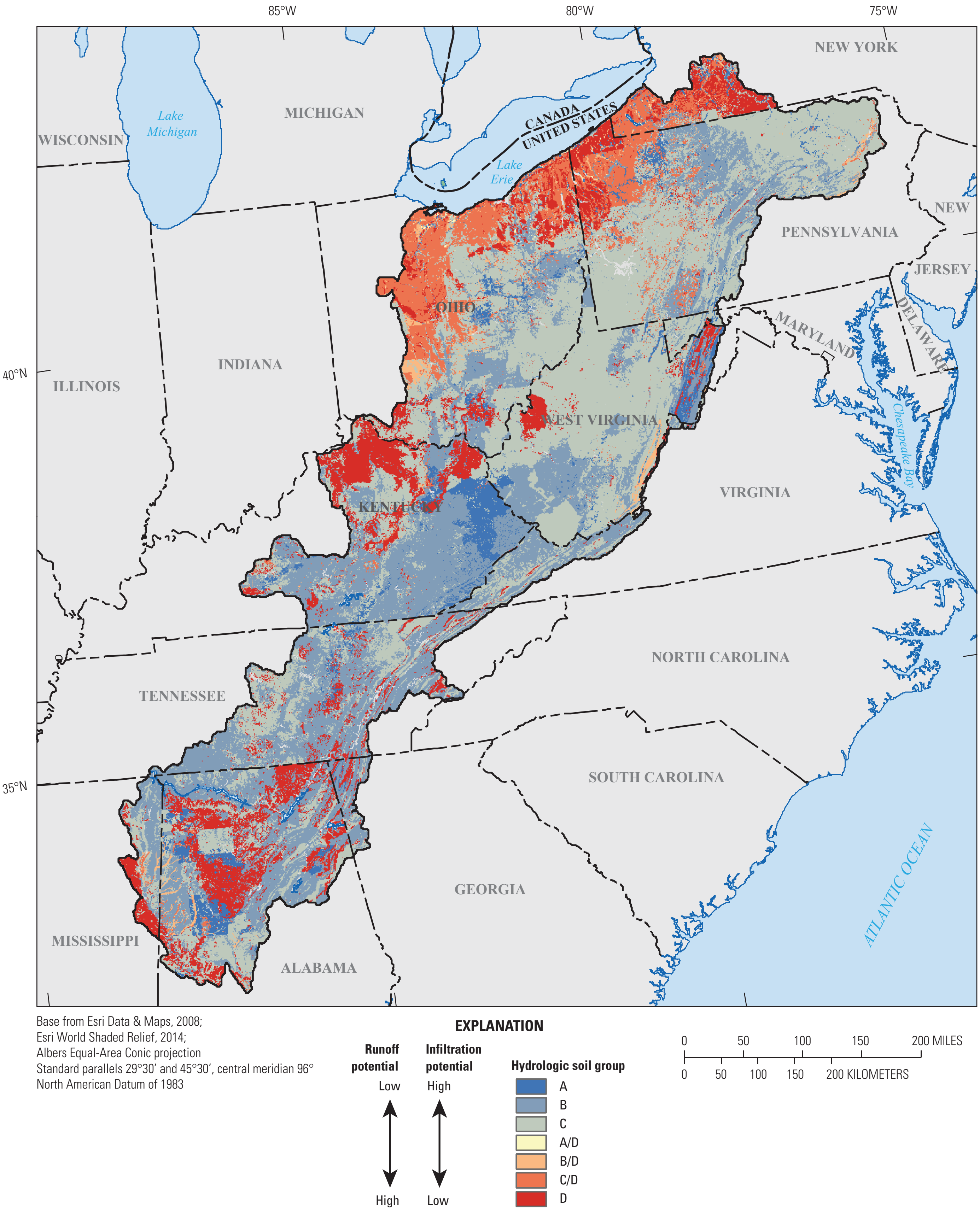


Figure 13. Major hydrologic soil groups from the Soil Surface Geographic Database (SSURGO) (U.S. Department of Agriculture, Natural Resources Conservation Service, Soil Survey Staff, variously dated) within the Appalachian Plateaus Soil-Water-Balance (SWB) model area.

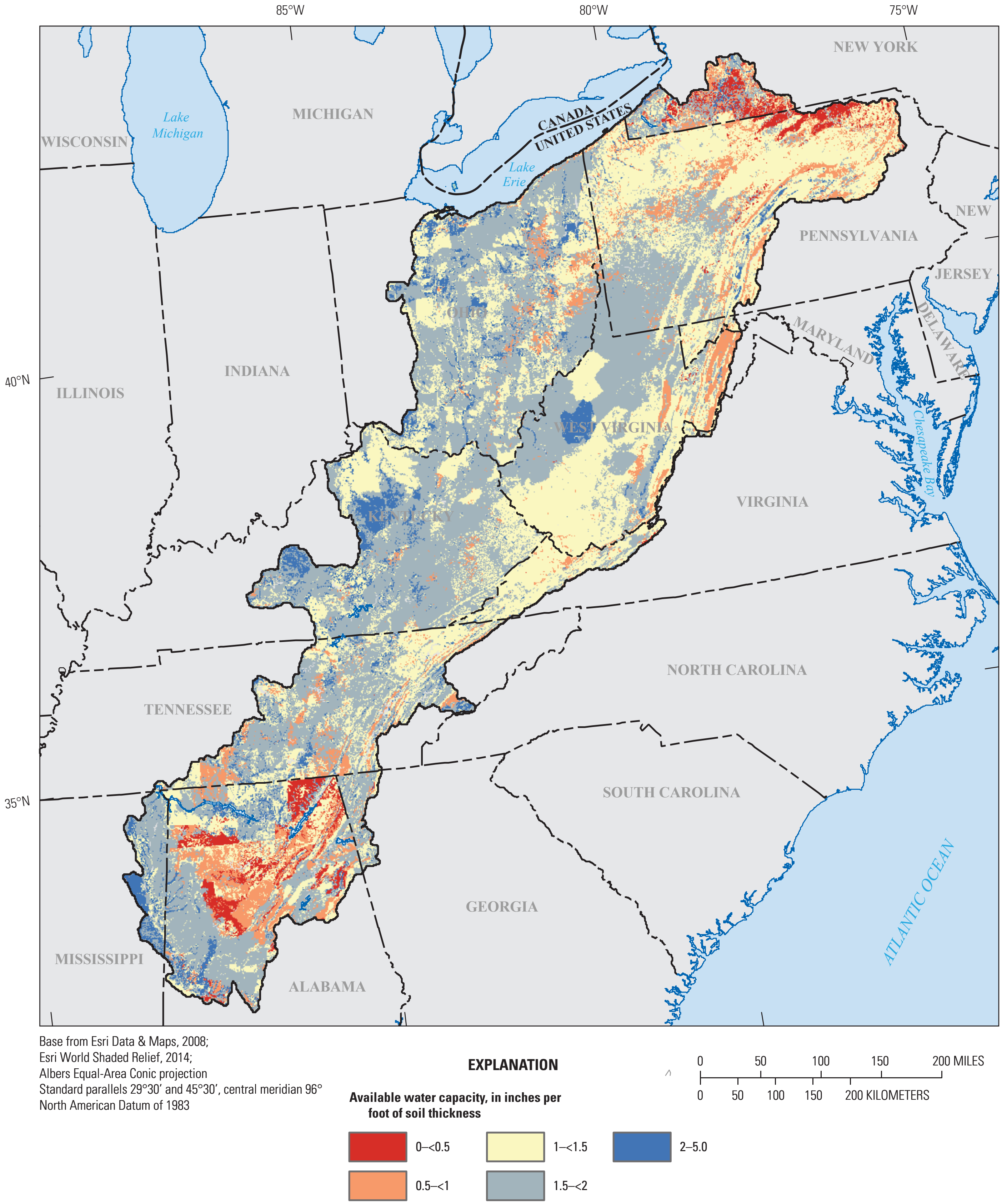
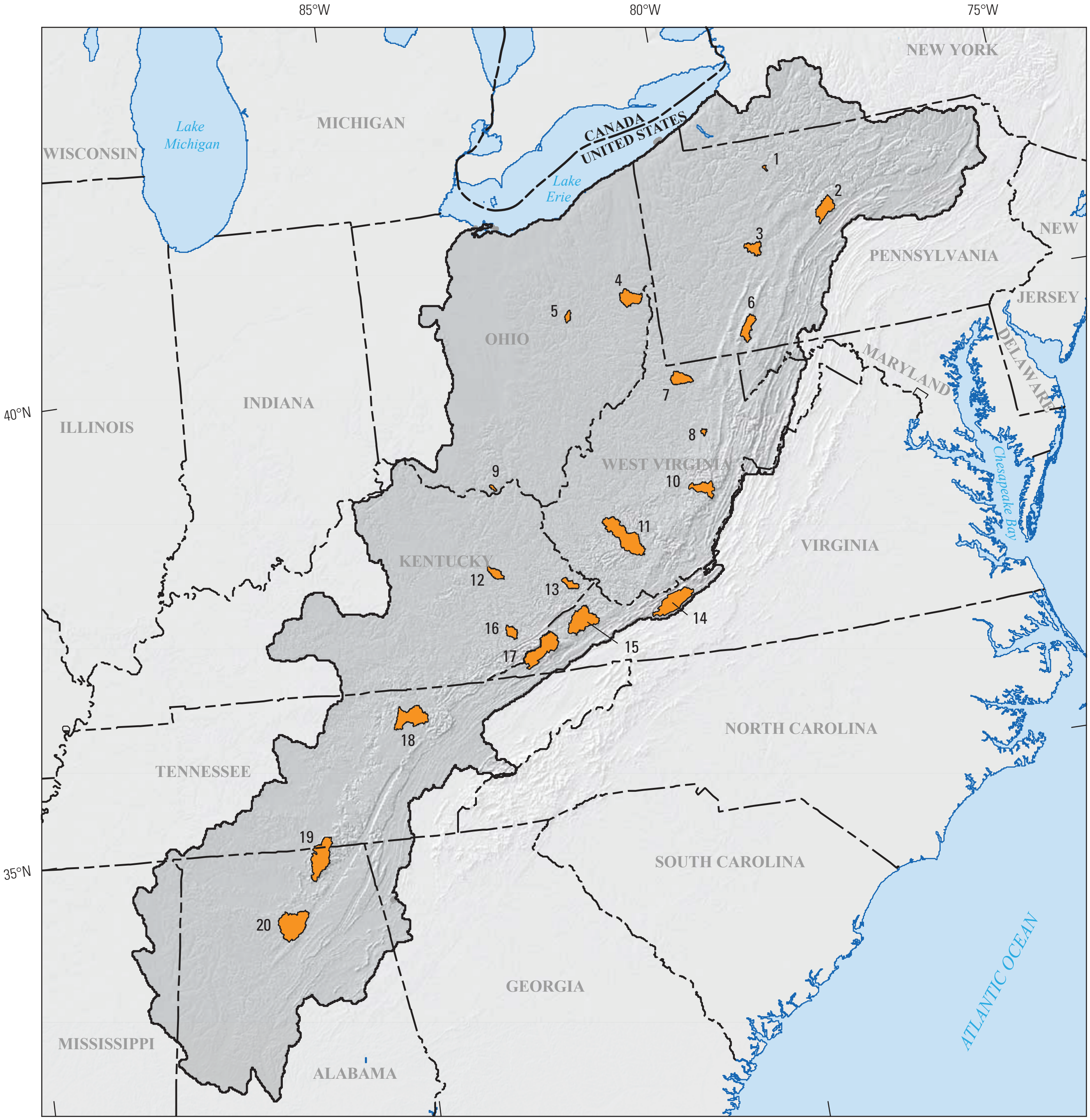


Figure 14. Available water capacity of soils from the Soil Surface Geographic Database (SSURGO) (U.S. Department of Agriculture, Natural Resources Conservation Service, Soil Survey Staff, variously dated) within the Appalachian Plateaus Soil-Water-Balance (SWB) model area.



Base from Esri Data & Maps, 2008;
Esri World Shaded Relief, 2014;
Albers Equal-Area Conic projection
Standard parallels 29°30′ and 45°30′, central meridian 96°
North American Datum of 1983

EXPLANATION

Calibration streamgaging station basin with watershed number

Study area

0 50 100 150 200 MILES

0 50 100 150 200 KILOMETERS

Figure 15. Location of watersheds used for calibration of Soil-Water-Balance (SWB) model.

Table 4. Watersheds selected to calibrate the Soil-Water-Balance (SWB) model of the Appalachian Plateaus study area.

[USGS, U.S. Geological Survey. Gray shading indicates watersheds used in calibration of soil groups B and C properties. SWB, Soil Water Balance]

Watershed	USGS site number	USGS site name	Drainage area (square miles)	Average base flow from PART, 1980 through 2011		Numbers of years of record, 1980 through 2011	Hydrologic soil group	Standard error[a] (inches)	Percent error[b]
				Inches	Cubic feet per second, per square mile ([ft³/s]/mi²)				
1	3026500	Sevenmile Run near Rasselas, PA	7.8	18.3	1.35	31	C	5.2	28.6
2	1547950	Beech Creek at Monument, PA	152	17.9	1.32	31	B/C	5.8	32.2
3	3034500	Little Mahoning Creek at McCormick, PA	87.4	13.5[c]	0.99	30	C	5.4	39.7
4	3110000	Yellow Creek near Hammondsville, OH	147	9.7	0.71	31	B/C	1.5	15.8
5	3140000	Mill Creek near Coshocton, OH	27.2	7.9	0.58	31	C	2.6	32.5
6	3080000	Laurel Hill Creek at Ursina, PA	121	18.5	1.36	31	B	4.2	22.8
7[d]	3061500	Buffalo Creek at Barrackville, WV	116.0	8.7	0.64	31	C	1.4	15.6
8	3052500	Sand Run near Buckhannon, WV	14.3	13.9	1.02	31	C	4.0	28.5
9	3237280	Upper Twin Creek at McGaw, OH	12.2	7.0	0.34	31	C	2.2	32.0
10	3186500	Williams River at Dyer, WV	128.0	18.9	1.39	31	C	6.0	31.8
11[d]	3198500	Big Coal River at Ashford, WV	391.0	11.0	0.81	31	B	2.0	18.0
12	3282500	Red River near Hazel Green, KY	65.8	7.9[c]	0.57	24	B	3.3	41.3
13[d]	3210000	Johns Creek near Meta, KY	56.3	8.3[c]	0.61	28	B	0.9	11.4
14	3173000	Walker Creek at Bane VA	299	9.3	0.68	31	B/C	2.3	25.0
15	3208500	Russell Fork at Hyasi, VA	286	8.9	0.66	31	B	3.0	33.0

Table 4. Watersheds selected to calibrate the Soil-Water-Balance (SWB) model of the Appalachian Plateaus study area.—Continued

[USGS, U.S. Geological Survey. Gray shading indicates watersheds used in calibration of soil groups B and C properties. SWB, Soil Water Balance]

Watershed	USGS site number	USGS site name	Drainage area (square miles)	Average base flow from PART, 1980 through 2011: Inches	Average base flow from PART, 1980 through 2011: Cubic feet per second, per square mile ([ft^3/s]/mi^2)	Numbers of years of record, 1980 through 2011	Hydrologic soil group	Standard error[a] (inches)	Percent error[b]
16	3280700	Cutshin Creek at Wooton KY	61.3	9.0	0.66	31	B	5.2	57.1
17	3531500	Powell River near Jonesville VA	319	13.1	0.96	31	B/C	1.7	13.1
18	3409500	Clear Fork near Robbins TN	272.0	10.5[c]	0.77	23	B/C	1.8	17.4
19	3574500	Paint Rock River near Woodville AL	320.0	11.5	0.85	31	D	3.0	25.9
20	2450000	Mulberry Fork near Garden City AL	365	12.8	0.94	31	B/C/D	3.0	23.4

[a]Residual between base flow computed by PART and recharge computed by SWB model.

[b]Standard error in inches, divided by PART baseflow in inches.

[c]Less than 31 years of record between 1980–2011.

[d]Extensive mining in watershed.

Table 5. Soil properties used in Soil-Water-Balance (SWB) model for deciduous forest underlain by hydrologic soil groups B and C.

[Shaded values were estimated]

Soil group	Mined or unmined	Curve number[a]	Maximum daily recharge (inches per day)	Root zone depth (feet)	Plant-canopy interception (percent)
B					
	Unmined	65	0.6	3	3.5
	Mined	60	0.6	4	3.5
C					
	Uunmined	75	0.24	1.5	2
	Mined	70	0.24	2.5	2

[a]Higher number corresponds to higher runoff potential.

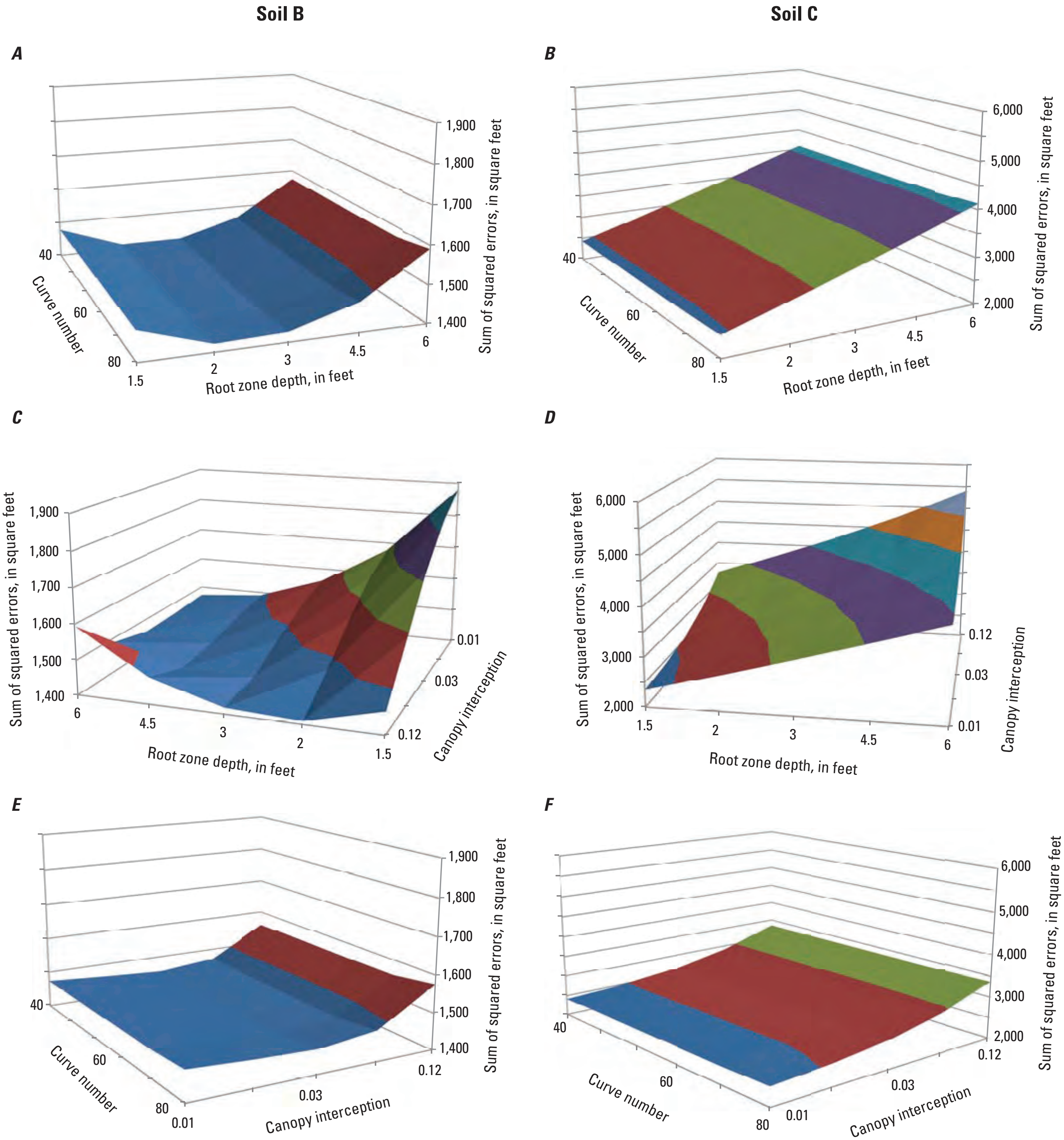


Figure 16. Soil-Water-Balance (SWB) model sum of squared error surfaces for (*A*, *C*, *E*) hydrologic soil type B and (*B*, *D*, *F*) hydrologic soil type C.

The values of root-zone depth and plant-canopy interception that yielded the lowest *SSE* for the calibration watersheds assigned to soil groups B and C were 3 and 1.5 ft, and 3.5 and 2 percent, respectively (table 5). The estimated root-zone depth for soil group B is deeper than for soil group C, providing a greater water-holding capacity. This result is consistent with the characterization of soil group B as coarse-textured and lower clay content than soil group C (Natural Resources Conservation Service, 2007). The estimated plant-canopy interception values for each soil group were similar, which is fortuitous because only one value can be specified for the entire domain of the SWB model. The root-zone depths in mined areas were increased by 1 ft and the curve numbers were decreased (table 5) to reflect the assumed increase in soil porosity in response to settling of the land surface caused by subsurface coal mining.

Model Fit

The *SE* in residuals between PART base flow and SWB recharge on an annual basis was 3.7 and 4.2 inches for the soil group B and C watersheds, respectively, which is an error of about 30 percent from the average annual base flow computed by PART. The largest errors in soil group B were for watersheds 6 and 16, where recharge was underestimated and overestimated, respectively, by the SWB model (fig. 17*A*). Watershed 6 is located at high altitudes (median 2,162 ft) near the boundary of the Appalachian Plateaus with the Allegheny Structural Front in Pennsylvania and has a base-flow of 1.36 cfsm. Watershed 16 is located in the Cumberland Plateau of eastern Kentucky. The largest errors in soil group C were for watersheds 1 and 10, where recharge was underestimated by the SWB model (fig. 17*B*). Both watersheds have unit base flow values greater than 1.3 cfsm and are located at high altitudes (median 2,014 and 3,468 ft, respectively) near the Allegheny Structural Front in Pennsylvania and West Virginia, respectively, where mean annual temperatures are below 50 °F (fig. 9) and fracturing in response to tectonic folding in Mississippian rocks has potentially enhanced aquifer permeability (Wyrick, 1968; Schneider, 1965). The SWB recharge compares favorably with the PART base flow for the remaining seven watersheds that contain mixtures of soil groups B and C (fig. 17*C*), but recharge was underestimated by the SWB model for watershed 2. This watershed also has a unit base flow value of greater than 1.3 cfsm and is located near the Allegheny Structural Front in Pennsylvania. These results indicate that other processes not accounted for in SWB are causing increased base flow in these high altitude areas near the Allegheny Structural Front.

Annual recharge computed by SWB was compiled for 297 watersheds within the study area for which annual base flow was computed with PART to validate the SWB results. All 297 watersheds have a streamflow record greater than 10 years during the calibration period for the SWB model (1980 through 2011). Residuals for all watersheds were calculated as base flow from PART minus recharge calculated from SWB. The SWB recharge compares to the PART base flow (fig. 18*A*) with a *SE* in residuals of 4.1 inches (fig. 18*B*), which is comparable to the *SE* for the 13 calibration watersheds. The SWB recharge is generally less than the PART base flow, however, particularly for watersheds with base flows greater than 20 in/yr (fig. 18*A*). The pattern in the residual plot (fig. 18*C*) indicates a consistent bias in SWB recharge compared to groundwater discharge from PART, with increasing underprediction of recharge for watersheds with unit base flow values larger than 1 cfsm. The spatial distribution of the residuals indicates that most of these watersheds are either located in the northeastern portion of the study area or near the Allegheny Structural Front (fig. 19). It is possible that results from PART could include erroneous groundwater sources in high-altitude, high-relief areas near the Allegheny Structural Front, where overprediction of base flow by PART could result from snowmelt runoff (Rutledge, 1998), interflow on convex hillslopes (Freeze and Cherry, 1979), or slow-draining wetlands (Neff and others, 2006).

Model Limitations

SWB results can be used to construct a generalized representation of recharge throughout the study area, and to estimate the magnitude of annual recharge for years when climatological data are available. The SWB model is designed to simulate several processes that affect the infiltration of precipitation through the soil column. As a result, it is necessary to specify values for many parameters in the model to cover the complete range of soil groups and land covers that are present within the study area. Although model discretization (0.6 mi) is sufficient to represent the relatively detailed spatial distribution of soils and land cover, only general information is available concerning the properties of the soils themselves (for example, root-zone depth and maximum saturated hydraulic conductivity). Some of the properties for the combinations of soil groups and land cover that compose most of the study area could be estimated through calibration against the base flows computed by PART, but many other parameters were insensitive and required specified values. In addition, the properties associated with a particular soil group and land cover combination could vary from one part of the study area to another. The disparity between the SWB and PART results reflects the spatial differences in soil properties within the study area, and the uncertainty in the values of those properties. The bias in results from large unit base flow watersheds (figs. 18 and 19) indicate other hydrologic properties or processes are not explicitly accounted for in the SWB model or are misrepresented by PART.

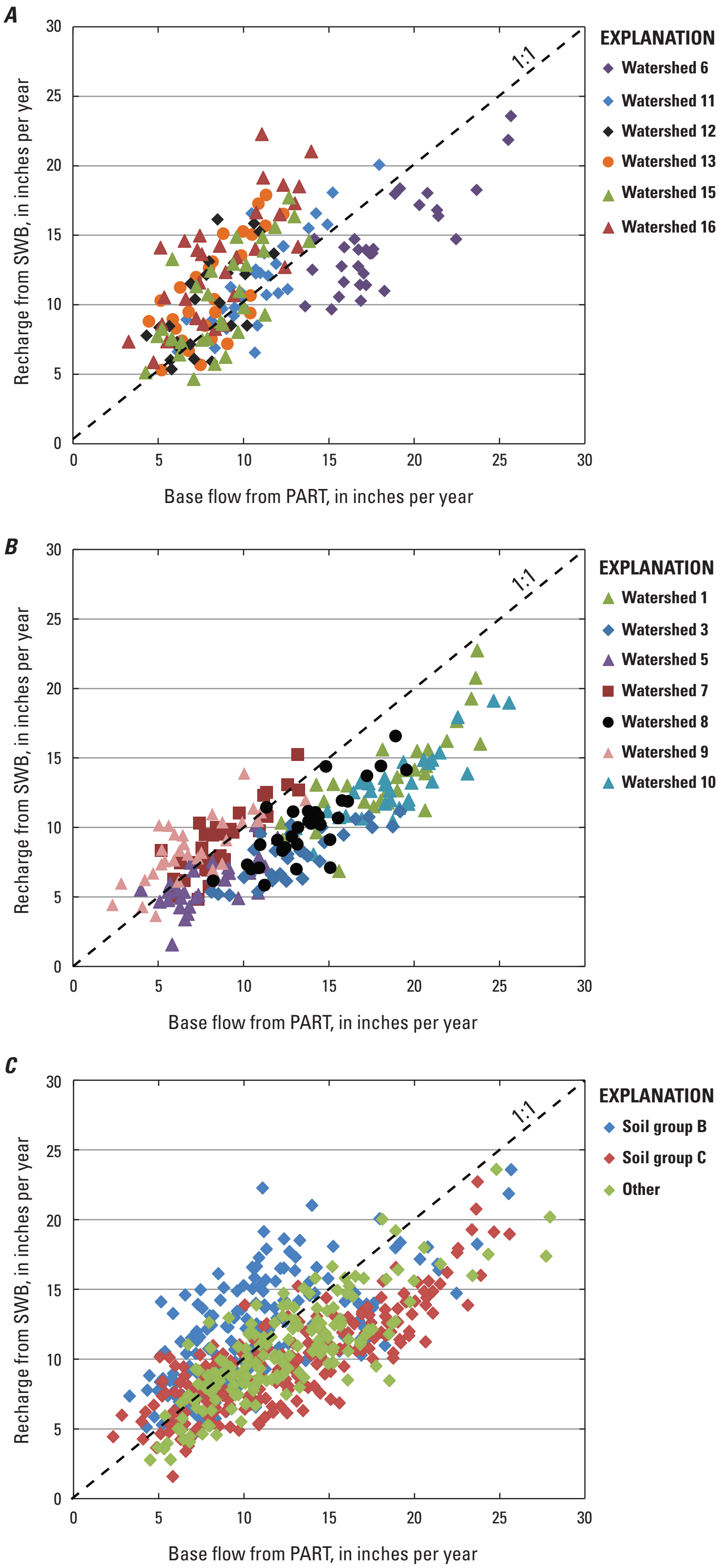


Figure 17. Annual base flow from PART versus Soil-Water-Balance (SWB) model simulated recharge for 20 calibration basins: (*A*) hydrologic soil group B, (*B*) hydrologic soil group C, and (*C*) all hydrologic soil groupings.

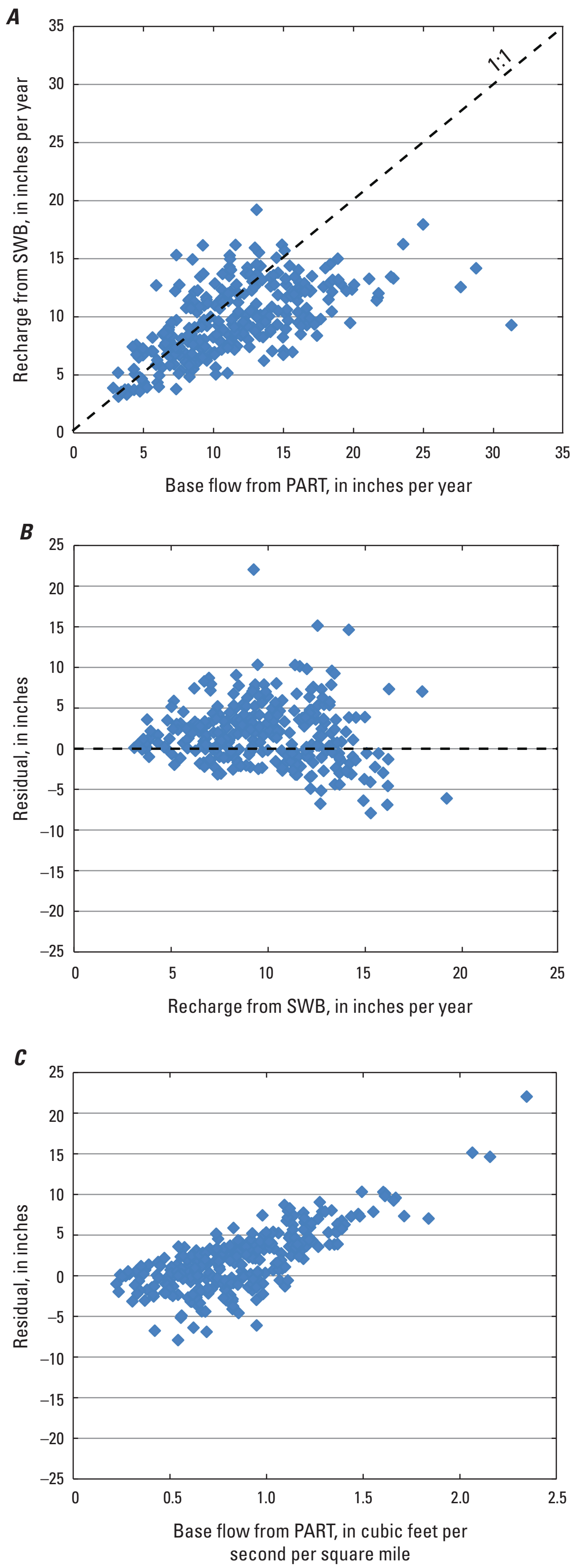


Figure 18. Base flow from PART versus Soil-Water-Balance (SWB) model simulated recharge for 297 basins in the study area with greater than 10 years of record between 1980 and 2011 in the Appalachian Plateaus study area.

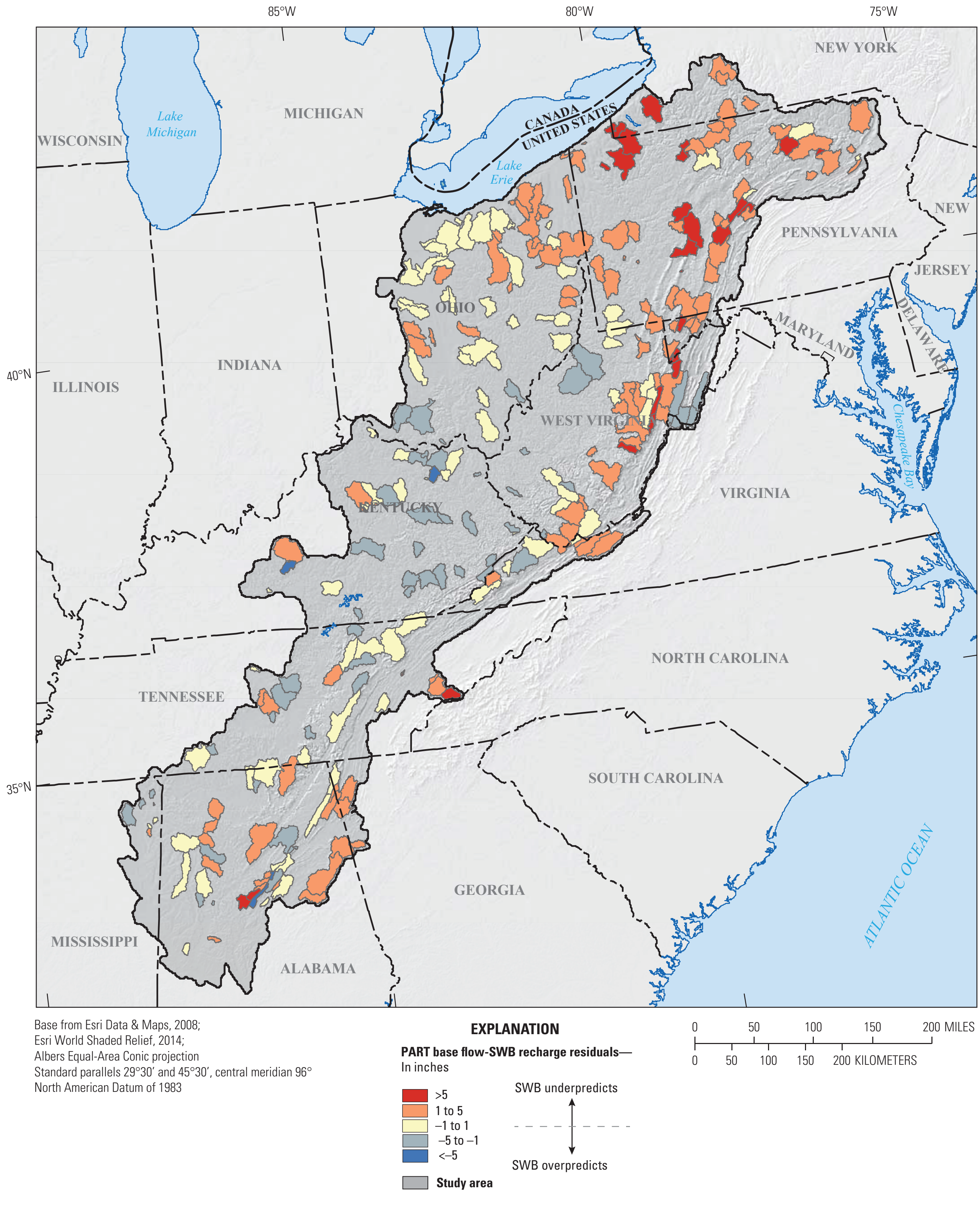


Figure 19. Soil-Water-Balance (SWB) model residuals from PART showing areas of model over and underestimation for 297 basins in the study area.

The SWB model of the Appalachian Plateaus underestimates the highest base flows computed with PART for watersheds with unit base flow values larger than 1.0 cfsm. Many of these watersheds are located at high altitudes near the eastern boundary of the Plateaus with the Allegheny Structural Front. This portion of the Appalachian Plateaus supports many high-yield bedrock wells (Wyrick, 1968) possibly as a result of a high density of fractures that formed in response to folding of the bedrock (Wyrick, 1968). The SWB model represents water movement through the soil profile and not the underlying bedrock. Recharge to bedrock aquifers in the SWB model is limited by the maximum saturated hydraulic conductivity (K_{sat}) specified for each combination of soil group and land cover. The K_{sat} values of soils that overlie permeable bedrock near the Allegheny Structural Front could be larger than the K_{sat} values of similar soils further to the west that overly less permeable bedrock. The difference between SWB recharge and PART base flow near the Allegheny Structural Front could be explored in future runs by specifying larger K values for soils at higher altitudes.

Aquifer Recharge and Base-Flow Discharge

The magnitude and spatial distribution of 1980–2011 mean annual recharge estimates from the SWB model of the Appalachian Plateaus (fig. 20*A*) were compared to previous recharge estimates covering the study area for 1951–1980 (Wolock, 2003a) (fig. 20*B*) and estimates derived from use of PART for 1980–2011 (fig. 20*C*) (Nelms and others, 2015). Imbedded in this methods comparison is the assumption that recharge to Appalachian Plateaus aquifers is equivalent to groundwater discharge as base flow to streams. In the SWB model (fig. 20*A*), this assumption is satisfied once the maximum soil-water capacity in a model cell is reached. At that time, excess water in the soil column, computed as the difference in water-budget sources (precipitation, snowmelt, and runoff from upstream cells) and water-budget sinks (interception, ET, and runoff to downstream cells), conceptually exits the model as recharge to underlying aquifers. The annual sum of recharge in a basin of the SWB model was assumed to be comparable to base-flow estimates from PART (fig. 20*C*). Groundwater recharge estimates in figure 20*B* were extracted from a conterminous U.S. dataset derived by Wolock (2003a). Wolock (2003a) used a hydrograph separation technique (Wahl and Wahl, 1988, 1995) to estimate BFI at 19,589 selected USGS streamflow gaging stations across the United States (Wolock, 2003c). By multiplying the interpolated BFI values (Wolock, 2003b) by 1951–1980 long-term average streamflow values (Gebert and others, 1987), Wolock (2003c) provides recharge estimates for the 30-year period prior to the SWB model period. Figure 20*C* shows recharge estimates from PART for 297 USGS streamflow gaging stations with at least 10 years of record from 1980 to 2011, contemporaneous with the SWB model period (Nelms and others, 2015).

Annual results from the 1980 to 2011 SWB model runs were averaged to create a mean annual recharge distribution for that time period (fig. 20*A*). The SWB model produced a mean annual recharge of about 9 in/yr for the entire study area with a range from about 1 in/yr in central Ohio to 93 in/yr in a single model cell located near the Tennessee/North Carolina border. Rainfall can locally exceed 100 in/yr in mountainous regions of the southeastern United States (fig. 9); however, estimates of bedrock recharge exceeding 35 in/yr should be considered anomalous and inconsistent with the current conceptual knowledge of bedrock recharge to areas with thick colluvial cover (Nelms and Moberg, 2010). High recharge values in excess of 20 in/yr were considered potentially anomalous and observed in less than 2 percent of the model area. Mean annual recharge values for the 1980–2011 SWB model period were generally consistent with typical ranges of 6 to 25 in/yr recharge reported from period-of-record streamflow analyses for various locations within the Appalachian Plateaus (Reese and Risser, 2010; Dumouchelle and Schiefer, 2002; Kozar and Mathes, 2001; Zurawski, 1978). Recharge values less than 10 in/yr compose about 61 percent of the model area, predominantly in low altitude areas north of central Kentucky and west-central Ohio.

The primary natural factors that may affect recharge include, but are not limited to, geology, topography, soils, land-use, and variable climate conditions (Price, 2011). The SWB model does not directly account for geology but does include input of other climatic and landscape datasets that correspond with computed recharge distributions. Values of recharge in excess of 15 in/yr in the northcentral Pennsylvania correspond to locally high precipitation (fig. 9) and hydrologic soil types A and B (fig. 13). Less than 5 in/yr of recharge is received over much of the western, northwestern, and northern parts of the study area in northern Kentucky, Ohio, and New York, respectively, and is associated with low precipitation and hydrologic soil types B/D, C/D, and D; soil types with low to very low infiltration potential. Recharge generally increases to the south with increased precipitation and a transition in hydrologic soil types from those with lower to higher infiltration potential. In Alabama, the juxtaposition of high (>20 in/yr) and low (<5 in/yr) recharge areas corresponds to the variability in soil properties in locations of hydrologic soil types A and D, respectively.

Within the Appalachian Plateaus study area, 1951–1980 annual recharge estimates extracted from Wolock (2003a) range from about 2 to 23 in/yr and average about 7 in/yr. Wolock's (2003a) annual average is 2 in/yr lower than the average computed from SWB result for 1980–2011. The 1980 to 2011 mean annual base-flow estimates from PART for 297 basins range from 3 to 32 in/yr, with a mean value of about 12 in/yr (fig. 20). For the same basins, SWB recharge estimates range from about 3 to 19 in/yr, with a mean value of about 9.7 inches. In general, the recharge estimates derived by Wolock (2003a) using the BFI hydrograph separation method (Wahl and Wahl, 1988, 1995), also produce lower recharge values than those computed using other hydrograph separation

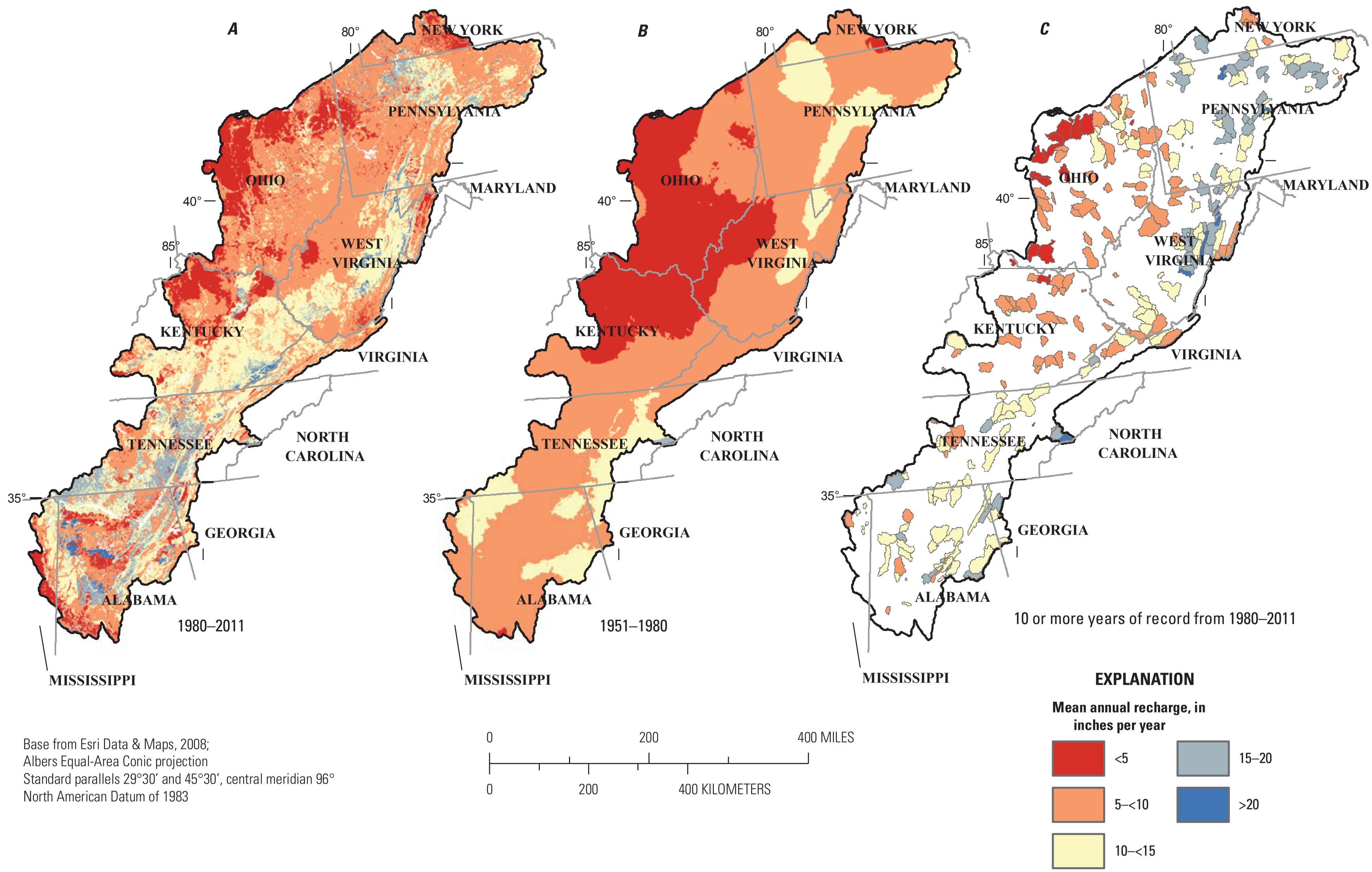


Figure 20. Distribution of mean annual recharge from *A*, the Soil-Water-Balance (SWB) model, 1980–2011, *B*, Wolock (2003a) 1951–80, and *C*, PART, displayed as averages for basins with at least 10 years of record from 1980–2011.

techniques (fig. 21) and a less spatially variable distribution of recharge than that computed by the SWB model (fig. 20*A–B*). The differences between Wolock's (2003a) results and that of SWB or PART could certainly result from variability in drier climate conditions from 1951 to 1980 versus wetter climate conditions from 1980 to 2011 (fig. 10). A comparison of base-flow estimates from six hydrograph separation methods for 849 streamflow gaging stations in the study area (Nelms and others, 2015), however, indicates that the BFI method consistently results in lower annual base-flow estimates than that computed with PART (fig. 21). The differences in results from the two methods tend to become larger with increasing base flow. The comparison indicates high recharge rates in upland areas of Pennsylvania, Maryland, and West Virginia and high recharge rates in rain-soaked areas of Alabama and Tennessee are underestimated by the recharge estimates from Wolock (2003a).

It should be noted that Wolock's (2003a) and Gebert and others' (1987) input datasets were highly generalized, with coarse discretization intended to assess hydrologic conditions at a national scale. Despite the lack of spatial detail in the recharge estimates and potential differences caused by prevailing climate conditions, similar patterns are evident in the results shown in figure 20*A–B*. Recharge estimates below 5 in/yr occur predominantly in central Ohio, central Kentucky, and bordering areas of West Virginia, and along the border of New York and Pennsylvania and in southwestern Alabama (fig. 20*B*). Although SWB recharge estimates below 5 in/yr cover a smaller portion of the model area than those from Wolock (2003a), the lowest estimates derived from both methods were obtained in the same general locations (fig. 20*A–B*). Recharge estimates from Wolock (2003a) only exceeded 15 in/yr in eastern Tennessee and western North Carolina, consistent with the area of highest recharge computed by SWB. PART estimates are higher than SWB estimates throughout much of the model area, with the biggest differences computed in the northern and northeastern parts of the study area in Pennsylvania and West Virginia along the Allegheny Mountain section of the Appalachian Plateaus Physiographic Province (Fenneman and Johnson, 1946).

Evapotranspiration

Evapotranspiration can be derived by subtracting surface-runoff and base flow from the average annual precipitation for the watershed (eqn. 5) and is commonly based on results of hydrograph separation analysis by one of the methods previously described. Other available methods for estimation of ET are not based on streamflow analysis, but rather on remotely-sensed data or climatological data, specifically, precipitation, and air-temperature records. The SWB model produces a spatially variable estimate of potential ET (PET) from daily minimum and maximum air-temperature data based on the Hargreaves-Samani method (Hargreaves and Samani, 1985). For SWB estimates of ET (fig. 22*A*), PET is subtracted from precipitation. If precipitation exceeds PET, then ET equals PET; otherwise, ET is limited to water that can be extracted from the soil (Westenbroek and others, 2010). Long-term estimates of actual ET were computed by Sanford and Selnick (2012) from precipitation and streamflow records at 838 watersheds across the United States. Sanford and Selnick (2012) presented ET estimates as a percentage of precipitation, or ET/P. National maps of ET were developed by through regression of ET/P with national climate and land-cover datasets (fig. 22*B*). Another technique commonly used to estimate evapotranspiration, especially in western States having substantial irrigated acreage, is the simplified surface energy balance (SSEB) method (Maupin and others, 2012). The main modeling principle of the SSEB approach documented by Senay and others (2007, 2011) is the combined use of reference ET (ETo) and land-surface-temperature (LST) data. The Operational Simplified Surface Energy Balance (SSEBop) model is based on the SSEB model, but SSEBop has been simplified so that ET can be estimated from air temperature, land-surface temperature, and reference ET values (fig. 22*C*) (Senay and others, 2013).

In the study area, SWB estimates for 1980–2011 mean annual ET range from about 9 to 47 in/yr, with a mean of about 30.5 in/yr (fig. 22*A*). ET values exceed 35 in/yr predominantly in areas south of West Virginia, where average annual temperatures exceed 54 °F, average annual precipitation exceeds 50 in/yr (fig. 9), and hydrologic soil type B covers a majority of the area (fig. 13). ET decreases to the north where estimates are less than 20 in/yr in less than 1 percent of the model area, predominantly in areas of low precipitation and cool temperatures in Ohio, Pennsylvania, and New York. In the south, where precipitation and temperature are relatively high, the lowest ET values are consistent with low AWC and low-infiltration-potential hydrologic soil types C and D overlying aquifers of Lower Pennsylvanian age in Alabama.

In general, a comparison of long-term ET estimates from all methods noted in figure 22 indicates that ET rates increase from north to south, consistent with precipitation patterns and decrease with altitude gain, and also consistent with temperature patterns (fig. 9). The 1971–2000 annual regression-based estimates of ET from figure 22*B* (Sanford and Selnick, 2012) are the lowest estimates of those presented in figure 22. Within the Appalachian Plateaus study area, the Sanford-Selnick model yielded ET estimates ranging from about 16 to 37 in/yr, with a mean of about 26.5 in/yr for the model area. The SWB model estimates of ET exceed the Sanford-Selnick model estimates by more than 1 inch over approximately 87 percent of the model area and by more than 5 inches over about 47 percent of the model area. ET estimates from the SSEBop model represent mean annual ET estimates for 2000–11 (fig. 22*C*). The SSEBop model yielded ET estimates ranging from approximately 10 to 50 in/yr, with a mean of about 30 in/yr for the model area. The lowest values of ET correspond to developed areas (fig. 22*C*), probably representing a reduction in ET with increased impervious area and high land-surface temperatures. Streamflow is less sensitive to temperature

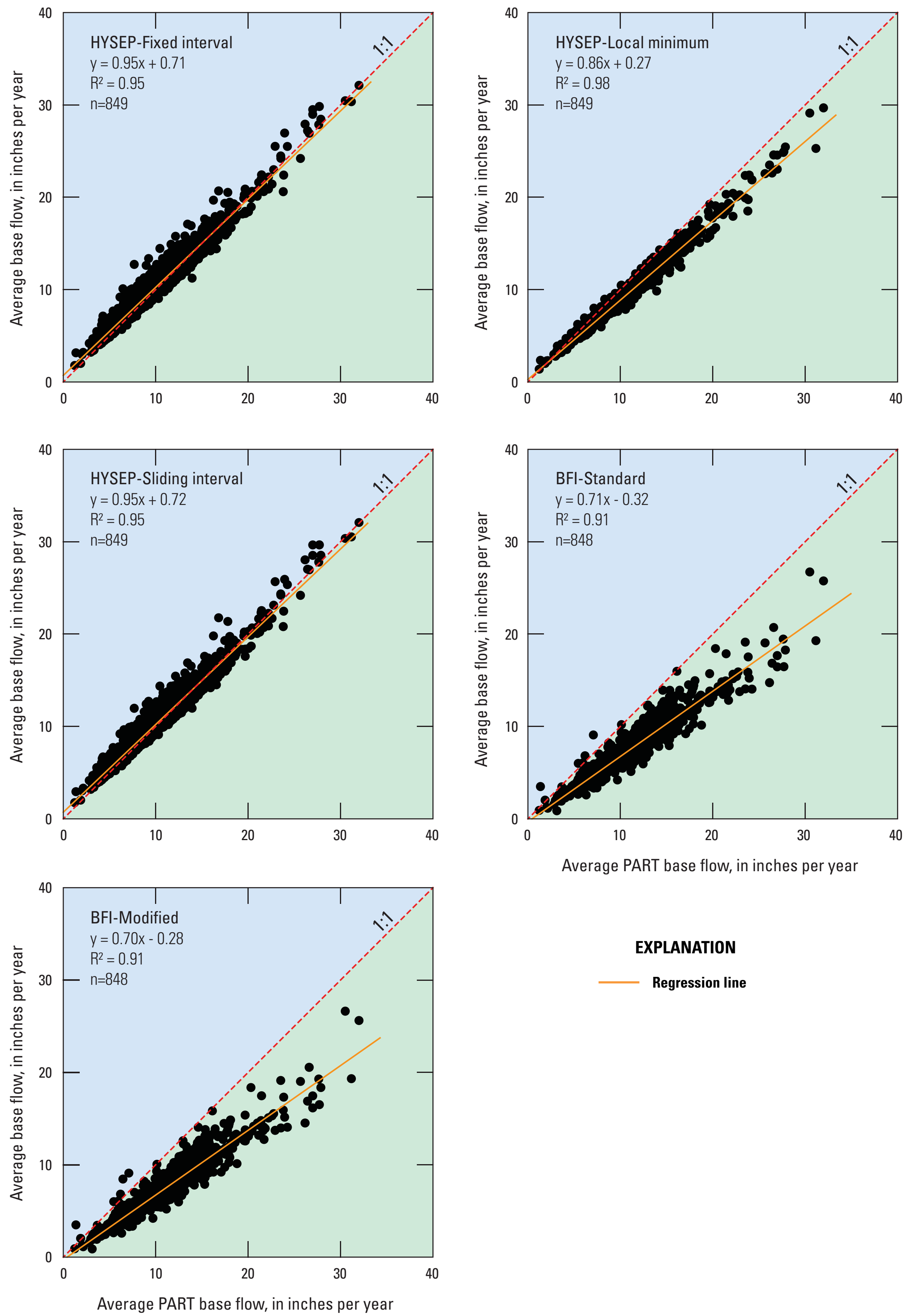


Figure 21. Comparison of average base-flow values estimated from PART hydrograph separation method to average base-flow values estimated from HYSEP and Base Flow Index (BFI) methods for streamflow gaging stations in the Appalachian Plateaus region. Regression lines are in orange.

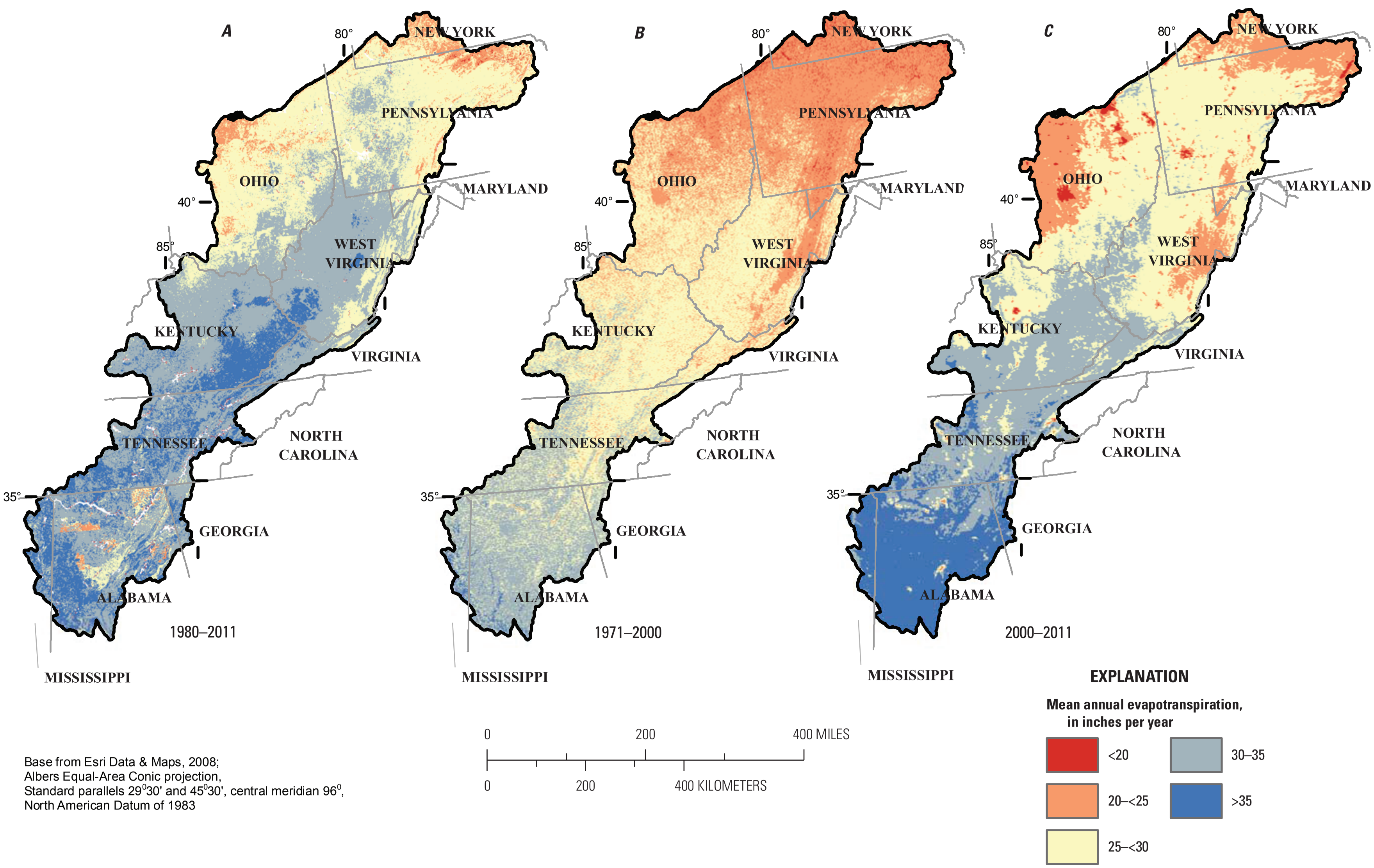


Figure 22. Distribution of mean annual evapotranspiration (ET) from *A*, the Soil-Water-Balance (SWB) model, 1980–2011, *B*, Sanford and Selnick (2012) 1971–2000, and *C*, Operational Simplified Surface Energy Balance (SSEBop) model 2000–11 (Senay and others, 2013).

changes in urban centers (Dewalle and others, 2000), where large areas of impervious surface coupled with the widespread loss of vegetation have presumably reduced the plant transpiration component of ET. SWB-computed ET estimates exceed SSEBop estimates by more than 1 inch over about 60 percent of the model area. In the northern part of the model area, SWB estimates exceed SSEBop estimates in developed areas and areas at relatively high altitudes with high AWC. SSEBop ET estimates exceed SWB estimates in areas at relatively low altitudes with low AWC. SSEBop estimates exceed SWB estimates predominately in the southern part of the model area, where SSEBop ET estimates are highest and the model area is underlain by soils with low AWC (fig. 14) and low to very low infiltration potential (fig. 13). All three methods shown in figure 22 yield results slightly higher than the previously reported values of 21 in/yr in Pennsylvania and West Virginia (Mohamoud, 2004) to 29 in/yr in the Tennessee River Valley (Zurawski, 1978).

Water Withdrawals and Water Use

Groundwater withdrawn from the Appalachian Plateaus aquifer system is an important source of freshwater for drinking water, commercial, industrial, mining, and agricultural, and other uses. Estimates of water withdrawals were synthesized from Kenny and others (2009) for a total of 186 counties that have either their center or more than 50 percent of their area underlain by Appalachian Plateaus aquifers (fig. 23). At the time of analysis, water-use data describing withdrawals through 2010 (Maupin and others, 2014) were not available. In 2005, fresh groundwater withdrawn from aquifers in those counties totaled 639 Mgal/d and water withdrawn from streams totaled about 17,400 Mgal/d (Kenny and others, 2009). The distribution of groundwater withdrawals by county during 2005 (fig. 23) indicates that Ohio withdrew the largest quantity of groundwater, almost 50 percent of the 639 Mgal/d throughout the region. It is likely that a majority of the Ohio groundwater withdrawals were from alluvial and outwash aquifers overlying Appalachian Plateaus aquifers. Surface-water use, however, greatly exceeded groundwater use in all States within the Appalachian Plateaus region (fig. 24).

The total drinking-water use for approximately 13.6 million people living in the Appalachian Plateaus in 2005 was about 1,930 Mgal/d, about 23 percent of which (or about 449 Mgal/d) was derived from public and domestic groundwater sources in the underlying aquifer system (tables 6, 7; Kenny and others, 2009). For 5.94 million people, groundwater withdrawn for drinking water accounted for 70 percent of all the groundwater withdrawals in counties underlain by Appalachian Plateaus aquifers (fig. 23, table 7). The majority (71 percent) of these groundwater withdrawals occurred in Ohio and Pennsylvania (41 counties in Ohio; 26 counties in Pennsylvania). Of the 449 Mgal/d of groundwater withdrawn for drinking water, 64 percent was from public-supply facilities, and 36 percent was from self-supplied domestic sources (table 6).

Groundwater withdrawals for industrial and thermoelectric purposes composed approximately 17 percent (110 Mgal/d) of the total groundwater withdrawn in 2005, and Ohio and West Virginia accounted for about 85 percent of this type of use (table 7, fig. 25). Thermoelectric use from groundwater sources occurred mostly in Ohio and composed less than 1 percent (9 Mgal/d) of the total groundwater withdrawn. The agricultural use category included irrigation, livestock, and aquaculture uses, as defined by Kenny and others (2009). In 2005, the estimated agricultural use of groundwater composed less than 10 percent (about 54 Mgal/d) of the total groundwater withdrawn. Pennsylvania and Alabama together used about 55 percent of the 54 Mgal/d withdrawn from aquifers for agricultural use (table 7, fig. 25). Dewatering was not reported as a mining withdrawal unless the water was used for other purposes prior to being discharged. Ground-water withdrawals for mining in 2005 were estimated to be about 26 Mgal/d. Ohio and Pennsylvania mining withdrawals from the Appalachian Plateaus aquifers accounted for 14 of the 26 Mgal/d.

Smith and others (2011) present a series of equations to calculate the disposition of water withdrawn for various economic sectors directly from the water-use data published by Kenny and others (2009). Following the lifecycle of water from withdrawal to disposition is common in heavily irrigated agricultural areas of the United States, where return flows are important contributor to groundwater budgets (Stanton and others, 2011; Faunt, 2009). In the Appalachian Plateaus, water withdrawals for domestic use eventually return to aquifers through septic drainfields, but most of the water withdrawn discharges to surface waters through centralized sewer systems, becomes surface runoff, evaporates, or is consumed (fig. 26). About 471 Mgal/d, or 74 percent, of the water withdrawn from Appalachian Plateaus aquifers is either returned to the hydrologic budget in the form of wastewater or unconsumed water that is discharged to streams and rivers (fig. 26). The remaining 26 percent is assumed to be removed from the system through consumption or evaporation, 131 Mgal/d of which is attributed to losses from public supply, domestic, and industrial/commercial use categories. Livestock and irrigation (including golf courses) are the only two uses that consume or evaporate more water than is returned as surface-water discharge.

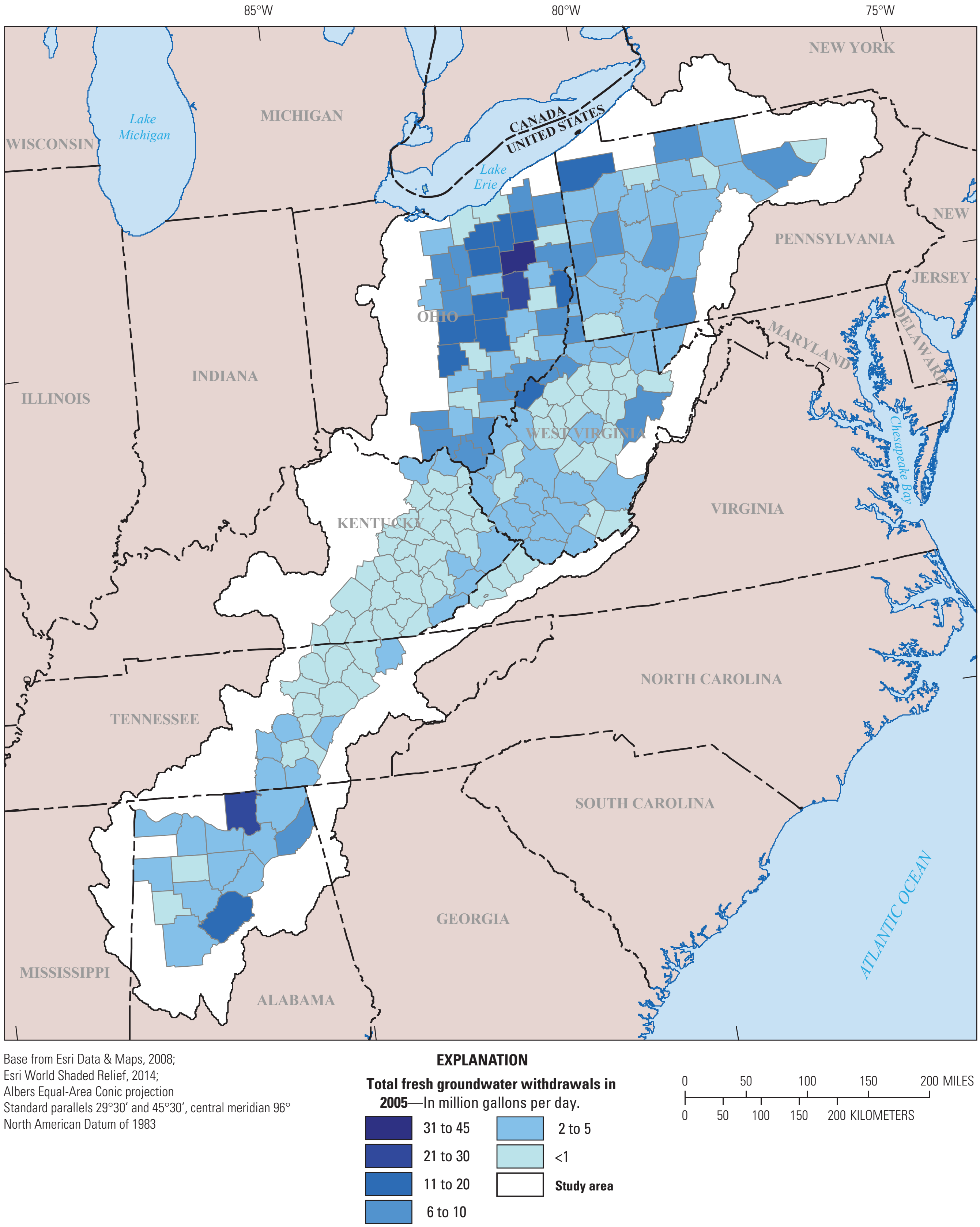


Figure 23. Total fresh groundwater withdrawals in 2005 from selected counties within the Appalachian Plateaus region.

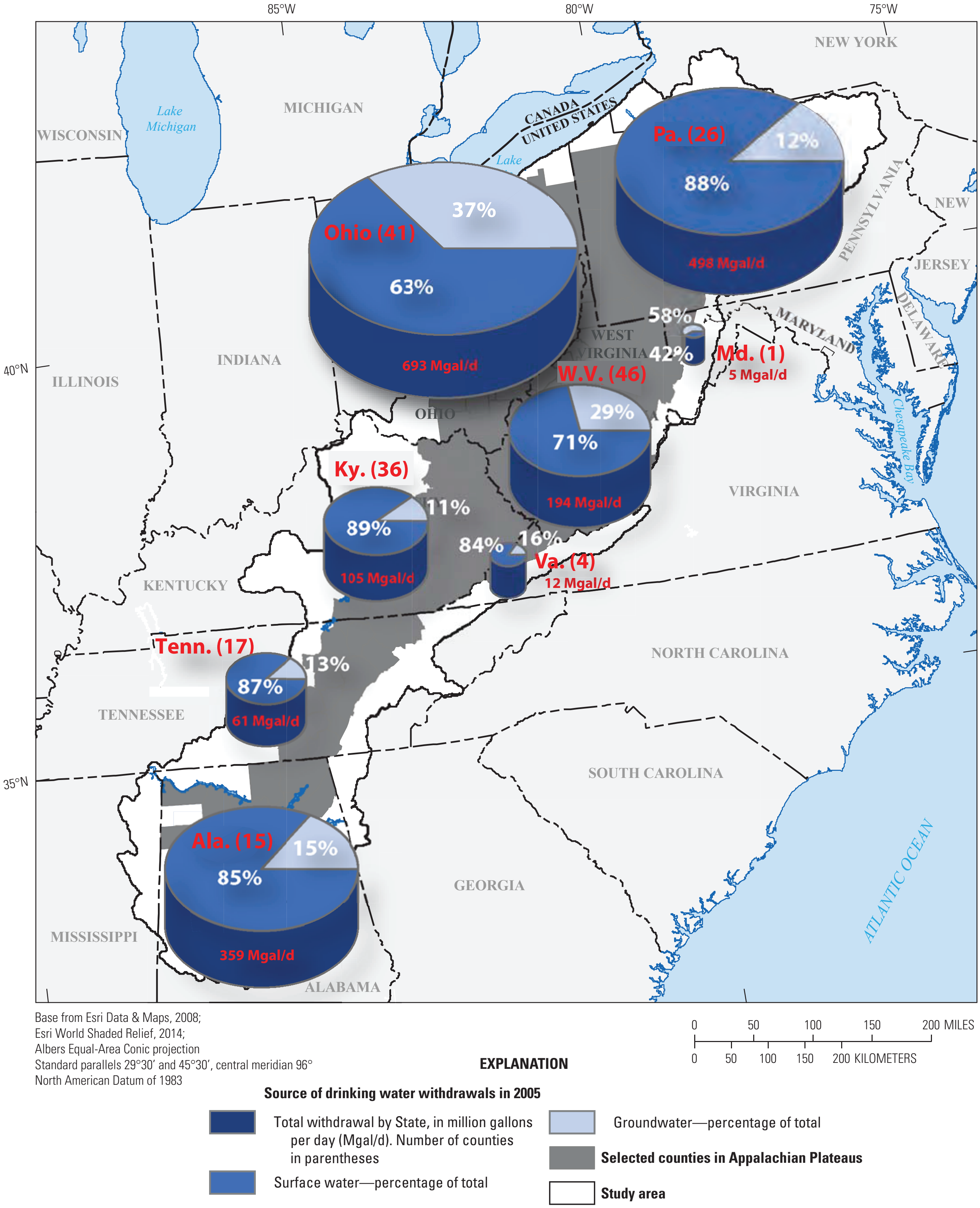


Figure 24. Total surface and groundwater drinking-water withdrawals in 2005 from selected counties within the Appalachian Plateaus region.

Table 6. Distribution of drinking water sources by State for 186 counties within the Appalachian Plateaus study area.

[Values are in million gallons per day. May not add to totals shown because of independent rounding. Water use data are from Kenny and others (2009). Gray shading indicates that value was rounded to zero]

State	Public groundwater source	Public surface-water source	Domestic groundwater	Total
Alabama	43	306	10	359
Kentucky	3	94	8	105
Maryland	1	2	2	5
Ohio	174	438	80	693
Pennsylvania	29	436	33	498
Tennessee	5	54	3	61
Virginia	0	10	2	12
West Virginia	30	139	25	194
Total	286	1,479	163	1,928

Table 7. Distribution of groundwater use by State for 186 counties within the Appalachian Plateaus study area.

[Values are in million gallons per day. May not add to totals shown because of independent rounding. Water use data are from Kenny and others (2009). Gray shading indicates that value was rounded to zero]

State	Public and domestic supply	Agricultural	Industrial and thermoelectric	Mining	Total
Alabama	53	11	4	4	72
Kentucky	12	1	1	3	16
Maryland	3	0	0	0	3
Ohio	254	9	45	8	317
Pennsylvania	62	19	11	6	98
Tennessee	8	6	1	1	16
Virginia	2	0	0	0	2
West Virginia	56	8	48	4	115
Total	449	54	110	26	639

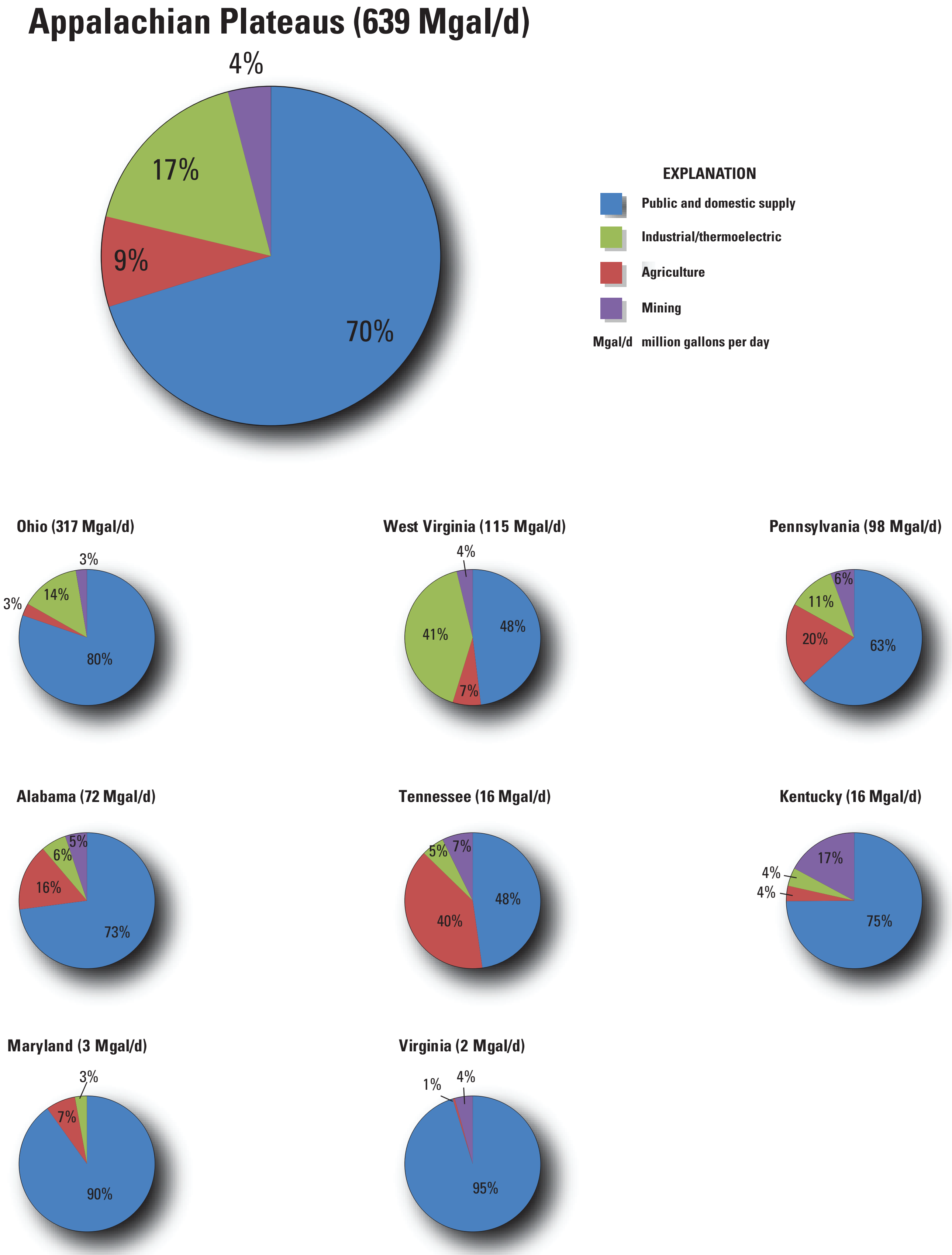
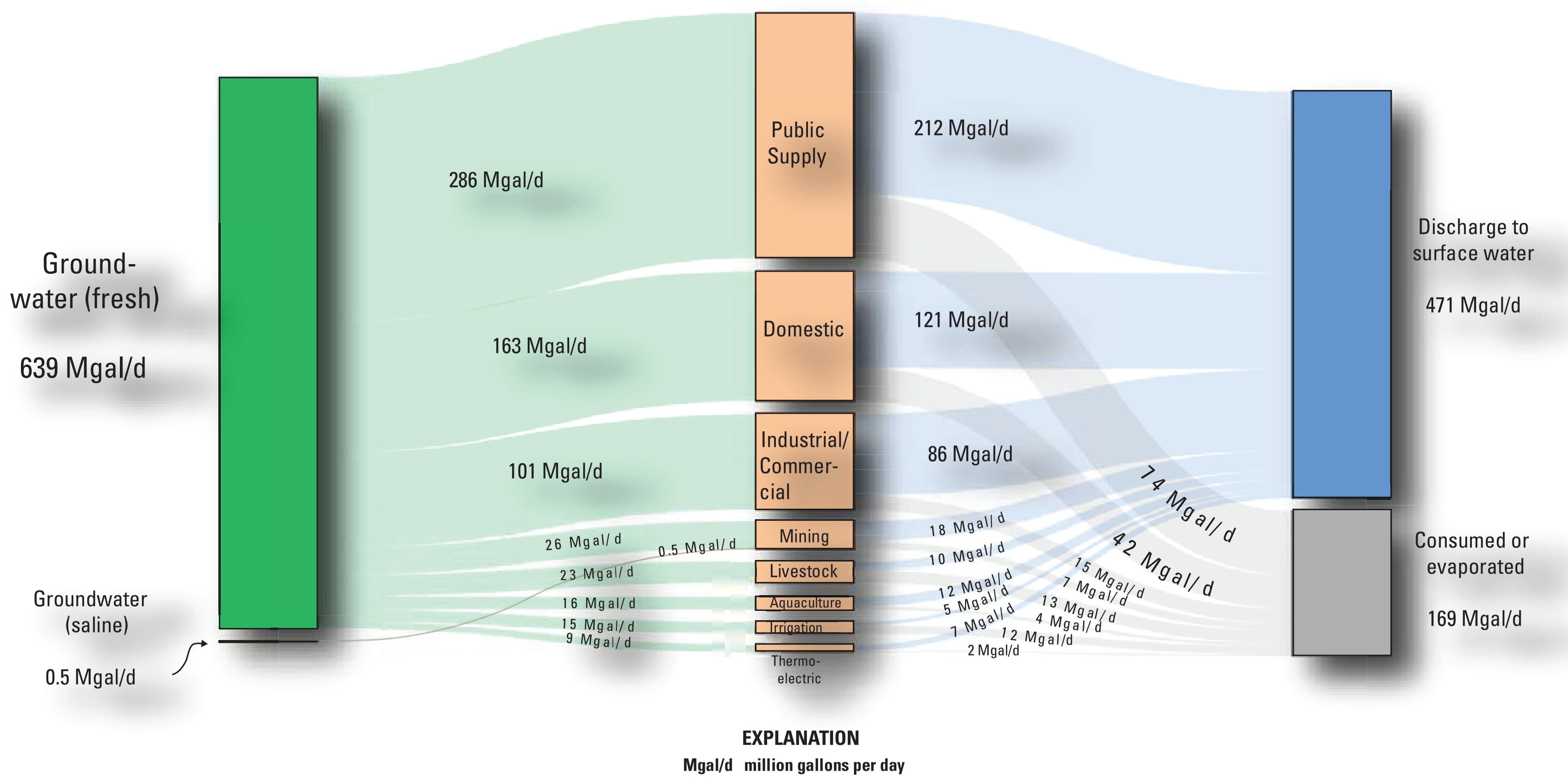


Figure 25. Groundwater withdrawals by State and percentages by water-use category in 2005 from selected counties within the Appalachian Plateaus region.



Note: Numbers may not sum to equivalent totals because of independent rounding.

Figure 26. Groundwater-use cycle in the Appalachian Plateaus (data from Kenny and others, 2009). Further details about flow calculations are provided in Smith and others (2011).

Groundwater in Storage

Appalachian Plateaus aquifers respond to precipitation over relatively short timeframes (fig. 4). Variations in water levels of even a few feet are the result of losses or gains in aquifer storage that can be substantial over periods of months to a year. With the exception of flooding in abandoned coal-mine aquifers, water levels in Appalachian Plateaus aquifers tend to vary about a well-defined range (fig. 4). Over increasingly long periods, changes in aquifer storage (indicated by water-level fluctuations) in the Appalachian Plateaus approach zero as annual losses and gains caused by prevailing climate conditions are balanced.

Abandoned Coal-Mine Aquifers

Surface and underground mining for coal in Permian- and Pennsylvanian-age rocks often causes substantial changes in local hydrogeology. The distribution of individual components of the water budget varies greatly between mined and unmined watersheds (Zegre and others, 2014; Messinger and Paybins, 2003; Larson and Powell, 1986). Local changes in the hydrologic budget of mined watersheds in the Appalachian Plateaus result from post-mining increases in base flow as deep coal mines or spoil material fill and then discharge to surface waters through springs (Borchers and others, 1991; Larson and Powell, 1986). Surface spoil materials increase the storativity in coal-mined watersheds (Cunningham and Jones, 1990), which tends to reduce peak flows and flatten streamflow recession curves (Larson and Powell, 1986).

Historically, underground coal mining has shown the greatest potential to impact the hydrology of aquifers in the Appalachian Plateaus on a relatively large scale (Callaghan and others, 1998). Coal mining can temporarily lower the water table and change the direction of flow across former groundwater divides (Trapp and Horn, 1997) and, in some cases, result in permanent interbasin transfer of water (Kozar and McCoy, 2012). The lateral continuity of deep coal-mine voids on a regional scale contributes to the formation of mine pools, which are vast areas of underground workings that may be hydrologically connected over tens of miles (Donovan and Fletcher, 1999). Interconnection between coal mines on a regional scale leads to groundwater capture areas that greatly exceed the boundaries of individual mines (Lopez and Stoertz, 2001; Donovan and Fletcher, 1999). Coal mine voids in the subsurface substantially change the spatial distribution of recharge (McCoy, 2002) and have enhanced the limited aquifer storage available prior to the development of mine aquifers (Trapp and Horn, 1997). Large volumes of water are known to be stored in underground coal mines in Kentucky, Ohio, Pennsylvania, West Virginia, and Virginia (Stoertz and others, 2004; Donovan and Fletcher, 1999; Minns, 1993; Ferrell, 1992; Hobba, 1981). An estimated 1,390,000 Mgal of water are stored in the Upper Pennsylvanian Pittsburgh coal mine aquifers of Ohio, Pennsylvania, and West Virginia (Donovan and Fletcher, 1999) and all other known abandoned underground coal mine aquifers in West Virginia (McColloch, 2012) (as shown in red in fig. 27). Fully saturated Pittsburgh coal mine aquifers in Upper Pennsylvanian rocks of northern West Virginia and southwestern Pennsylvania form the largest spatially continuous high-yield aquifers exclusive of Cambrian-Ordovician karst aquifers in the Northern Appalachian region (Donovan and Leavitt, 2004). The quality of water from coal-mine aquifers ranges from high quality, potable (Kozar and others, 2012) and beneficial for aquaculture and recreation (Miller, 2008) to poor quality acid-mine drainage.

Alluvial and Outwash Aquifers of the Ohio River Basin

Unconsolidated deposits of glacial and alluvial origin lie at or near the surface in northern portions of Ohio and Pennsylvania and along many of the major river valleys, particularly that of the Ohio River (fig. 28). These valley-fill sediments form alluvial and outwash aquifers that consist of Pleistocene-age gravels, sand, silts, and clays deposited in broad, deep valleys during periods of glacial advance and retreat (Rosenshein, 1988). Alluvial and outwash aquifers are underlain and laterally bounded along uplands by Appalachian Plateaus aquifers of Pennsylvanian and Mississippian age. Groundwater generally flows from uplands toward valleys (Seyoum and Eckstein, 2014; Williams and others, 1998) and Appalachian Plateaus aquifers discharge to alluvial and outwash aquifers rather than directly to streams (Buckwalter and Moore, 2007). Subsurface discharge from Plateaus aquifers in the study area has been estimated to be 39 percent of the total inflow to a buried-valley aquifer in Ohio (Seyoum and Eckstein, 2014). Outside the study area, the range of discharge to alluvial aquifers has been estimated to range from 0.2 Mgal/d per mile (Unthank, 2013) to 1 Mgal/d per mile of surrounding upland (Williams and others, 1998).

About 4,915,000 Mgal of water is stored in alluvial and outwash aquifers in the study area (Bloyd, 1974) (table 8). The amount of water stored in tributary valleys of the Ohio River is generally higher in glaciated portions of the study area than in unglaciated portions. Values ranges from 1,650,000 Mgal in the Muskingum River Valley to 45,000 Mgal in the Kanawha-Little Kanawha River Valleys. With the exception of areas where acid-mine drainage is prevalent, most of the alluvial and outwash aquifers in the study area produce potable water (Bloyd, 1974).

Predevelopment and Postdevelopment Water Budgets

It is assumed that precipitation that falls directly on Appalachian Plateaus aquifers is the sole source of inflow to the groundwater system and that only a relatively small, insignificant amount of groundwater flows into the system from outside the surface drainage area. Daymet data indicate the 1980–2011 average annual precipitation rate for the entire

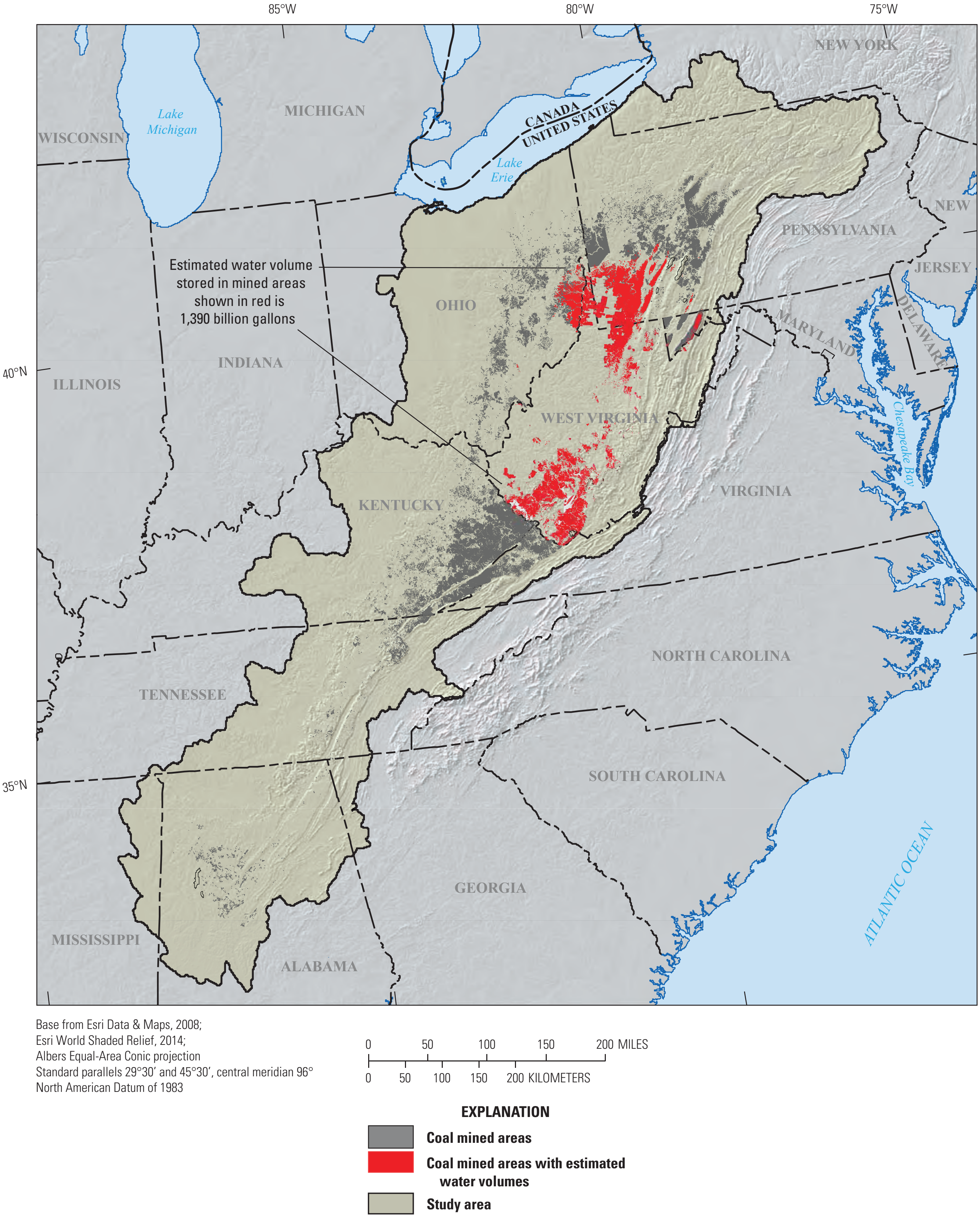
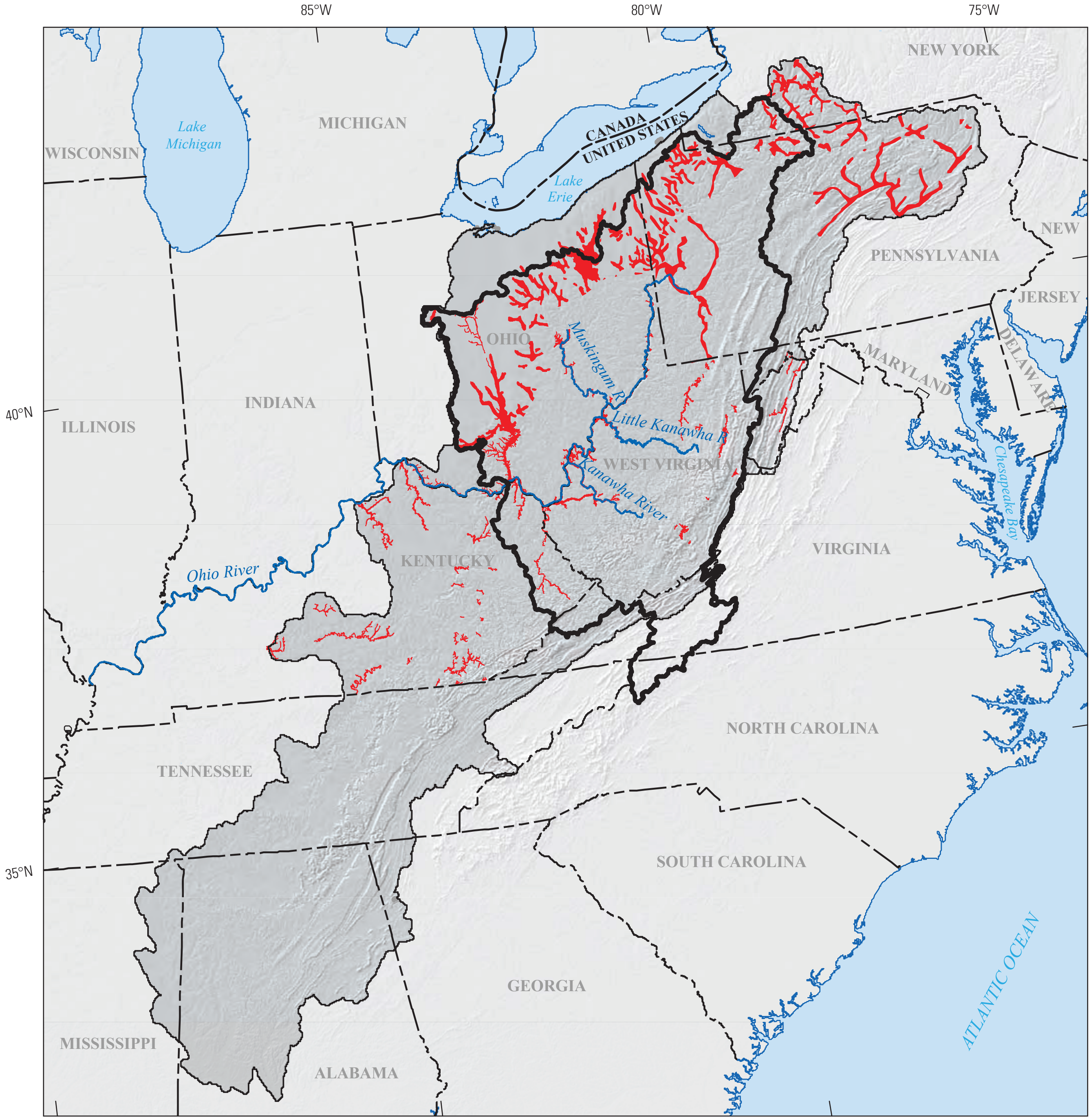


Figure 27. Extent of coal mining in the Appalachian Plateaus and abandoned mine areas with estimated water volumes in storage.



Base from Esri Data & Maps, 2008;
Esri World Shaded Relief, 2014;
Albers Equal-Area Conic projection
Standard parallels 29°30′ and 45°30′, central meridian 96°
North American Datum of 1983

0 50 100 150 200 MILES
0 50 100 150 200 KILOMETERS

Aquifer data from Nicholson and others (2005); USGS (2003); Ohio Department of Natural Resources (2000); Trapp and Horn (1997),

EXPLANATION

- **Sand and gravel aquifers**
- **Study area**
- **Portion of Ohio River Basin draining Permian, Pennsylvanian, or Mississippian aquifers of The Appalachian Plateaus**

Figure 28. Alluvial and outwash aquifers in the Appalachian Plateaus study area.

Table 8. Estimated potable water available from storage in outwash and alluvial aquifers of the Ohio River Basin in the Appalachian Plateaus study area (from Bloyd, 1974).

Basin name	Stored groundwater (billions of gallons)
Ohio River[a]	401
Allegheny River	1,270
Monogahela River	60
Upper Ohio River	576
Muskingum River	1,650
Kanawha-Little Kanawha Rivers	45
Scioto River	748
Big and Little Sandy-Guyandotte Rivers	165
Total	4,915

[a]Segment of Ohio River between Pittsburgh, Pennsylvania, and Portsmouth, Kentucky.

Appalachian Plateaus aquifer system was 216,200 Mgal/d (fig. 29). This time period nearly covers the full range of observed climate conditions with respect to the historic PDSI record but includes more wet years than dry years (fig. 10). Output from the SWB model indicate precipitation falling on the Plateaus is ultimately consumed by evapotranspiration (133,300 Mgal/d or 62 percent of outflow), groundwater recharge and discharge to streams (41,100 Mgal/d or 19 percent of outflow), and surface runoff to streams (41,800 Mgal/d or 19 percent of outflow) (fig. 29). Generalized predevelopment and postdevelopment water budgets presented in figure 29 indicate that most of the precipitation is returned to the atmosphere as ET. Implicit in the water budgets shown in figure 29 are the assumptions that the only differences in predevelopment and postdevelopment conditions are groundwater withdrawals by pumping and that annual climate fluctuations have produced no net change in storage over long time periods. Postdevelopment pumping withdrawals from aquifers are minimal (less than 1 percent) and are returned to the system as wastewater discharge to streams, septic effluent, or irrigation return flows.

Changes in storage attributed to either the construction of surface-water impoundments or the mining of coal can have substantial local influence (Zegre and others, 2014; Kozar and McCoy, 2012) but probably have limited influence on long-term, regional water-budget fluxes of the Appalachian Plateaus, as shown in figure 29. Prior to coal mining, fracture porosity in typical sub-bituminous coal seams is about 5 percent (Rodrigues and Lemos de Sousa, 2002). Assuming the known 1,390,000 Mgal of stored water in mine aquifers (fig. 27) occupies a post-mining porosity of 80 percent (McCoy, 2002), and confined storage in mines is negligible (Donovan and Fletcher, 1999), the additional post-development water in storage attributed to the increase in mine-void related porosity from 5 to 80 percent is 1,303,000 Mgal. Put in context of an annual budget, the increase in coal-mine aquifer storage from pre- to post-development periods is less than that evapotranspired from the Appalachian Plateaus in 10 days.

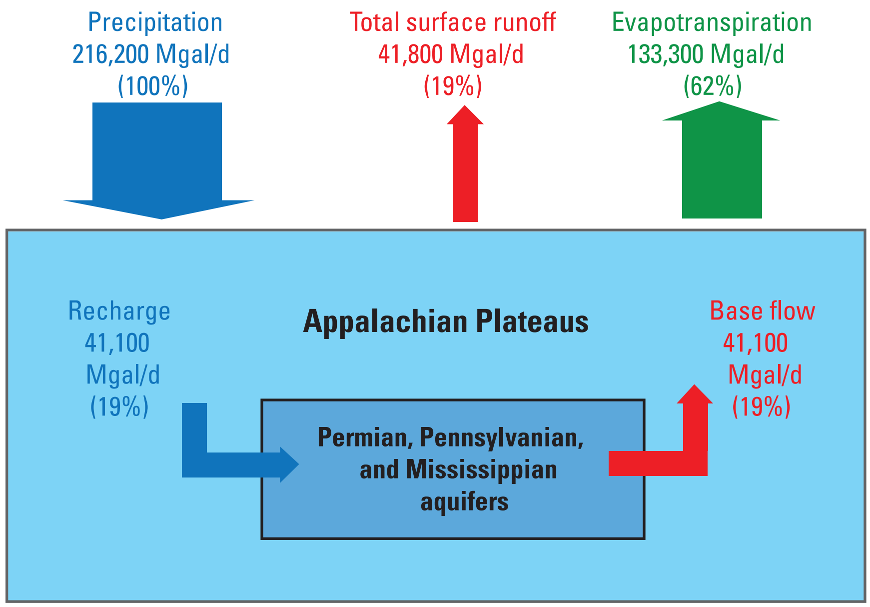
The correlation between precipitation and annual rates of groundwater recharge is strong (Risser and others, 2008), and annual water budgets in the Appalachian Plateaus are highly dependent on prevailing climate conditions (fig. 30). The percentage of annual precipitation that reaches the water table varies based on local conditions and precipitation totals, and ranged from 13 percent (2000) to 22 percent (2011) annually in the SWB model. Uncertainty associated with SWB standard error in recharge residuals of 3.7 to 4.2 in/yr equates to a maximum 3- to 5-percent difference in the percentage of annual precipitation estimated to result in base flow. Estimated direct runoff to streams for the period of 1980–2011 was 17 to 26 percent of annual precipitation and is consistent with previous estimates of 17 to 20 percent of annual precipitation in southern portions of the Appalachian Plateaus (Zurawski, 1978). The dominant form of outflow in the water budget is ET, which composes 52 to 69 percent of annual precipitation. The results in figure 30 indicate changes in the annual water budget are primarily accounted for by the response of ET to various climate conditions. Generally, ET as a percentage of precipitation peaks during dry periods, while base flow and runoff tend to reach minimum values. During wet periods, this relationship is reversed and base flow and runoff as a percentage of precipitation generally peak while ET percentages reach minimum values.

Hydrologic Conditions

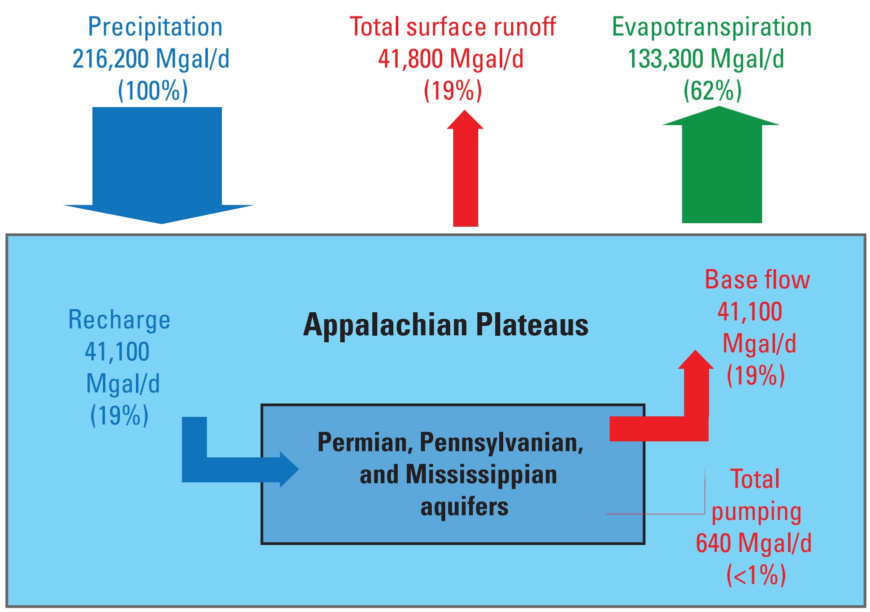
Changes in hydrologic budgets in humid regions of the United States are attributed to many natural factors, principally geology, topography, soils, vegetation, and climate (Price, 2011; Risser and others, 2008; Dumouchelle and Schiefer, 2002). Non-natural, or human-induced changes to the hydrologic budget of the Appalachian Plateaus study area in particular can result from various land-use practices, such as urbanization, deforestation, and coal mining (Zegre and others, 2014; Swank and others, 2001; Dewalle and others, 2000), as well as water withdrawals and interbasin transfers of water (Kozar and McCoy, 2012). Other socioeconomic factors, soil properties, and vegetative succession processes that are transient in nature may also influence the hydrologic budgets; limited data, however, make it difficult to quantify the magnitude of such influences (Jones and others, 2012; Holman, 2006). Decadal hydrologic responses to vegetative succession causing shifts in periods of maximum ET/P, for example, can be mistakenly attributed to climate-change trends (Jones and others, 2012).

Long-term climate records of precipitation, temperature, and hydrologic records of streamflow are the most prevalent and accessible data for determining trends and variability in hydrologic budget components of the Appalachian Plateaus. Climate variability in particular can affect the quantity of

A. **Predevelopment landscape water budget**



B. **Developed landscape water budget**[1]



[1]Negligible increase of 1,390,000 Mgal of water in storage caused by development of coal-mine aquifers not shown.

EXPLANATION

Mgal/d million gallons per day

Figure 29. Predevelopment and postdevelopment water budgets for the Appalachian Plateaus.

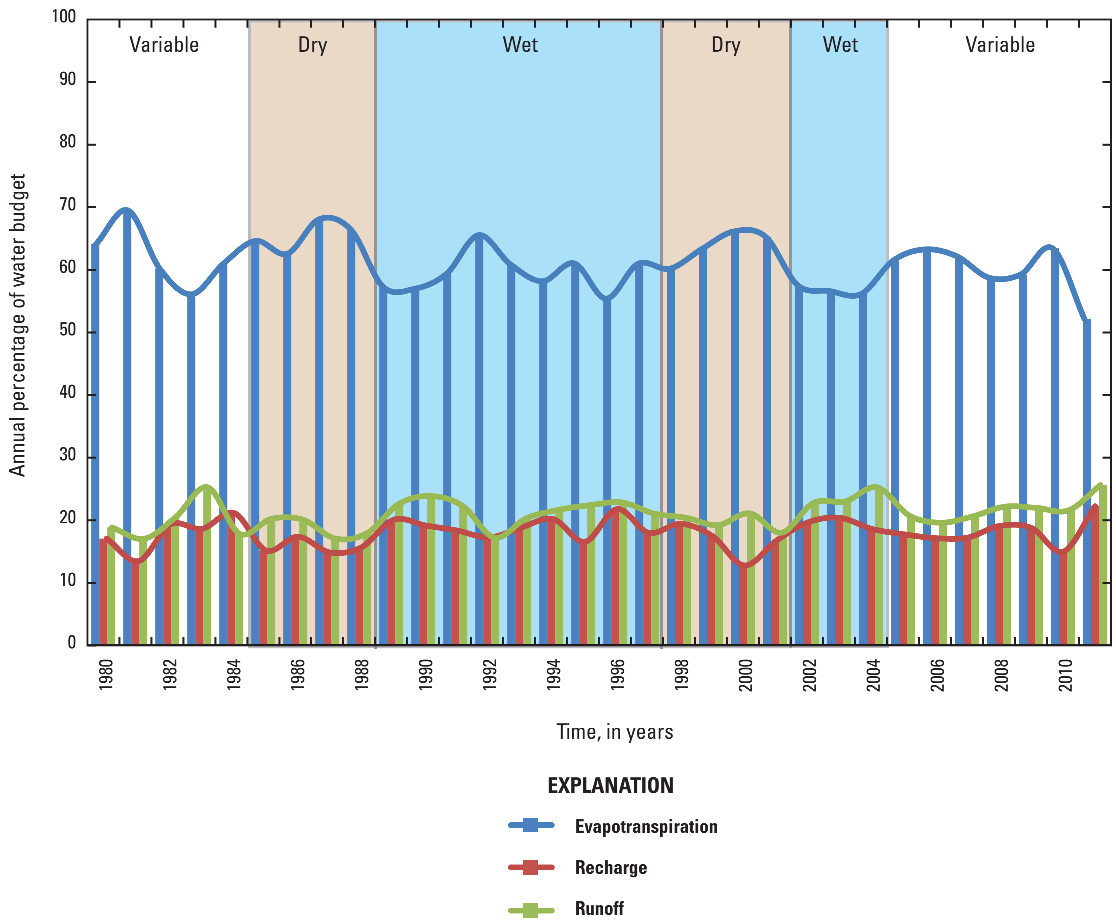


Figure 30. Percent changes in water-budget components in the Appalachian Plateaus study area between years 1980 and 2011.

various components of the hydrologic cycle (Gurdak and others, 2009) and is a major issue potentially limiting water availability in the study area. To address the issue, output from SWB is used to assess differences in water-budget components during mean average (1980–2011), dry (1988), and wet (2004) climate conditions in the Appalachian Plateaus study area (fig. 10). Climate conditions were consistently dry or wet, respectively, for at least 2 years prior to both the 1988 dry and 2004 wet years. An analysis of period-of-record hydrograph separation data provided historical perspective about the magnitude and overall direction of historical and current trends in the components of streamflow in the study area.

Climate-Driven Variability in Water Budgets

Long-term mean annual estimates of recharge magnitudes (fig. 31*A*, reproduced from figure 12*A*) and percentages of precipitation (recharge/P) (fig. 31*D*) for 1980 to 2011 derived from the SWB model were compared to annual estimates for dry climate conditions in 1988 (fig. 31*B*, *E*) and wet climate conditions in 2004 (fig. 31*C*, *F*) identified in figure 10. Recharge magnitudes and percentages of precipitation (recharge/P) generally increase moving from dry conditions to wet conditions, although recharge/P varies less than the magnitude of recharge within this range of climate conditions (fig. 31*A*–*F*). Mean recharge for the model area during dry conditions in 1988 is about 6 in/yr (fig. 31*B*), which is about 3 in/yr less than the mean annual average and about 15 percent of the 1988 total precipitation (fig. 31*E*). Mean recharge for the model area during wet conditions in 2004 is about 11 in/yr, which is about 2 in/yr higher than mean annual conditions. The mean recharge value represents 18 percent of the 2004 precipitation, roughly equivalent to the percentage of 1980–2011 mean annual average recharge/P (19 percent). Mean annual conditions from the SWB model are biased toward wetter conditions (fig. 10) because of the simulation period; however, the similarity between recharge/P in figure 31*D*, *F* indicates that annual recharge in the Appalachian Plateaus reaches a maximum of nearly 20 percent of annual precipitation, regardless of the severity of wet conditions. The percentage of precipitation that reaches the groundwater table tends to vary much less (15 to 19 percent) between dry and wet periods than does the actual magnitude of annual recharge, which varies by nearly a factor of two (6 to 11 in/yr). Both the magnitude of annual recharge and recharge/P are positively related to annual precipitation, consistent with trends observed in the adjacent Valley and Ridge Physiographic Province (Nelms and Moberg, 2010).

The water budget in the Appalachian Plateaus is not only influenced by precipitation patterns, but also by the energy budget as interpreted by the consistent north-to-south variation in ET (fig. 32*A–F*) and temperature (fig. 9). Warmer air holds more water vapor than colder air, as reflected in the enhanced ET in Alabama, Kentucky, and Tennessee relative to Maryland, Ohio, Pennsylvania, and northern West Virginia (fig. 32*A–C*). Shifts toward warmer temperatures in the southeastern United States (Patterson and others, 2012) and northeastern United States (Huntington and others, 2009) have resulted in projected changes to the water budget. Hayhoe and others (2007) project an increase in ET in the northeastern United States in response to climate trends—a relation assumed to include all regions of the United States, including the Appalachian Plateaus, where ET rates are related to ecosystem demands for water (Jones and others, 2012). The positive relation between ET and precipitation (fig. 32*A–C*) contrasts with the inverse relation between ET/P and precipitation (fig. 32*D–F*) during wet and dry climate conditions. ET/P reaches maximum values during periods of limited precipitation (fig. 32*E*). Average ET for the dry period in 1988 is about 27 in/yr, nearly 3.5 in/yr less than mean annual conditions and about 69 percent of the total precipitation for that year. Average ET for the wet period in 2004 is about 34 in/yr, nearly 3.5 inches higher than mean annual conditions and about 52 percent of the total precipitation for that year. Differences in the magnitude of annual ET between wet and dry periods are higher in Ohio, Maryland, Pennsylvania, and West Virginia than the differences indicated for States to the south (fig. 32*B–C*). During dry conditions, ET/P rates exceed 70 percent in much of the area in eastern Kentucky, Ohio, southwestern Pennsylvania, and West Virginia. Water stored in the soil during dry periods remains available for transpiration by ecosystems whose consumption adjusts to compensate for climate variability (Jones and others, 2012).

Seasonal changes in temperature and precipitation strongly influence annual water-budget components, particularly for basins having drainage areas less than 10 mi^2 that incorporate a wide range of year-to-year variability (Delin and Risser, 2007). Nonuniformity of weather systems (Delin and Risser, 2007) and different hydrologic processes operating across a range of scales (Blöschl and Sivapalan, 1995) tend to be homogenized in large basins, and the variability of the base-flow estimates decreases with increasing drainage area (fig. 33). The period-of-record annual average base flow for each of the 849 continuous-record streamflow gaging stations within the Appalachian Plateaus study (figs. 11, plotted as diamonds in fig. 33), show the range of average annual base flow from PART for various individual streamflow gaging station periods of record between 1900 and 2011. The scatter of average annual base flow from PART decreases from approximately 1 to 30 in/yr for basins less than 50 mi^2 to 3 to 14 in/yr for basins greater than 300 mi^2. The regression envelope narrows as shown by a decrease in the difference between minimum and maximum annual base flow with increasing drainage area. A regression of the average annual base-flow values from each drainage-area bin interval, however, has a near-zero slope along an annual base-flow value of approximately 10 in/yr. Base-flow data analyzed over long temporal scales, in this case 1900 to 2011, or large spatial scales (>200 mi^2) fail to characterize the annual variability depicted by small basins in fig. 33. Climate variation results in annual conditions that differ up to a factor of two or more from the long-term base-flow estimate of 10 in/yr for all study area basins shown in fig. 33. The effect of year-to-year variations in the amount and spatial distribution of precipitation on water budgets tend to be magnified in basins of limited area (generally less than 50 mi^2), which usually are located in the headwaters sections of watersheds. Delin and Risser (2007) suggest caution should be used when transferring recharge rates across various spatial scales.

Observed changes in low-flow discharges in streams of the eastern United States have been associated with the effects of spatial and temporal changes in precipitation (McCabe and Wolock, 2002; Lins and Slack, 1999; Karl and Knight, 1998) and temperature (Hodgkins and others, 2003). The response of streamflow to changes in climate conditions in the northeastern United States (Hodgkins and Dudley, 2011; Hodgkins and others, 2003) and southeastern United States (Patterson and others, 2012) have temporal trends consistent with monthly, interannual, and decadal-scale oscillations. In the northeastern United States, streamflow and bedrock groundwater levels respond similarly over time, although their responses differ in magnitude and duration (Weider and Boutt, 2010).

McCabe and Wolock (2002) noted an abrupt change in streamflows around 1970, coincident with increased variability in precipitation, particularly in the fall (September 1–November 30) in the eastern United States (Patterson and others, 2012; Karl and Knight, 1998). An apparent positive shift in minimum streamflows around 1970 was noted by Wiley (2006) in his evaluation of low-flow trends in the Appalachian Plateaus of West Virginia. Causes for the low-flow trends observed around 1970 have been associated with precipitation patterns, variations in ocean characteristics, and increasing atmospheric carbon dioxide in climate-model simulations (Patterson and others, 2012; Lins and Slack, 1999). The strength of such relations with the AMO (Patterson and others, 2012), NAO (Hodgkins and Dudley, 2011), El Nino-Southern Oscillation, Pacific Decadal Oscillation, and Pacific North America Index (McCabe and Wolock, 2014) in the eastern United States, however, is unclear.

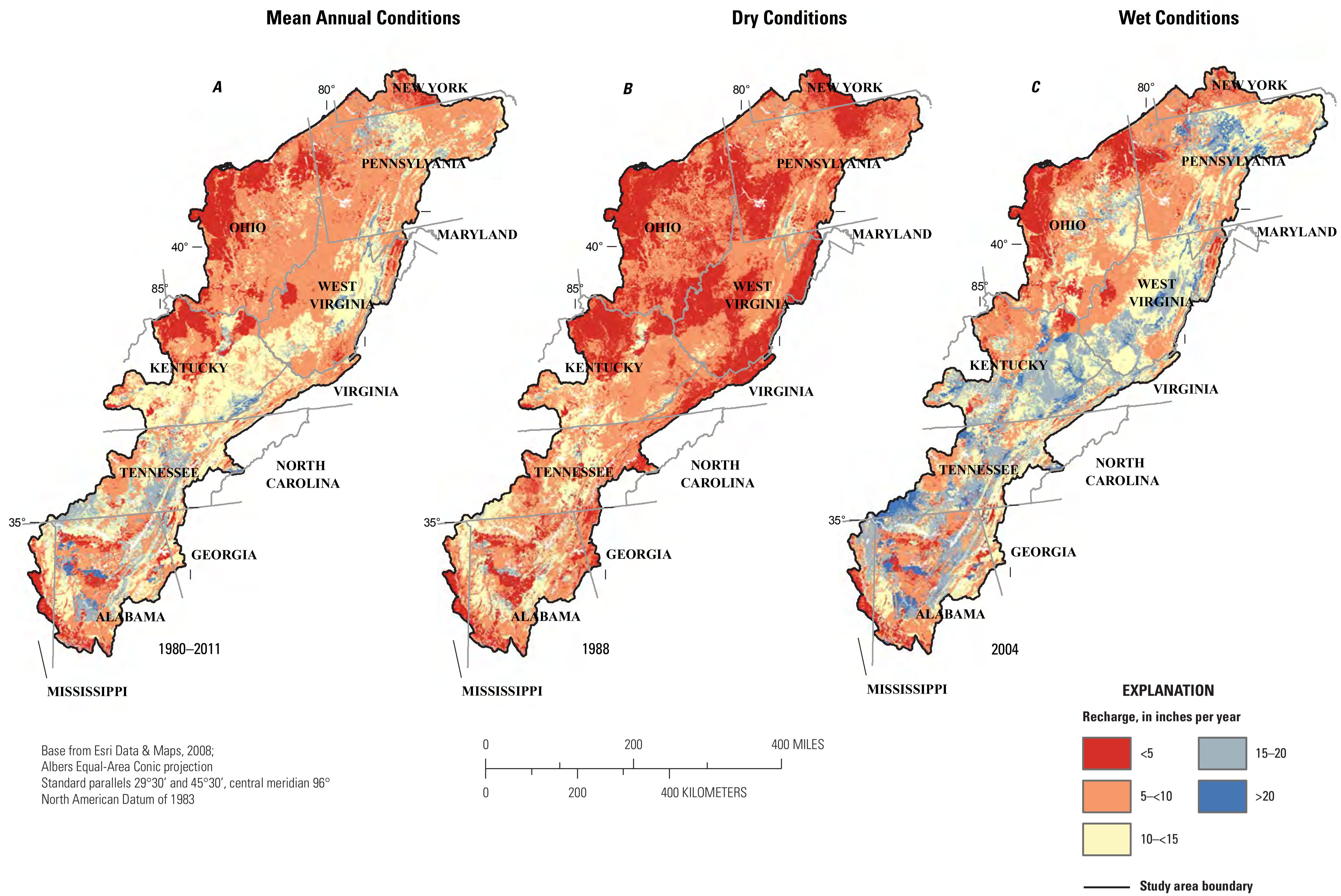


Figure 31. Recharge from the Soil-Water-Balance (SWB) model in inches per year for *A*, mean annual conditions from 1980–2011, *B*, dry conditions from 1988, *C*, wet conditions from 2004, and as percentage of precipitation for, *D*, mean annual conditions from 1980–2011, *E*, dry conditions from 1988, and *F*, wet conditions from 2004.

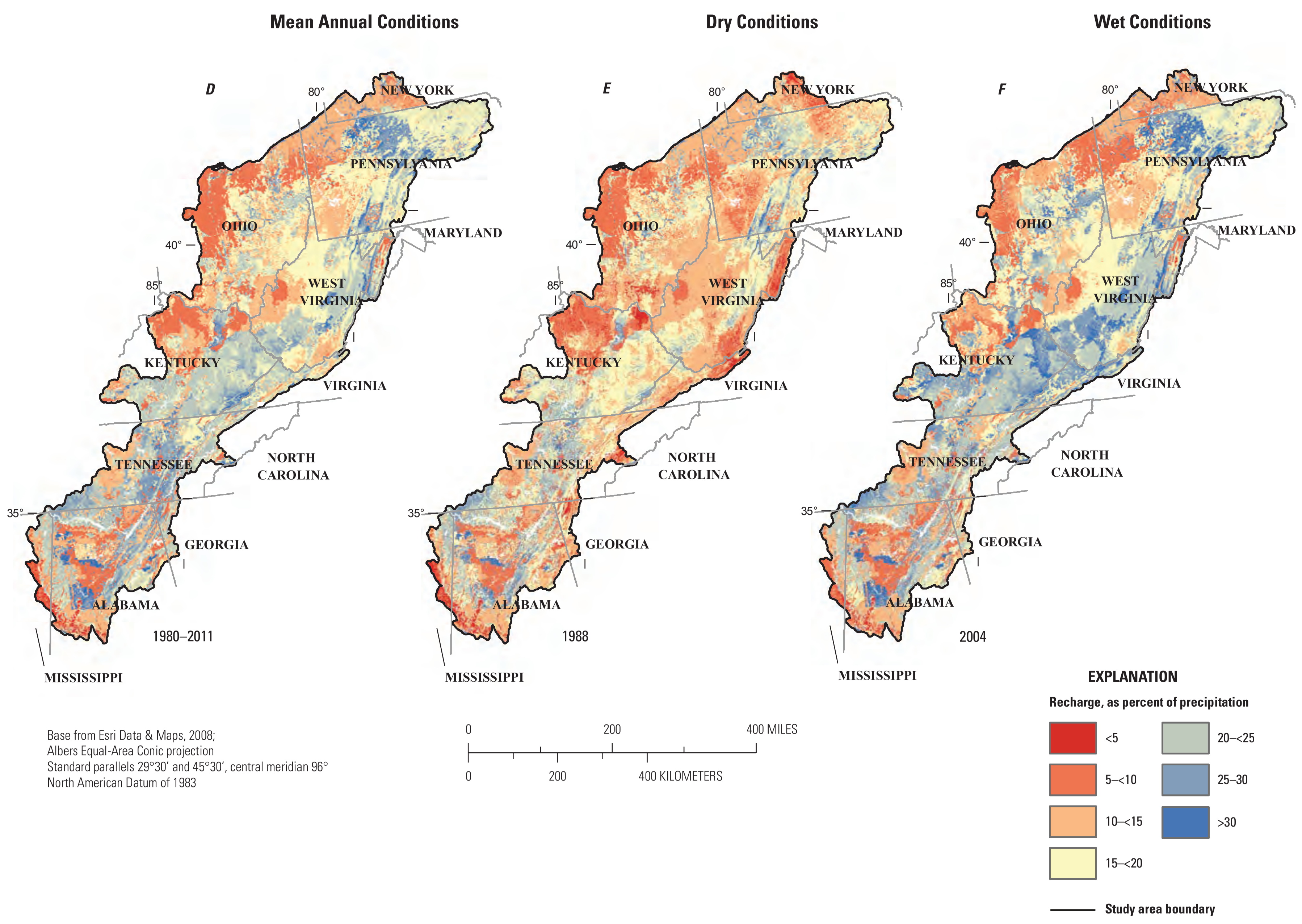


Figure 31. Recharge from the Soil-Water-Balance (SWB) model in inches per year for *A*, mean annual conditions from 1980–2011, *B*, dry conditions from 1988, *C*, wet conditions from 2004, and as percentage of precipitation for, *D*, mean annual conditions from 1980–2011, *E*, dry conditions from 1988, and *F*, wet conditions from 2004.—Continued

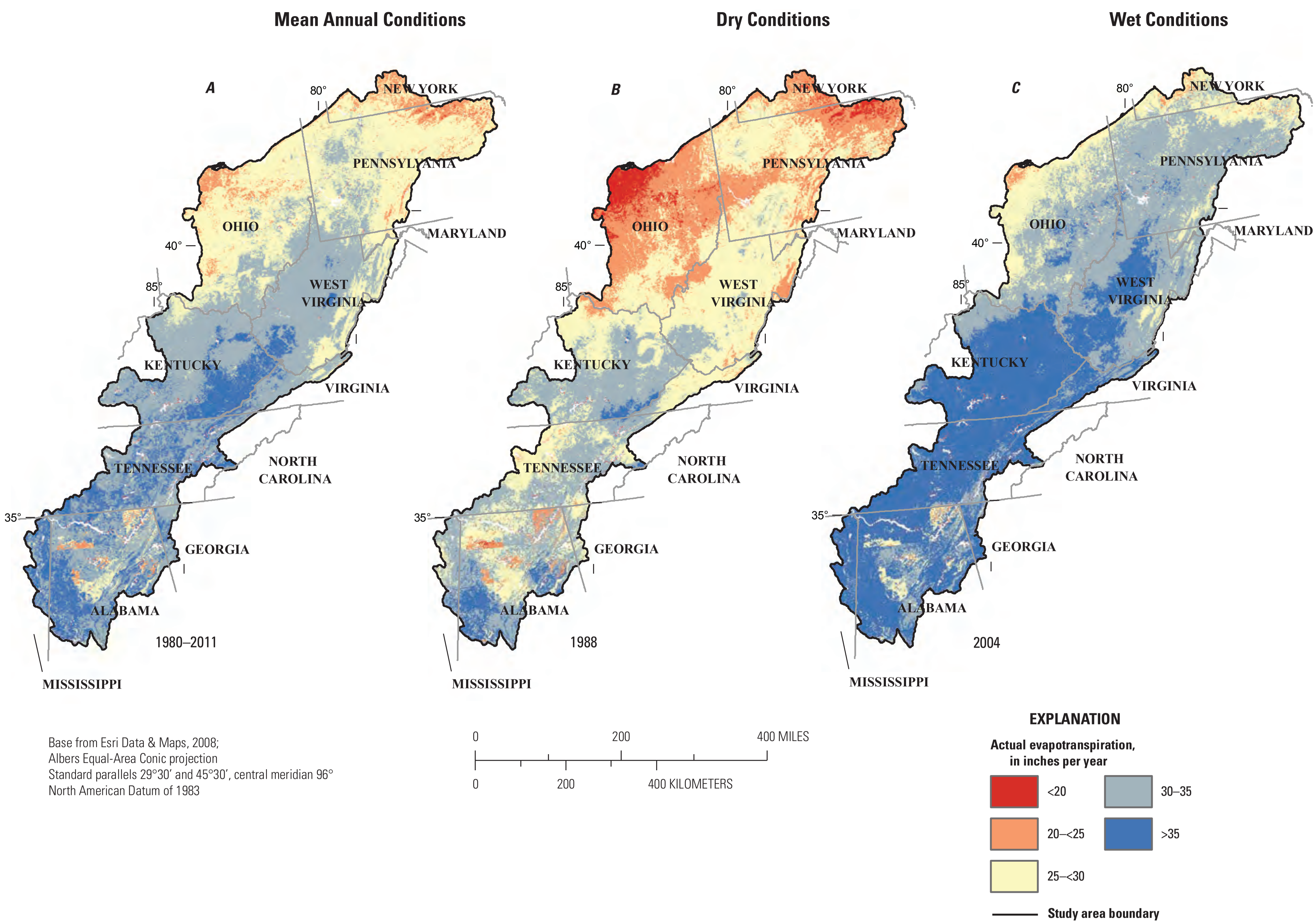


Figure 32. Actual evapotranspiration (ET) from the Soil-Water-Balance (SWB) model in inches per year for *A*, mean annual conditions from 1980–2011, *B*, dry conditions from 1988, *C*, wet conditions from 2004, and as percentage of precipitation for *D*, mean annual conditions from 1980–2011, *E*, dry conditions from 1988, and *F*, wet conditions from 2004.

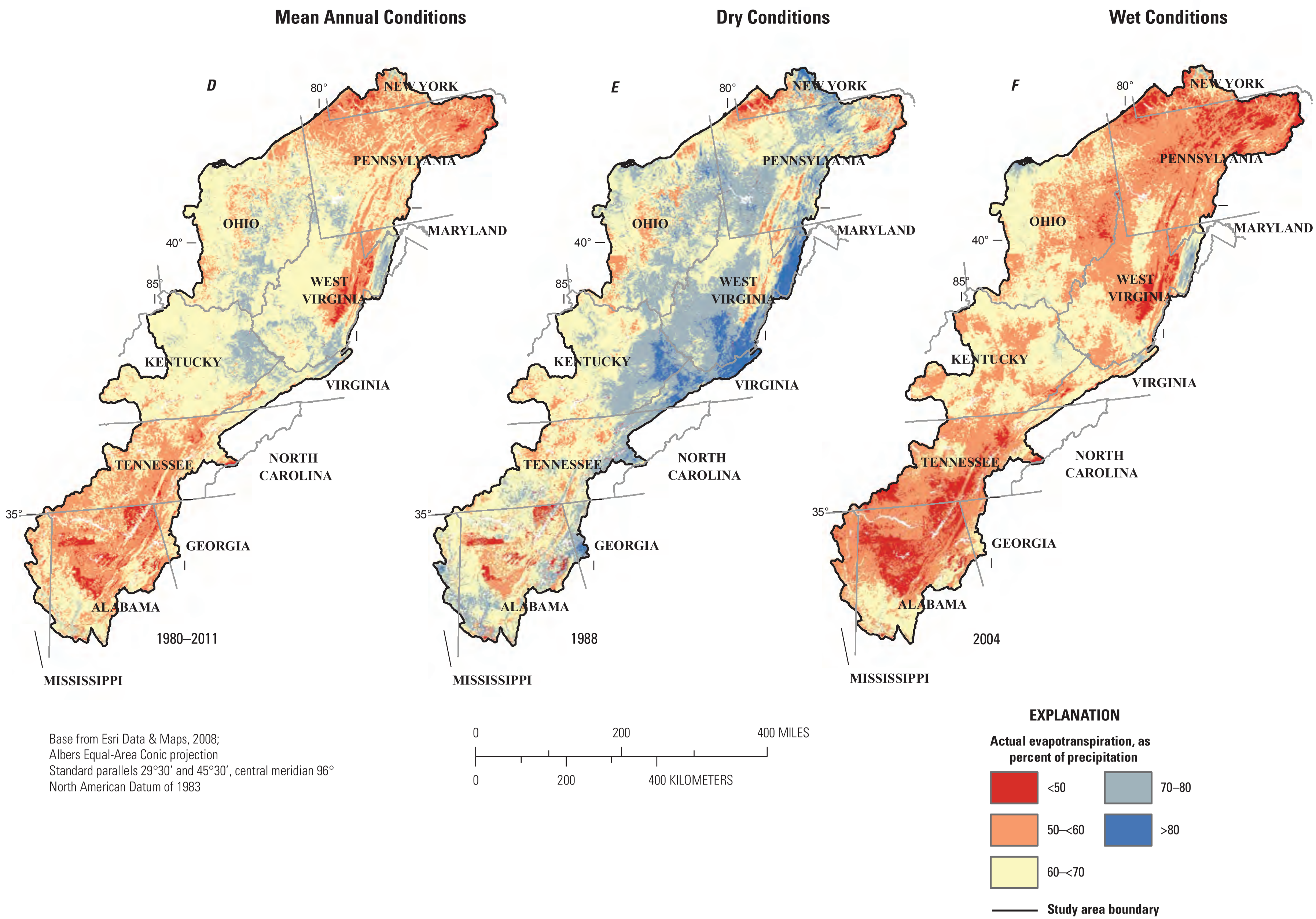


Figure 32. Actual evapotranspiration (ET) from the Soil-Water-Balance (SWB) model in inches per year for *A*, mean annual conditions from 1980–2011, *B*, dry conditions from 1988, *C*, wet conditions from 2004, and as percentage of precipitation for *D*, mean annual conditions from 1980–2011, *E*, dry conditions from 1988, and *F*, wet conditions from 2004.—Continued

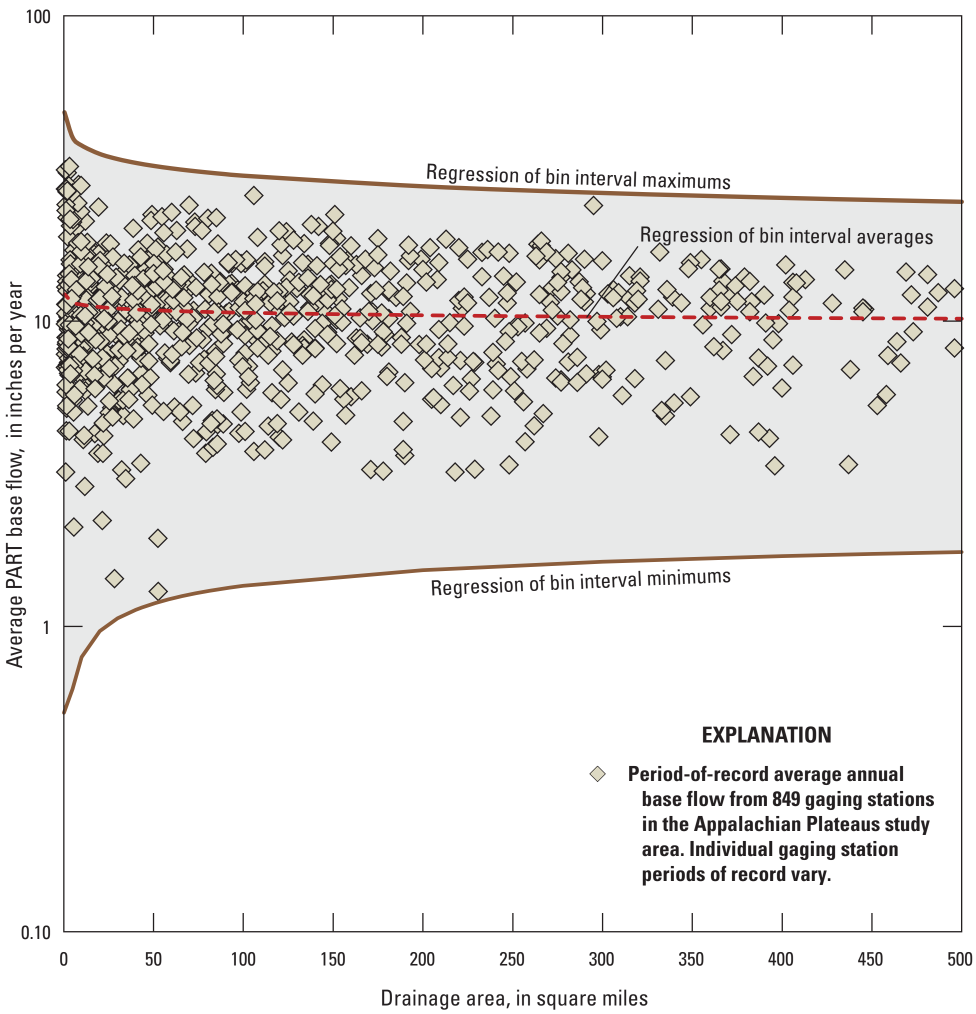


Figure 33. Relation between average base-flow values estimated from the PART method and drainage area of 849 streamflow gaging stations in the Appalachian Plateaus study area for the period 1900 to 2011.

Typical hydrologic responses to various climate conditions observed in the study area are depicted for three index streamgages in figure 34. Three patterns of change over time are evident in the components of streamflow (base flow and runoff) and BFI and were fairly consistent across the dataset. Because interannual variations in base flow, runoff, and BFI can be substantial, long-term trends in figure 34 were assessed by smoothing the data using a locally weighted scatterplot smoothing (LOWESS) (Helsel and Hirsch, 2002). Annual base flow, annual runoff, and BFI data all exhibit flat trends over the period of record for index streamgage 03051000, Tygart Valley River at Belington, WV (fig. 34*A*–*C*) despite a total increase in precipitation, as summarized from PRISM datasets, of 2.48 in/yr over the same period ending in 2011. The absence of a response in mean annual streamflow (base flow plus runoff) at gage 03051000 to increasing precipitation over the period analyzed indicates additional water from precipitation is either stored in aquifers, released during peak flows, or consumed by ET—all processes dependent on the timing and intensity of precipitation. Inverse trends are evident between base flow and runoff at index streamgage 03208500, Russell Fork at Haysi, Virginia (fig. 34*D*–*E*). Annual values of base flow have increased, while annual runoff has decreased over the period of record at gage 03208500. During this same period, BFI has increased and streamflow in the basin has become more dominated by base flow over time (fig. 34*F*). Many basins in the present dataset that are within the coal-measure areas show similar trends of increasing annual base flow and BFI, and decreasing annual runoff, with time. This phenomenon of increasing base flow over time in surface coal-mined areas of Virginia (Larson and Powell, 1986), and valley fill and deep coal mine areas of West Virginia (Zegre and others, 2014) is well documented. Many of the basins in the northern part of the study area in Ohio and Pennsylvania experienced changes in trends around 1970. Index streamgage 03118500, Nimishillen Creek at North Industry, OH illustrates that although annual base flow (fig. 34*G*) has increased over the period of record, the trend in annual values of runoff (fig. 34*H*) increased around 1970, causing the trend in BFI (fig. 34*I*) to flatten. It is interesting to note that base flow at 03118500 (fig. 34*G*) is increasing both prior to and following the 1970 step-change noted by McCabe and Wolock (2002). This is despite a change from a negative trend in total precipitation from 1930 to 1969 (–1.6-in/yr total for 40-year period) to a positive trend in precipitation post-1970 (0.7-in/yr total for 42-year period ending 2011) (PRISM Climate Group, 2012). Total streamflow at gage 03118500 has historically (1930–2011) increased due to increasing base flow (fig. 34*G*), but post-1970 precipitation appears to have resulted in both increased base flow and runoff (fig. 34*H*) amounts. Hodgkins and others (2007) noted similar post-1970 increases in runoff in the Great Lakes Basin, which includes portions of northern Ohio.

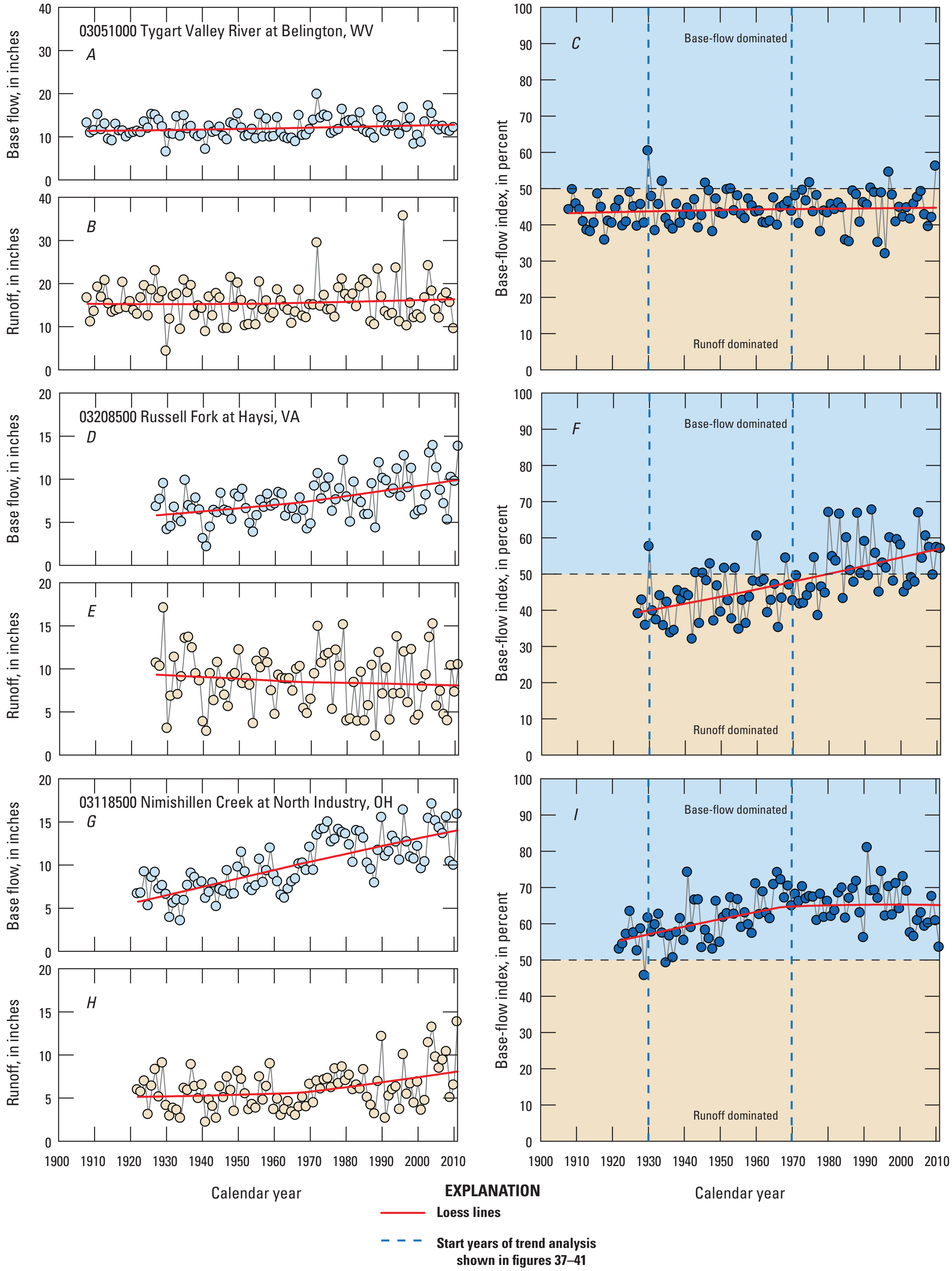


Figure 34. Annual values of base flow, runoff, and base-flow index at streamflow gaging stations. *A–C*, 03051000 Tygart Valley River at Belington, West Virginia, *D–F*, 03208500 Russell Fork at Haysi, Virginia, and, *G–I*, 03118500 Nimishillen Creek at North Industry, Ohio.

Trends for individual streamflow gaging stations such as those in figure 34 are informative on a local scale, but when combined with data from all 849 streamflow gaging stations (fig. 11) can provide meaningful insight into broader-scale trends across the entire Appalachian Plateaus study area. Because periods of record amongst the individual streamflow gaging stations are not synoptic, the study-area-wide synthesis combines data from several time periods in order to evaluate long-term trends. For each year of streamflow record at each respective streamflow gaging station, equation 1 was used to compute normalized annual anomalies of precipitation, total streamflow, base flow, runoff, and BFI from PRISM and hydrograph separation datasets (fig. 35; Nelms and others, 2015). When plotted together, normalized annual anomalies from each streamflow gaging stations appear as overlapping time series of standard deviation that in sum appear as a single band around the normals as shown in red in figure 35. An annual average of all anomalies for a given year (shown by black line) and a 10-year moving average of all anomalies (shown by yellow line) provide two windowing techniques to highlight data trends (fig. 35). The range of anomalies for each of the plot in figure 35 increases with time and number of gages, particularly during the period from 1910 to 1945. It is possible that the range of anomalies is due to the size of the dataset across a large study area, an increase in number of small (>50 mi^2) gaged basins in the dataset, or that greater temporal resolution in the anomaly data would result in a narrower range of fluctuation, such as that seen in monthly data from the northeastern United States (Weider and Boutt, 2010).

The annual and 10-year moving average for precipitation (fig. 35*A*) show a change from negative to positive anomalies about 1970, consistent with the previously defined transition to a wetter climate in the Appalachian Plateaus (fig. 10) and the timing of a step-change in runoff across the United States noted by McCabe and Wolock (2002). The shift from negative to positive anomalies is present in the streamflow (fig. 35*B*), base flow (fig. 35*C*), and runoff anomalies (fig. 35*D*) with little apparent time lag (fig. 36). Similarity in annual trends in figure 35*A*, *C* indicates the condition of Appalachian Plateaus aquifers is susceptible to prevailing climate patterns, a condition strongly influenced by seasonal differences in precipitation and temperature (Reese and Risser, 2010). Because the 10-year average line summarizes data across a larger window than the annual line, dry conditions in the late 1960s result in a 10-year moving average peak at 1980 that lags the annual average peaks in figure 35*A*–*D* from 1971 to 1975. Runoff anomalies show less-pronounced variability and a more subtle response to the shift toward wetter climate conditions around 1970 than the base-flow anomalies (fig. 36). This observation supports conclusions that historic changes in precipitation patterns in the eastern United States over the last century have had a pronounced effect on the groundwater system (Weider and Boutt, 2010).

Groundwater discharge to streams is particularly important during periods of drought. The drought of greatest severity in the Appalachian Plateaus was in the early 1930s, although longer but less severe periods of below-normal precipitation occurred from 1952 to 1955 and 1968 to 1971 (fig. 35*A*). Precipitation anomalies reach a maximum below-normal value in 1930 and continue to stay below normal until 1932 (fig. 35*A*). Except for BFI anomalies that reached maximum above-normal values in 1930 (fig. 35*E*), all anomalies show a rapid decrease from above- to below-normal values from 1929 to 1930. The historic regionwide drought of the 1930s shows that groundwater as a proportional source of water to streams reached maximum levels in 1930 (fig. 35*E*) but was near normal for the remainder of the drought. The BFI values were similarly near normal for longer periods of below-normal precipitation in the 1950s and 1960s. Stability in the BFI during below-normal precipitation periods is surprising in context of SWB model results that show a reduction in recharge of greater than 30 percent from 1980–2011 normals during a dry year (fig. 31*B*). The BFI anomalies do not mimic precipitation patterns, but trend upward over the last century (figs. 35*E*, 36) in contrast to the more variable anomalies shown in streamflow, base-flow, and runoff (figs. 35*B*–*D*, 36). Data in figure 36 show changes in the proportions of base flow and runoff in streamflow of the Appalachian Plateaus are occurring at temporal scales greater than 1 year, and that the base-flow component of total streamflow is increasing at a faster rate than the runoff component. Although beyond the scope of the current study, long-term shifts in the interannual timing of precipitation, such as those noted by Neff and others (2000), have more impact on seasonal streamflows (Tu, 2009) than annual streamflows. Despite annual climate fluctuations and seasonally variable precipitation totals, the absence of clear long-term trends in streamflow, base flow, and runoff (fig. 35*B*–*D*) support the assumption that net gains or losses in aquifer storage balance over multi-year periods.

Spatial Variation in the Hydrologic Budgets

Spatial variations in the hydrologic budgets were evaluated by calculating the magnitude of change for the various water budget components by multiplying the slopes of nonparametric Kendall-Theil robust lines for the time periods 1930–2011, 1930–1969, and 1970–2011 at the index streamgages by the number of years in each period. These time periods approximately parallel the periods used to assess streamflow changes in the South Atlantic region of the United States by Patterson and others (2012), base flow in the Midwest (Juckem and others, 2008), and the overall hydrologic cycle in the Northeast (Huntington and others, 2009). Changes in the input (precipitation) were calculated from the average values clipped from the PRISM annual precipitation grids

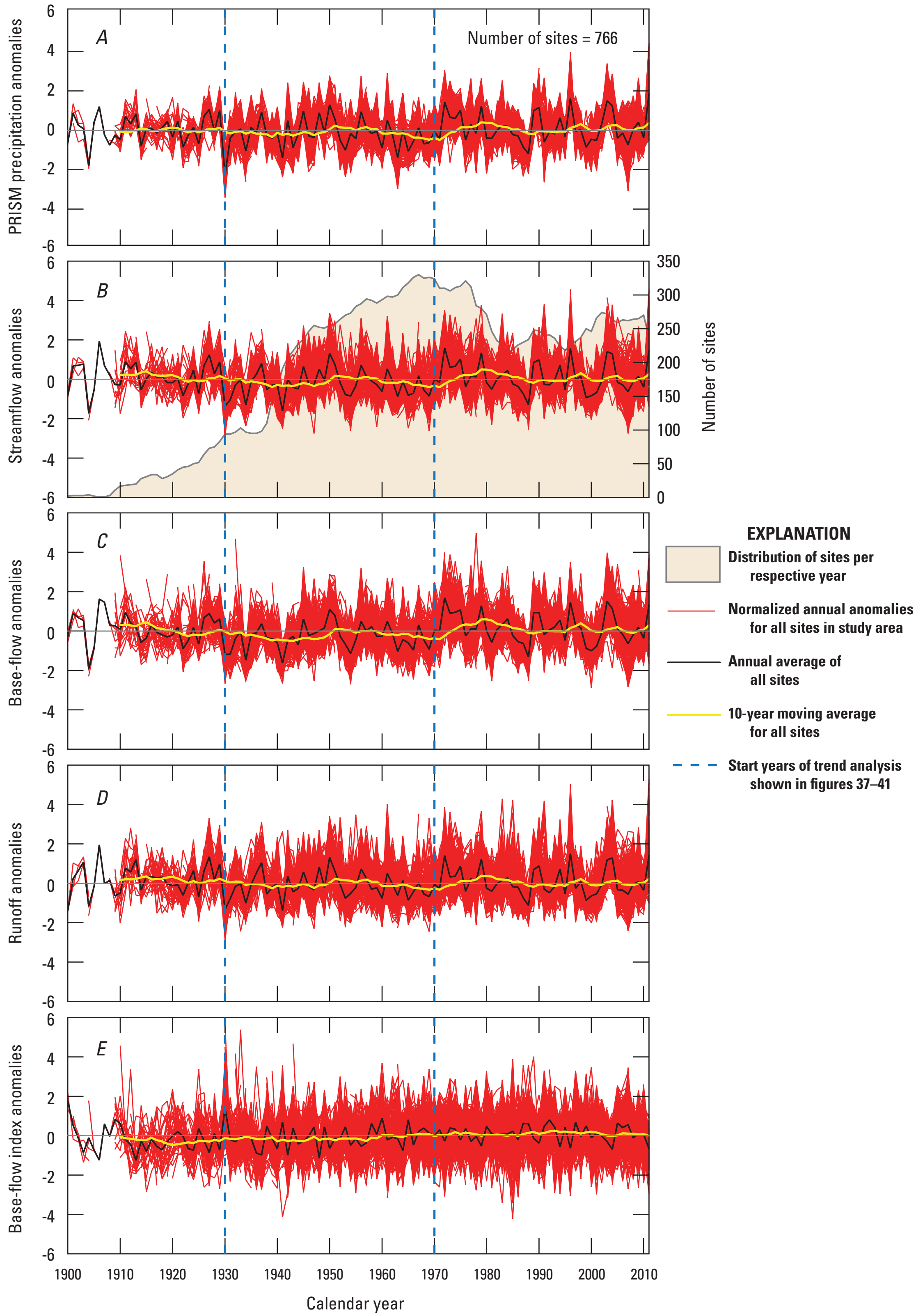


Figure 35. Normalized annual anomalies for *A*, precipitation, *B*, streamflow, *C*, base flow, *D*, runoff, and *E*, base-flow index for all sites in the Soil-Water-Balance (SWB) model (red lines). The black line through these data is the annual average of all sites. The yellow line is a 10-year moving average for all sites. The shaded region in graph *B* is the distribution of sites per respective year.

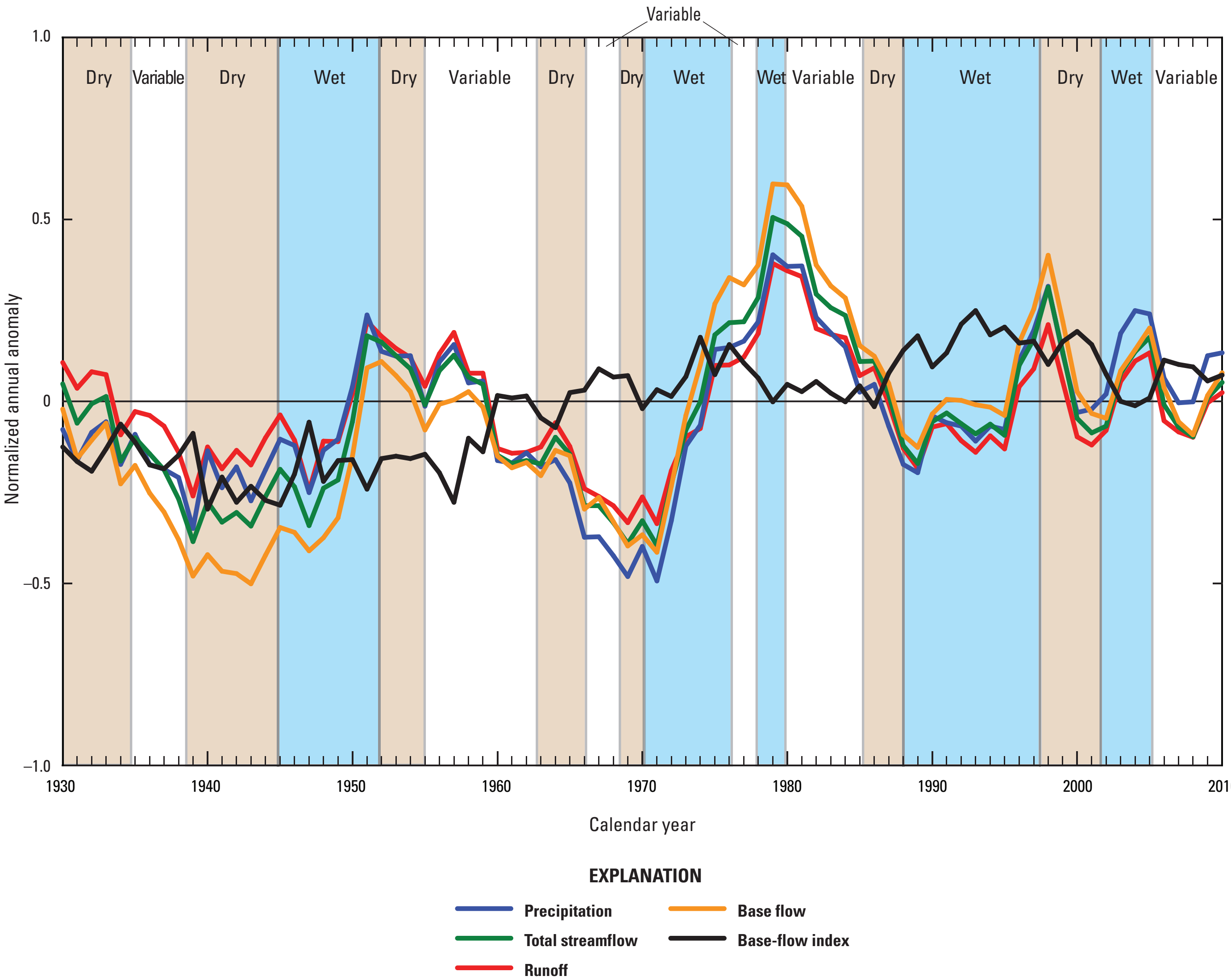


Figure 36. 10-year moving average of normalized annual precipitation, total streamflow, runoff, base flow, and base-flow index anomalies from 849 streamflow gaging stations within the Appalachian Plateaus study area. Periods of wet, dry, and variable conditions identified from Palmer Drought Severity Index in figure 10.

using basin boundaries from the GAGES–II dataset (Falcone, 2011). The magnitude of change of the output components (streamflow, base flow, and runoff) of the hydrologic budget were based on annual values from the PART hydrograph separation analysis from Nelms and others (2015). Changes in the percentage of base flow that composes streamflow were calculated from annual values of BFI from PART analysis.

Total annual precipitation (fig. 37*A*) increased throughout the study area over the entire record (1930–2011) and ranged from 0.97 to 9.59 inches. In the mid-20th Century (1930–1969), however, total annual precipitation in a majority of the basins in the northern half of the study area decreased from –5.87 to –0.08 inches (fig. 37*B*). Declines in total annual precipitation were dominant in the southern half of the study area in the late 20th Century (1970–2011), whereas declines and increases were roughly equal in the northern half during the same period (fig. 37*C*). The negative trends tended to result in values greater than –5.00 inches throughout the study area. Basins along Lake Erie consistently experienced positive trends for all time periods.

Increasing trends in annual precipitation during the entire period suggest that streamflow should have corresponding trends that increase over time, but mean annual streamflow decreased in many of the basins in the southern half of the study area (fig. 38*A*). Several of the basins in the southern half of the study area and along Lake Erie experienced positive trends in streamflow during the mid-20th Century, while a majority of basins in the northern half had negative trends that ranged from –8.02 to –0.15 inches (fig. 38*B*). The negative trends in streamflow became more widespread throughout the study area and intensified in magnitude during the late 20th Century, especially in the southern basins (fig. 38*C*).

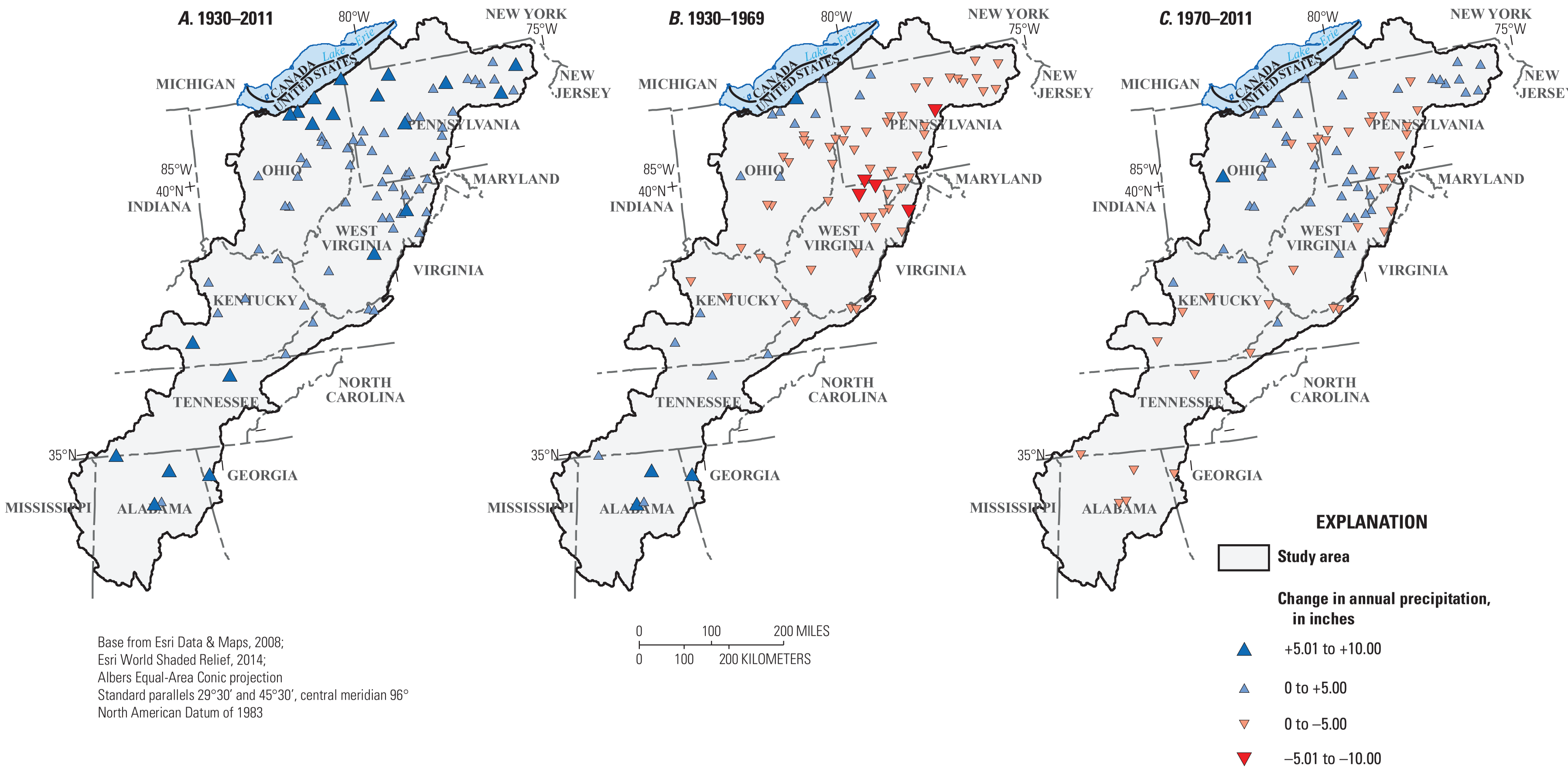


Figure 37. Changes in annual precipitation for *A*, 1930–2011, *B*, 1930–69, and *C*,1970–2011 at index streamgages in the Appalachian Plateaus study area.

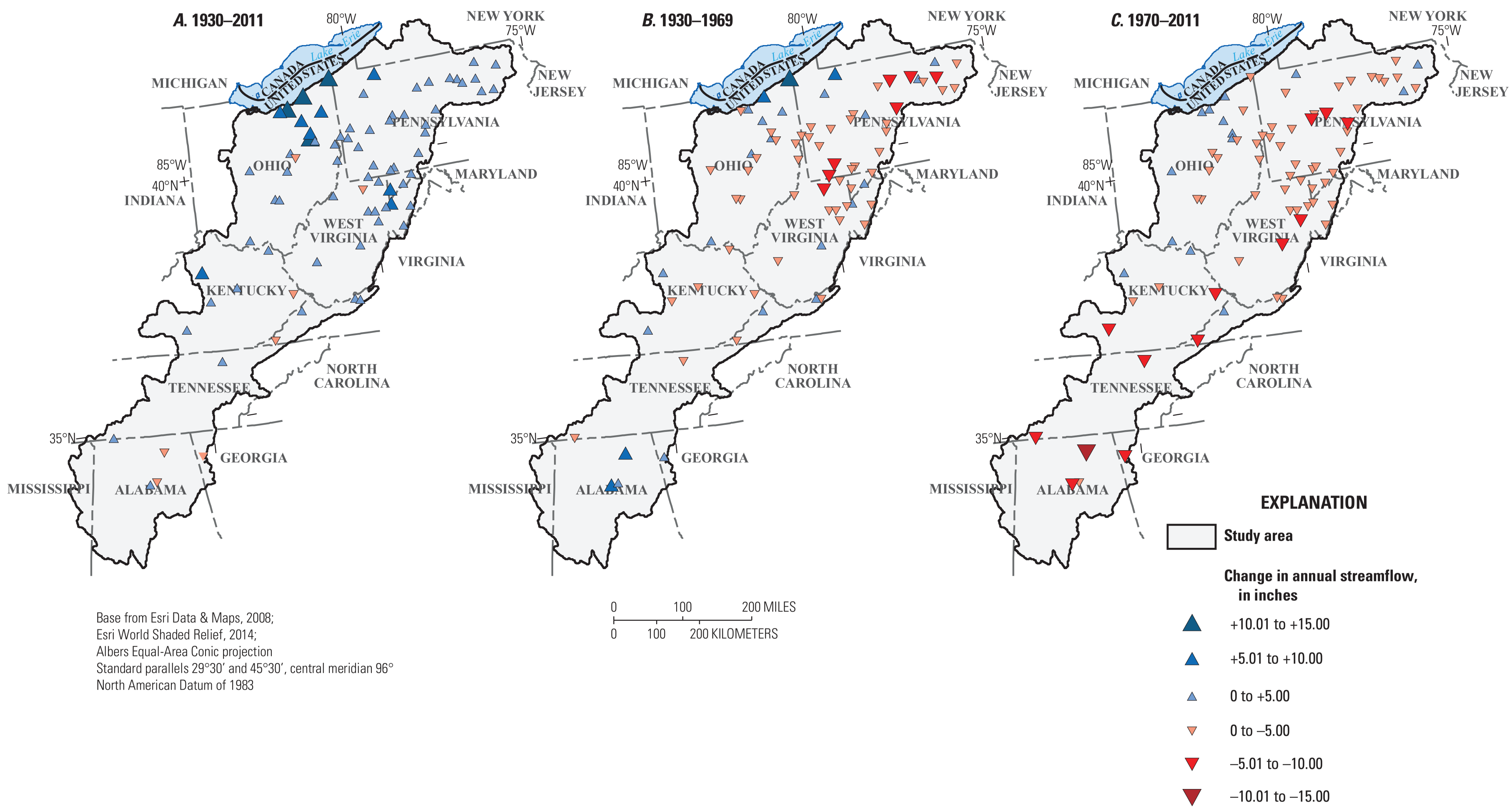


Figure 38. Changes in annual streamflow for *A*, 1930–2011, *B*, 1930–69, and *C*, 1970–2011 at index streamgages in the Appalachian Plateaus study area.

More detail about streamflow trends can be ascertained by evaluating trends for the individual components of streamflow, namely base flow and runoff. Base flow increased in all basins for the entire period (fig. 39*A*), but runoff both decreased and increased (fig. 40*A*). This indicates that for many basins in the Appalachian Plateaus region, the increase in streamflow between 1930 and 2011 is in response to increased groundwater discharge or base flow, as proven by the positive trends in BFI (fig. 41*A*). During the mid-20th Century, negative trends in annual base flow (fig. 39*B*) and runoff (fig. 40*B*) are evident throughout the study area. Many basins in the northern half of the study area where annual precipitation had negative trends in the mid-20th Century showed large increases in BFI (fig. 41*B*) that ranged from 0.2 to 16 percent, which indicates that base flow was the dominant component for sustaining streamflow during this period. Negative trends in annual base flow (fig. 39*C*) and runoff (fig. 40*C*) became more dominant and widespread in the late 20th Century. A reversal in BFI trends (fig. 41*C*) occurs between the mid- and late-20th Century periods. Negative trends in BFI ranged from –4.79 to –0.09 percent for many of the basins in the northern half of the study area that experienced large positive BFI trends in the earlier period. Many of these basins experienced negative trends in annual precipitation in the late-20th Century, which normally results in base flow being the major component of streamflow. This indicates that trends, especially in these basins, are not completely the result of climatic changes and may be more influenced by other factors, including changes in land-use activities. It is also worth noting that temporal trends in precipitation, streamflow, runoff, and base flow in the study area are influenced by the time period being investigated and that several time periods were evaluated prior to presentation of the results herein.

Groundwater Withdrawals

Total groundwater withdrawals generally decreased over the 20-year period from 1985 to 2005 (fig. 42; Kenny and others, 2009). There are exceptions resulting from changes in collection methods used by the USGS to gather water-use information. During 1985 and 1990, water use for mining practices included the volumes associated with mine dewatering practices. Water use for the mining category accounted for about 48 percent of the total groundwater withdrawn in 1990. Large increases in 1990 water-use for the dewatering of mines were reported for the States of West Virginia, Ohio and Tennessee, in respective order of rates of withdrawal. In 1995, 2000, and 2005, dewatering operations were neither considered nor accounted for in the mining water-use category. In 2005, water use for the mining category composed only about 4 percent of the total groundwater withdrawals in the 186 counties of the study area.

Population within the study area remained largely unchanged from 1985 to 2005, and averaged about 13.5 million people. The estimated rate of domestic groundwater withdrawals slightly increased from 1985 to 1995, followed by a 12-percent decrease from 1995 to 2005. This decrease may be due to the expansion of public-supply systems, which primarily use surface water in the study area (fig. 24), and residential areas that may have switched from a domestic groundwater self-supply system to the public surface-water supply system. Self-supplied industrial and commercial use of groundwater peaked at about 276 Mgal/d in 1990, but decreased more than 60 percent to about 101 Mgal/d in 2005.

Agricultural use of groundwater is the aggregate of irrigation, livestock, and aquaculture water-use categories for the study area. In 2005, groundwater used for agriculture composed about 9 percent of the total groundwater withdrawals (fig. 42). More than half of the water for agriculture was used for livestock purposes associated with livestock watering, feedlots, dairy operations, and other on-farm needs. Agricultural withdrawals are computed from reported crop, stock, or other agricultural yield information (Hutson, 2007). The lowest estimates for agricultural withdrawals were in 2000, during a relatively dry period (fig. 10). The decrease of more than 75-percent in withdrawals for agricultural use for 2000 (fig. 42) is not a reflection of changes in precipitation but, rather, is due to the unreported livestock withdrawals during that time period (Hutson, 2007). On average, the agricultural use of groundwater ranged from about 40 to 55 Mgal/d, and around 80 percent was attributed to livestock use of the groundwater. The States of Alabama and Pennsylvania, on average, were responsible for more than 75 percent of the livestock groundwater use in the study area due to an abundance of dairy and poultry farms in those States.

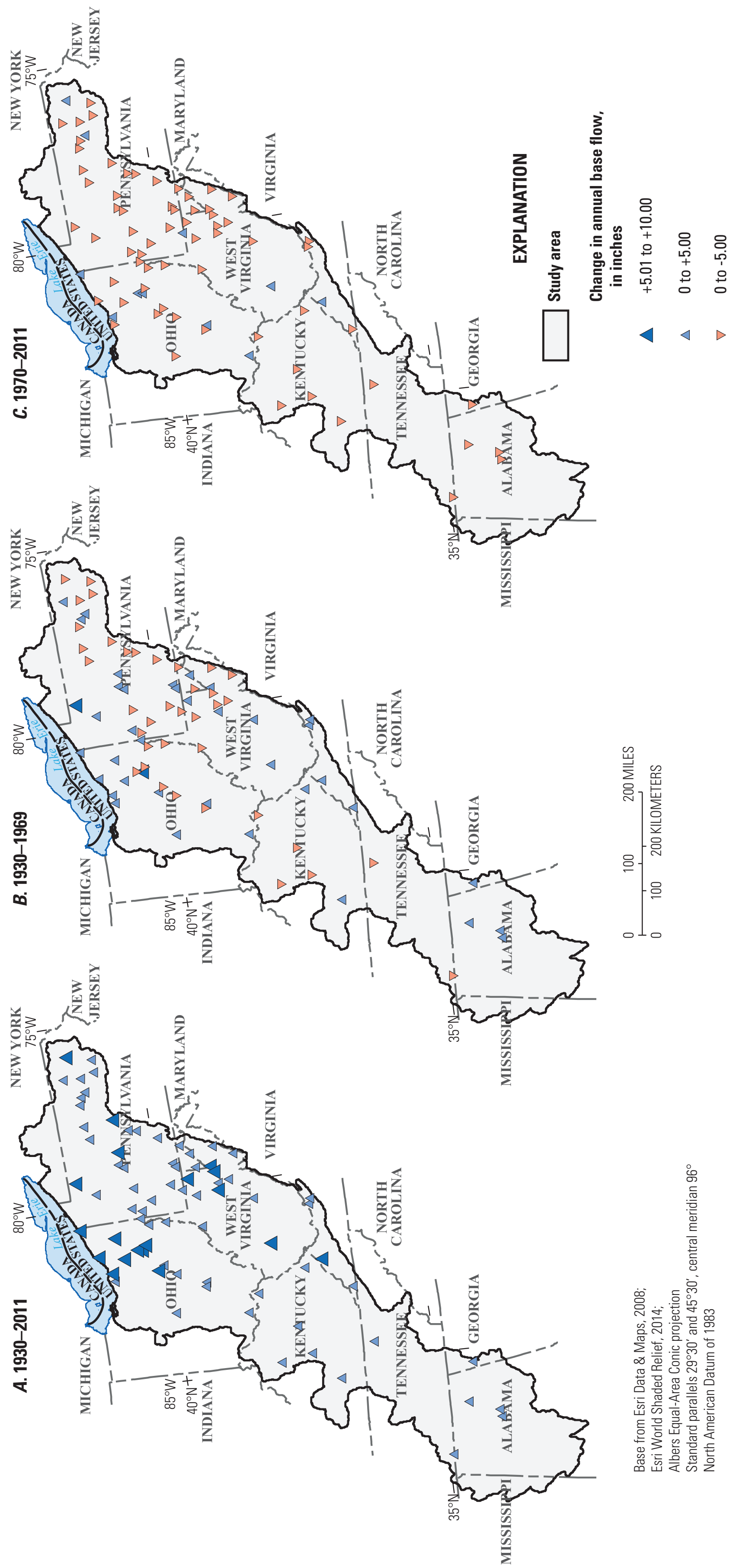


Figure 39. Changes in annual base flow for *A*, 1930–2011, *B*, 1930–69, and *C*, 1970–2011 at index streamgages in the Appalachian Plateaus study area.

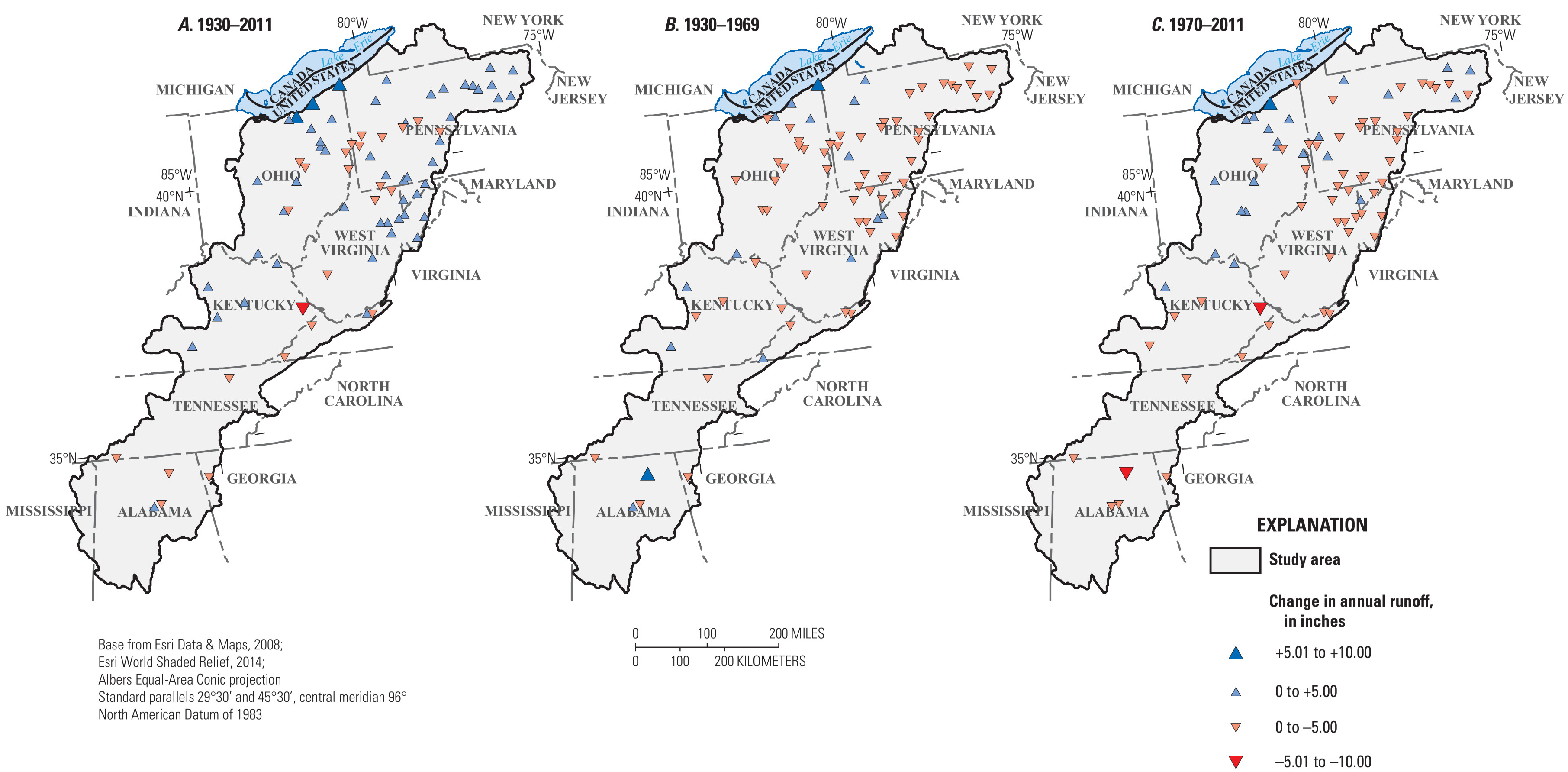


Figure 40. Changes in annual runoff for *A*, 1930–2011, *B*, 1930–69, and *C*, 1970–2011 at index streamgages in the Appalachian Plateaus study area.

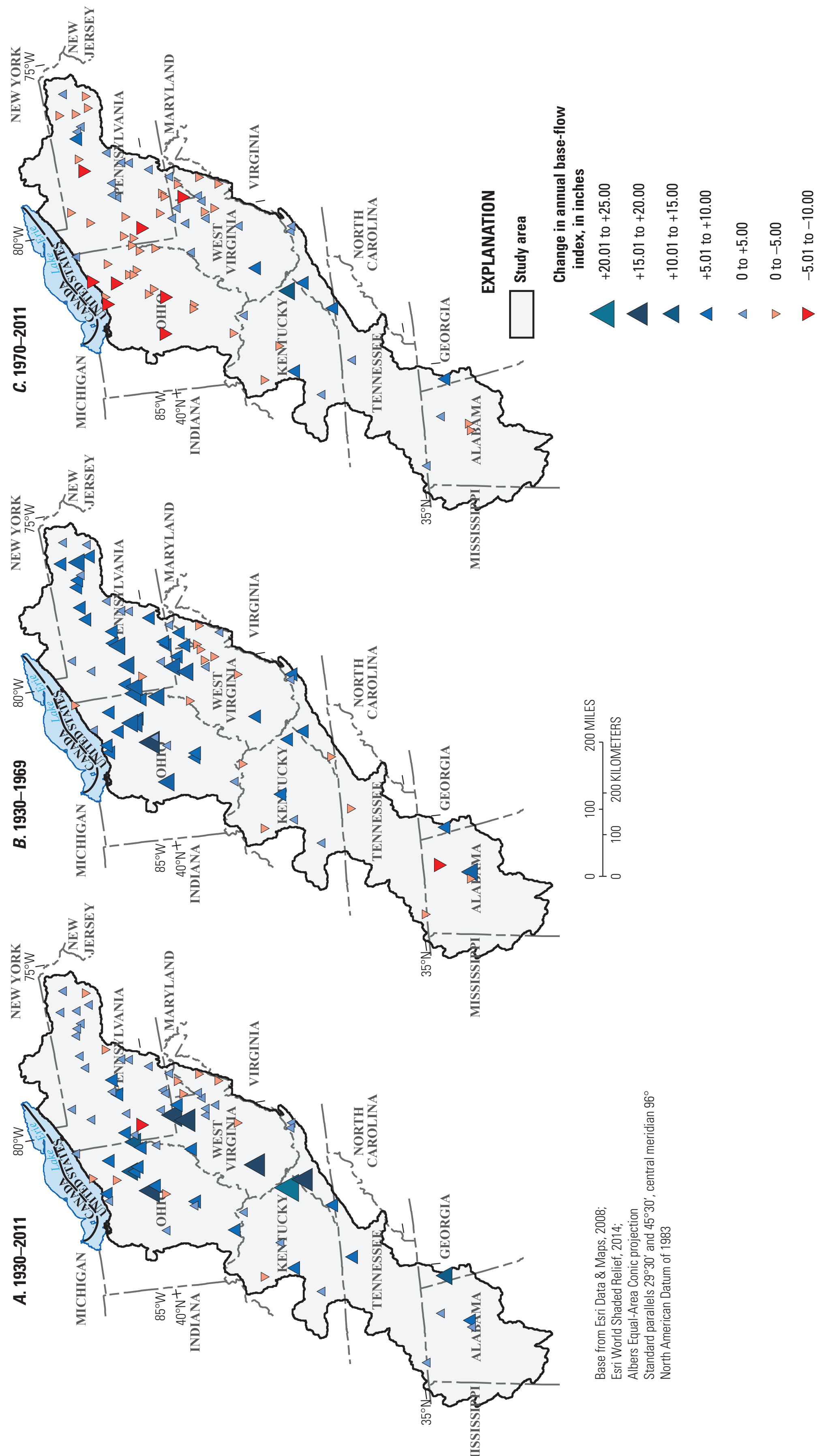


Figure 41. Changes in annual base-flow index for *A*, 1930–2011, *B*, 1930–69, and *C*, 1970–2011 at index streamgages in the Appalachian Plateaus study area.

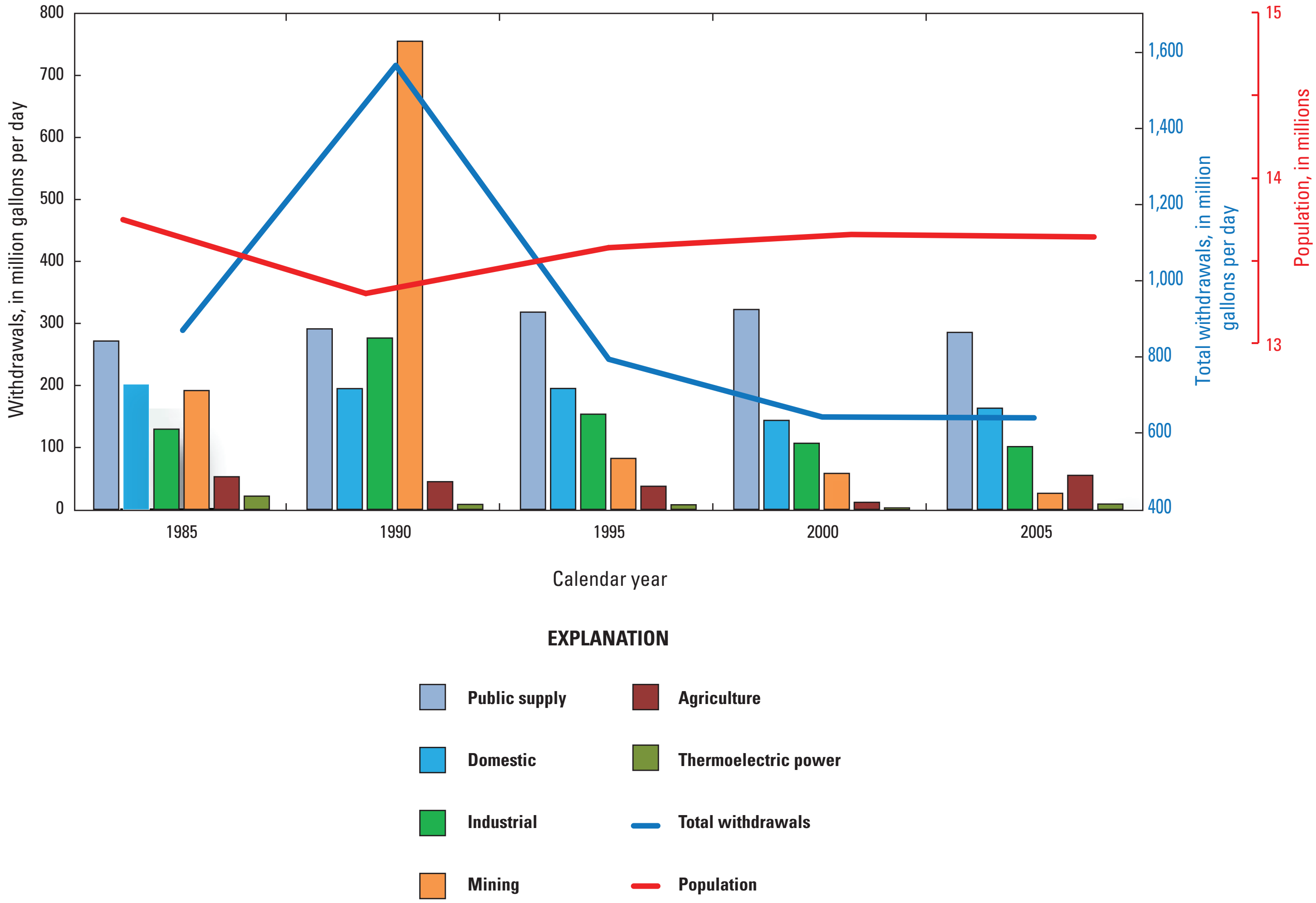


Figure 42. Trends in groundwater withdrawals by water-use sector for selected counties within the Appalachian Plateaus region (1985–2005). Data from Kenny and others (2009).

Summary and Conclusions

The Appalachian Plateaus occupy approximately 86,000 square miles in portions of Alabama, Georgia, Kentucky, Maryland, Ohio, Pennsylvania, Tennessee, Virginia, and West Virginia. Fractured rock aquifers of Permian-, Pennsylvanian-, and Mississippian-age sandstones, siltstones, shales, limestones, and coal underlie the area, which has extensive energy resources. Coal, oil, and natural gas have historically been extracted from the area for over a century, as have timber, aggregate, and other natural resources. The availability of freshwater resources is vital to the continued development of all of these natural resources in the region. A better understanding of groundwater availability in the Appalachian Plateaus thus plays a central role in sustained economic development of the region.

A regional water-budget approach supports previous concepts that groundwater discharge composes between 50 and 65 percent of mean annual streamflow in the region. Streams in the Appalachian Plateaus are dependent on an annual average of 9 inches per year of base-flow discharge. Depending on prevailing climate conditions, base-flow discharge across the area may differ by a factor or two around mean annual conditions. The wide range of basin characteristics in the area equates to high spatial variability in water budgets, and this variability tends to be greater in smaller basins than in larger basins where climate, geology, land use, and other factors are relatively homogenized. Evapotranspiration is the largest component of the Appalachian Plateaus water budget and composes 62 percent of the mean annual outflow from the system. This value can be as high as 69 percent regionally or greater than 80 percent locally during periods of drought. Rates of evapotranspiration in the humid Appalachian Plateaus are lowest, in terms of percentage of precipitation, in high altitude and low temperature areas of Maryland, Pennsylvania, and West Virginia.

The regional water budget for the Appalachian Plateaus has remained essentially unchanged from pre-development to the present despite large-scale changes in land use (coal mining) and temporally variable climate conditions. Although important at a local scale, post-development changes in the volumes of water stored in abandoned-coal mine aquifers of Appalachian Plateaus is negligible when compared to the annual volumes of water gained or lost to the system through recharge and ET processes. Annual climate fluctuations and seasonally variable precipitation totals result in net gains or losses in aquifer storage that are assumed to balance over long periods.

Streamflow and its components, base flow and runoff, are closely related to annual climate patterns in the region. Base flow displays greater annual variation than runoff, indicating that changes in annual climate conditions that influence streamflow are often the result of changes in the groundwater system. The timing of precipitation is probably important in an area such as the Appalachian Plateaus, where much of the warm-month precipitation is lost to evapotranspiration. Long-term changes in the water budget reflect previously documented patterns in streamflow across the conterminous United States. The proportion of stream water composed of groundwater, however, appears insensitive to annual climate patterns. A synthesis of all hydrograph separation data indicates a region-wide increase in the groundwater proportion of streamflow that can be attributed to both land-use and climate factors.

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For more information about this publication, contact:

Director
USGS Virginia Water Science Center
1730 East Parham Road
Richmond, VA 23228
(804) 261-2600
(804) 261-2657 fax

Or visit the Virginia Water Science Center Web site:

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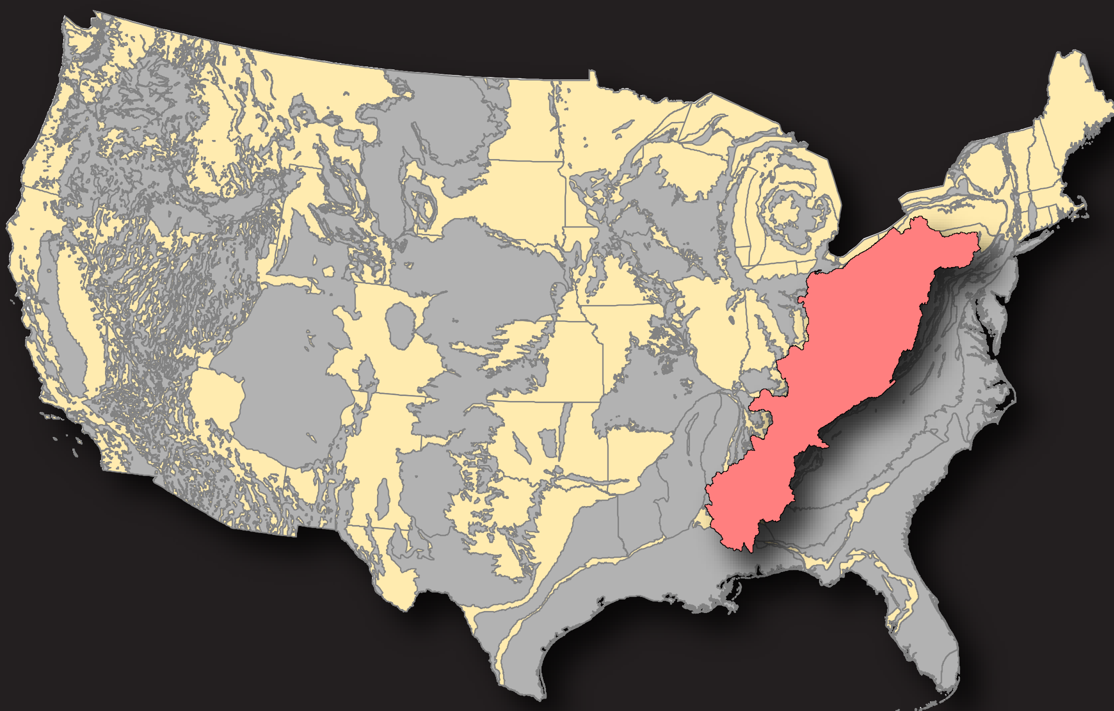

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