Hydrology of the Greenbrier Karst

3

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Abstract

Studies of the karst drainage systems of the Greenbrier limestone in southeastern West Virginia began in the early 1960s and were the first to make extensive use of water-tracing techniques and cave mapping in the USA. The carbonate aquifer is about 400 ft (120 m) thick in the Swago Creek area west of Marlinton (Pocahontas County) increasing to 1000 ft (300 m) in southern Monroe County. The basic hydrogeologic setting for the region consists of relatively flat-lying limestones exposed in valleys or plateaus and surrounded by higher elevation clastic units. Recharge to the conduit aquifer system is by capture of surface streams originating on the clastic rocks (allogenic recharge) and water infiltrating through the extensive areas of dolines (autogenic recharge). Only a few surface streams cross the carbonate outcrop, and even, these tend to loose water into the karst drainage systems. Much of the flow through the aquifer is through conduits under open channel conditions much like a surface stream with a roof. Discharge is concentrated at large springs that typically display rapid response to storm events, and the ratio of maximum to minimum discharge exceeds 100:1 for most of the springs. The karst caves and conduits are generally decoupled from surface topographic features, and the patterns of mapped cave passages are influenced by structural and stratigraphic characteristics. Insoluble beds within the Greenbrier Group may perch underground streams well above the apparent base level, and the underlying Maccrady Shale acts as an aquitard with several large caves developed along the contact of the shale and the overlying limestone. Much of this area can be considered a "contact karst" with the clastic rocks delivering concentrated recharge water onto the soluble limestones and the underling shales eventually forcing the return of conduit flow to the surface. The available data on water wells in the limestone suggest that most are actually producing from shaley units with the limestones acting as confining beds. The Greenbrier River and its tributaries represent base level for most of the area, and the relief of several hundred feet provides the hydraulic gradient. The area is underdrained by a well-integrated network of caves.

3.1 Introduction

This chapter presents an overview of studies of the karst hydrology of the Greenbrier limestone in southeastern West Virginia. The area ranges from Swago Creek west of Marlinton in Pocahontas County to Greenville in Monroe County (Fig. 3.1). The Greenbrier Group is exposed in an upland valley or plateau trending northeast/southwest. This is a mature karst aquifer with few surface streams making it across the width of the carbonate outcrop. The Greenbrier limestone outcrop area is less than one-mile wide (1.6 km) in the northern part to about 10 miles (16 km) wide in Monroe County, and the thickness increases from about 400 ft (120 m) in the Swago Creek area to 1000 ft (300 m) in the Greenville area. The sinking streams and caves drain to the

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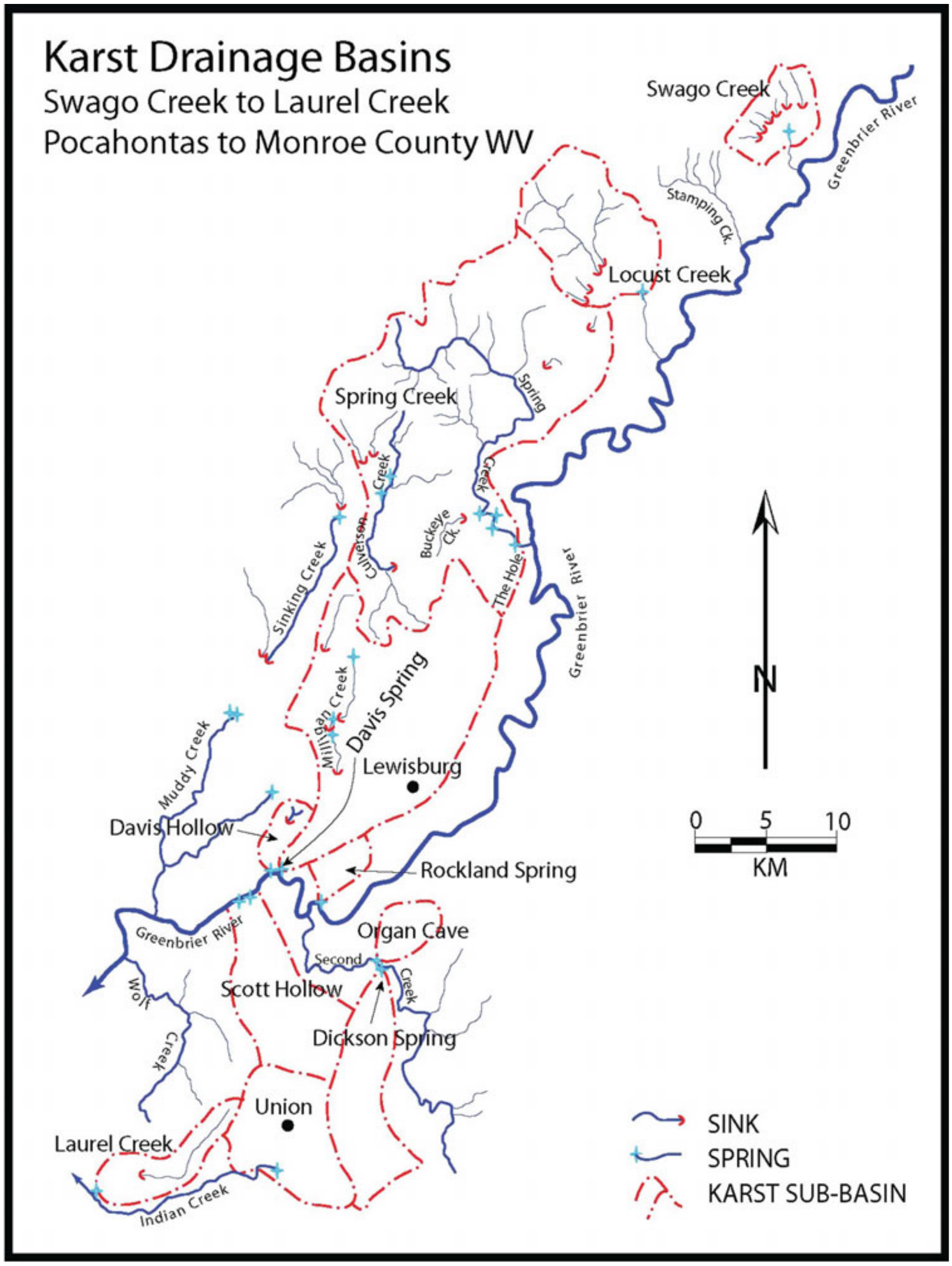

Fig. 3.1 Sketch map showing study area and principal drainage basins

Greenbrier River or its major tributaries of Locust Creek, Spring Creek, and Second Creek for the areas north of Union in Monroe County. Drainage South of Union is to Indian Creek, a tributary to the New River.

The Greenbrier River and Indian Creek represent regional base level for the area. Karst development in the northern part is confined to relatively narrow coves floored by limestone and surrounded by clastic hills. The carbonates thicken, and the width of exposure widens to the south. Karst weathering processes have produced a broad doline plain (Fig. 3.2) with numerous sinking streams, blind valleys, and large springs.

3.2 Climate

All recharge to the aquifer is from precipitation. The humid temperate climate provides relatively uniform rainfall throughout the year. The mean annual temperature at Lewisburg is 50 °F (10 °C) with a mean annual total precipitation of 40 in. (1011 mm) (1981–2010 normals). Average annual snowfall is 32 in. (81 cm). Union is slightly dryer and warmer than Lewisburg while the Swago Creek area is colder and wetter. Marlinton has a mean annual temperature of 47 °F (8.8 °C) and mean annual precipitation of 44 in. (1117 mm) with 40 in. (102 cm) average snowfall.

Fig. 3.2 Aerial photo showing doline plain on mature Greenbrier karst in Monroe County

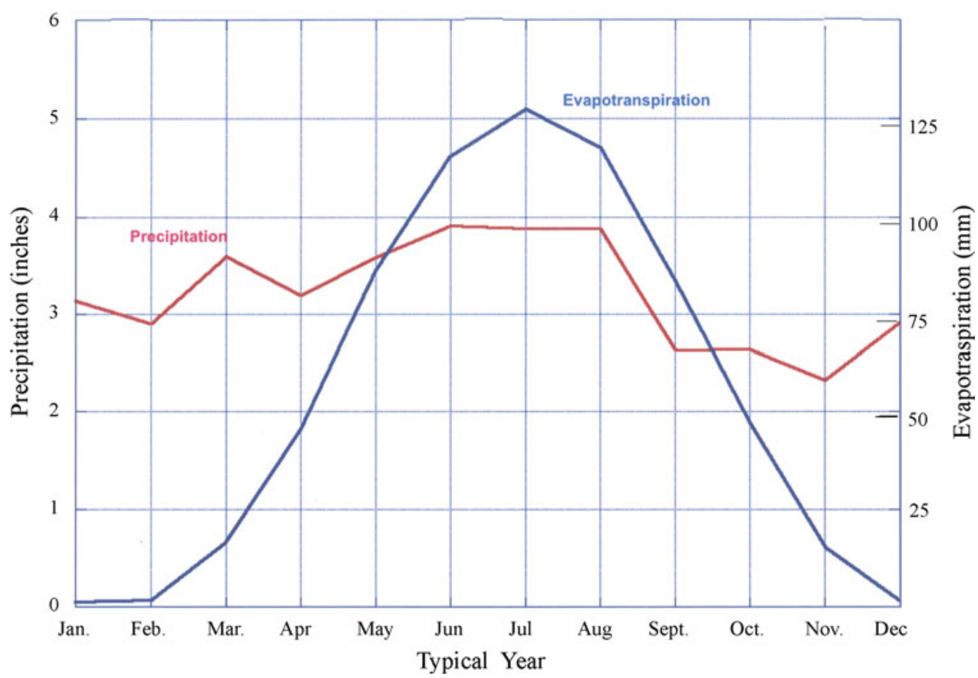

Fig. 3.3 Graph showing monthly precipitation and potential evapotranspiration at Lewisburg

Evapotranspiration exceeds precipitation during the summer months (Fig. 3.3) so stream and spring flows are at a minimum during late summer and early fall. Annual runoff for the Greenbrier River at Buckeye is 22.2 in. (560 mm) and 19.9 in. (505 mm) at Alderson. Roughly, half of the annual precipitation is lost to evapotranspiration.

3.3 Recharge and Yield of the Aquifer

The karst drainage basins are surrounded by clastic rocks but much or all of the flow is subterranean to base-level springs. The discharge at the springs is typically very flashy, and the storm hydrographs are similar to those of nearby surface streams. This is because much of the recharge is from the capture of surface streams at or near the contact with the carbonate outcrop. Recharge is very rapid at the points of capture of surface streams and through the more open drains at the bottom of dolines. Storage within the conduit part of the aquifer is in smaller fractures and the overlying epikarst zone. The spring hydrographs represent a mixture of allogenic recharge from captured surface streams originating on the surrounding clastic units and autogenic recharge captured by the doline plain. A study by Jones and Rauch (1977) showed that most of the springs have a high-flow to low-flow ratio exceeding 100. Discharge from Davis Spring typically ranges from about 7 to over 1000 cfs (0.2–28 cm).

3.4 Water Wells

Water wells that appear to be completed in the limestone may actually be producing from shaley units within the Greenbrier Group or the underlying Maccrady Shale (Ogden 1976; Heller 1980). There are few examples of interchange between water in the conduits and nearby water wells. Water levels in wells are frequently higher than in the conduit part of the aquifer. A study by Demrovsky (2003) examined two cases where water from uncased wells drilled close to cave passages "bursts" into the caves under pressure and created a constant flow or leakage into the cave. Demrovsky termed these "subterranean springs" and found that the constant drainage of the shale aquifer was creating a significant depression in the water table (potentiometric surface) and causing several area wells to go dry. This water was rich in sulfur and iron oxides (ferrihydrite) and of very different chemical quality from streams in the cave. High sulfate concentrations have been reported in scattered water wells in the area (Heller 1980) although these wells appear to be drawing water from sulfur-rich shales (Ogden 1976).

Recharge to the shaley aquifers, especially the Hillsdale limestone/Maccrady Shale, may be from outside of the main areas of the exposed carbonate outcrop (Ogden 1976). The groundwater flow direction for the lower aquifer may not be reflected in the drainage patterns and divides determined using tracer tests in the conduit part of the system. However, a study by Heller (1991) using water levels from wells and MODFLOW to draw the potentiometric surface for the area between Davis Spring and Spring Creek in central Greenbrier County did result in a north–south drainage divide that closely matched the divide predicted by Jones (1973) for the conduit flow part of the aquifer.

3.5 Water Tracing

Some of the earliest water-tracing studies in North America were conducted by Hermine Zotter in the Swago Creek area (Zotter 1965). These tests were planned primarily by cavers from the Pittsburgh Grotto to establish connections between sinking streams and caves and were viewed as an aid to exploring and understanding caves (Fig. 3.4). The tests were qualitative and used packets of activated carbon placed in springs to act as passive collectors for fluorescein sodium dye (CI45350) injected in sinking streams (Fig. 3.5). The use of passive collectors enabled researchers to do weekend trips to the area without the need for constant observation of a number of springs.

The essential procedure was described by Dunn (1957) and used a basic alcohol solution to elute the tracer. The solution was allowed to sit undisturbed for a time period ranging from one-half hour to a couple of days and then visually examined under sunlight or a high-intensity flashlight. A fluorescent-green sheen at the top of the carbon signified a positive test. This system was found to work with other dyes in the Xanthene family (Eosin sodium and Rhodamine WT) so tests starting in the 1980s often used multiple tracers and fluorometers in the analysis to increase the levels of detection and separate the different tracers based on emission wavelength.

The early tests conducted by Zotter in the Swago Creek area typically used 3–8 oz (85–230 g) of fluorescein over distances of one to two miles (1.6–3.2 km). Tests conducted by Jones (1973, 1997) in Greenbrier and Monroe Counties used one to ten pounds of dye over distances of 1–15 miles (1.6–24 km). Many of these tests resulted in visual coloration at the resurgences, and an approximation of the travel time was obtained. Although coloration of springs is discouraged today, one advantage of the high dye concentrations during the early reconnaissance work was that all the springs in a 100-square mile area could be checked from a small plane during a 1-h flight. Visiting all the possible resurgences by road and trail required almost two days of fieldwork.

The early tracer tests in the Swago Creek area were designed primarily to determine the connections between caves, sinking streams, and springs. The tests conducted in the early 70s as part of a US. Geological Survey study (Jones 1973) were an attempt to define drainage divides and determine the catchments for some of the larger springs. The results of the tracer tests were combined with data from cave surveys and plotted on 62,500 scale topographic maps. The

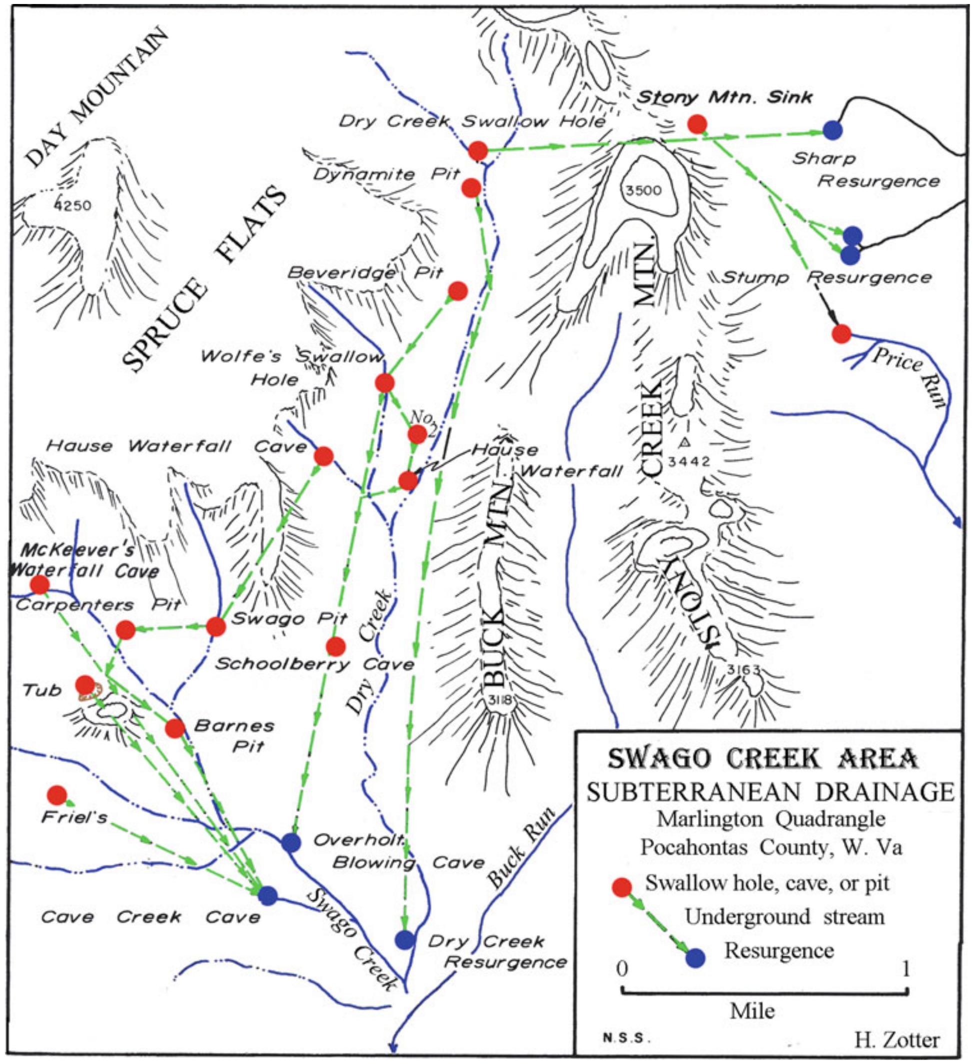

Fig. 3.4 Sketch map showing karst drainage patterns in the Swago Creek area of Pocahontas County (after Zotter 1965)

drainage basins were drawn along the topographic divides on the surrounding clastic rocks and estimated in parts of the limestone outcrop area. Additional tracer tests have been done by different workers over the years, and minor adjustments have been made to the boundaries originally shown in the 1973 publication (Jones 1997; Dasher and Boyer 2000; Tudek 2009).

3.6 Principal Drainage Basins

3.6.1 Swago Creek Basin

The Swago Creek watershed is a karst "cove" of about 4 mi^2 (10 km^2) that contains more than 80 known caves. The rocks are relatively flat lying, but faulting may exert some control over subsurface flow paths. Recharge to the karst aquifer is largely from the capture of surface streams flowing off of the surrounding highlands capped by clastic rocks. Karst groundwater flow is along hydraulic gradients toward the Greenbrier River, but the flow routes are not reflected by the surface topography. This basin has the highest average elevation and greatest relief of the basins described in this volume.

Studies of this area by Zotter (1963, 1965) described the connections between the various stream sinks, caves, and springs (Fig. 3.6). White and Schmidt (1966) used the results of the tracer tests and mapping of caves to define subterranean flow routes and determine drainage divides. They demonstrated that the underground drainage was often unrelated to the surface (or former surface) flow routes. Flow paths that seemed to underdrain the original surface channels were termed "well behaved," and paths that crossed into other basins were "misbehaved." Another concept from this paper was "lost waterfalls" where the cave stream comes to the surface perched on insoluble shale (Fig. 3.7), flows on the surface for some distance, and sinks underground again where it breaches the shale and again reaches the limestone.

Fig. 3.5 Fluorescein sodium dye injected in the stream from Boyds Cave in Monroe County

Streams on the western side of the Swago Creek basin lose water intermittently through gravel-floored stream channels and resurge at Cave Creek Cave spring. This appears to underdrain the dry surface valley and is "well behaved" using the classification of White and Schmidt (1966). Water from Hause Waterfall Cave is pirated from the Dry Creek catchment through Carpenters Cave and is "misbehaved."

Water at the very head of Dry Creek sinks at Dry Creek Swallow Hole and passes eastward under Stony Creek Mountain to resurge at Sharp Resurgence. Dry Creek water sinking slightly to the south at Beveridge Pit more or less underdrains the Dry Creek valley through Wolfs Swallow Hole and resurges at Overholt Blowing Cave.

3.6.2 Upper Spring Creek Locust Creek Cave Basins

The water resurging at Locust Creek Cave (Fig. 3.8) comes from both the Little Levels area west of Hillsboro and a partial contribution from streams sinking on the west side of Droop Mountain. This is a complex basin and was initially studied by Zotter (1965), White and Schmidt (1966). Tracer tests from these early studies showed that water sinking in the bed of Bruffey Creek at the entrance to Bruffey Creek Cave flowed east to Hughes Creek Cave. The water resurges at Upper Hughes Creek where it is perched on the Taggard Shale and flows on the surface for about 600 ft (180 m) until it breaches the shale and sinks again into Lower Hughes Creek Cave. The Bruffey/Hughes water then flows south to resurge at Locust Creek Cave. Also resurging at Locust Creek Cave is water from Grand View Pit and Blue Hole.

The study by Zotter showed that water from Bruffey Creek reached Locust Creek Cave after going through the Hughes Creek Caves, and Hills Creek water was believed to flow under Droop Mountain directly to Locust Creek Cave 1.65 miles to the southeast. Coward (1975) reported tests from Bruffey and Hills Creek Caves with dye recovery in Cutlip and Clyde Cochran Caves with the assumption that all the flow from Hills Creek was to the southwest to Clyde Cochran Cave and then to the east to Locust Creek Cave.

Further studies by Williams and Jones (1983) and Jones (1997) identified a somewhat more complex flow system (Fig. 3.9). Discharge measurements under base flow conditions showed that the flow in Hills Creek alone was greater than the flow in Locust Creek so there were obviously multiple resurgences for this area. Cave exploring and mapping in the Friars Hole system in the 1970s revealed a large master stream passage extending to the southwest toward Spring Creek. A quantitative test (Williams and Jones 1983; Jones 1984) involved the simultaneous injection of an optical brightener in Hills Creek, Rhodamine WT in Cutlip Cave, and fluorescein sodium in Friars Hole cave. The tracer from Hills Creek was recovered at Locust Creek Cave and in Clyde Cochran Cave. However, the tracers injected from Friars Hole and Clyde Cochran Caves were not found at Locust Creek Cave but at JJ Spring on Spring Creek 11 miles (18 km) to the south. Most of the water from Bruffey and Hills Creeks (at least under base-flow conditions) goes through Cutlip, Clyde Cochran, and Friars Hole Caves and ultimately resurges on Spring Creek. Jones (1997) calculated that at low flow only about 5% of the water from Hills Creek went to Locust Creek Cave, and some of the

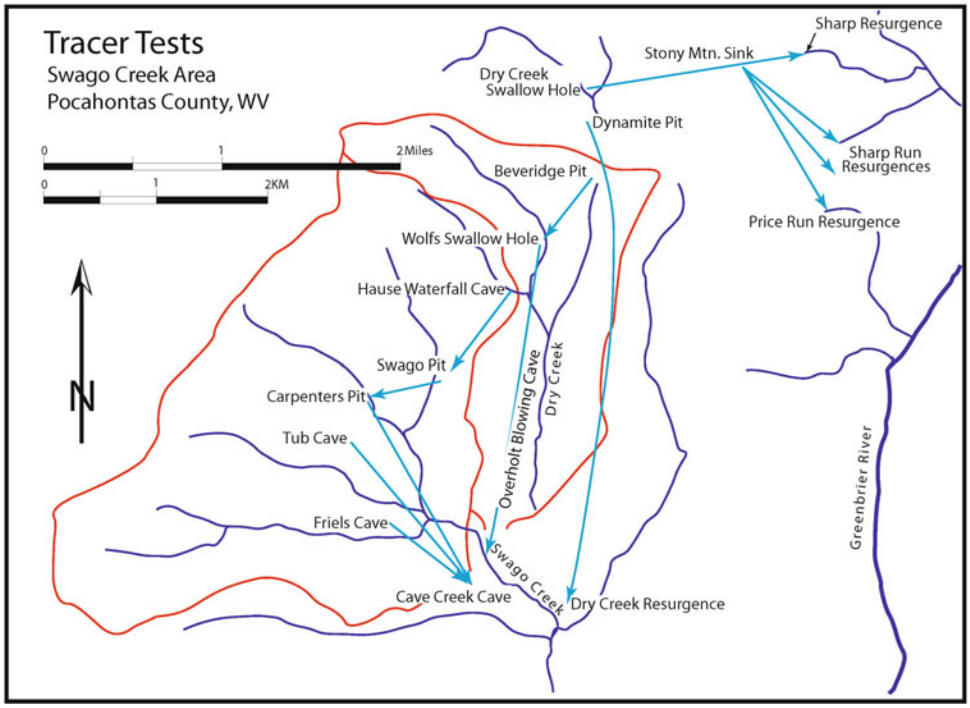

Fig. 3.6 Sketch map showing principle tracer tests in the Swago Creek area (after Zotter 1965)

Fig. 3.7 Lost waterfall on the wall of the Dry Creek Valley, Swago Creek basin. Photo by W. B. White. Used with permission

flow from Bruffey Creek and Rush Run was also diverted south to Spring Creek.

Under low-flow conditions, the bed of Spring Creek is dry for several reaches above the Rt 219 bridge, below the Cannon Hole, and below the upstream entrance to Spur Cave. A 1976 study by Charles and Barbra Williams (personal communication) established the connections between Spring Creek water sinking in the Cannon Hole and the water-filled sinkholes in the Spring Creek floodplain a half mile downstream. These steep-sided sinkholes were termed “floodplain cenotes” by Jones (1997) and are clustered together. The Cannon Hole is an estavelle (Fig. 3.10) that captures the surface flow of Spring Creek during base flow conditions and acts as a spring during periods of high flow. The Spring Creek water reappears in the cenotes and then at JJ Spring on the east side of the surface channel. The water from JJ Spring is on the surface for a couple of hundred feet before entering Spur Cave that acts as a subterranean

Fig. 3.8 Aerial photo of Locust Creek Cave (spring)

meander cutoff and then reappears a mile further downstream. During high-flow conditions, water is flowing in the surface channel, the Cannon Hole is functioning as a spring, and the Circulating Cenote is discharging overflow water as a spring (Figs. 3.11 and 3.12).

3.6.3 Lower Spring Creek Drainage Basin

Karst drainage to Spring Creek from the south (Fig. 3.13) consists primarily of flow from Culverson Creek Cave, Buckeye Creek Cave, and The Hole Cave. Some smaller sinks in the Frankford area also drain northeast to Spring Creek (Jones 1997).

Culverson Creek is fed by streams that flow off of clastic rocks on Cold Knob (Roaring, Little Roaring Creek, and Charley Run), sink near the limestone contact, reappear at springs and flow on the surface for a couple of miles, and sink again at the blind valley entrance to Culverson Creek Cave. The Culverson Creek Water then flows northeast to a series of four springs on the south bank of Spring Creek. The Culverson flow apparently passes underneath the headwaters of Buckeye Creek. A steam sinking near the old Pilgrims Rest Church also resurges with the Culverson Creek water on Spring Creek. The four Culverson resurgences include Matts Black Cave and are clustered along an escarpment on the southwest side of Spring Creek. Tracer tests started from the Culverson and Fullers entrances under low-flow conditions took 14 days to reach Spring Creek (Fig. 3.14).

Wolfe (1973) examined the dry valleys below the sink points and called them “karst sieves” due to the accumulation of clastic sediments left by water percolating through the porous limestone. Rader’s Valley is a large dry valley (Fig. 3.15) that is the former surface course of Culverson Creek.

Buckeye Creek flows to the north and sinks at the blind valley entrance to Buckeye Creek Cave. This passage has been surveyed to Spencer Cave on Spring Creek. Under low-flow conditions, the bed of Spring Creek is dry for some distance above Spencer Cave, and the Buckeye Creek water flows on the surface in the Spring Creek channel to sink at the Cannon Hole. A narrow flat-floored valley locally known as the “Race Track” trends south from the Buckeye Creek Cave entrance. This valley is a closed depression and essentially a small polje (Fig. 3.16) about 1 mile (1.6 km) long and averages 300 ft (110 m) in width. Springer (2002, 2004) conducted detailed studies of the sediments and passage sculpturing in Buckeye Creek to characterize paleoflooding in the basin.

The large contact cave system known as “The Hole” drains north to resurge at two springs on the south side of Spring Creek. The eastern most spring is Burns Cave Number 2, and the second spring is the “Blue Hole” (Fig. 3.17) about one-half mile upstream from the Burns Spring. These springs also drain the Frankford area.

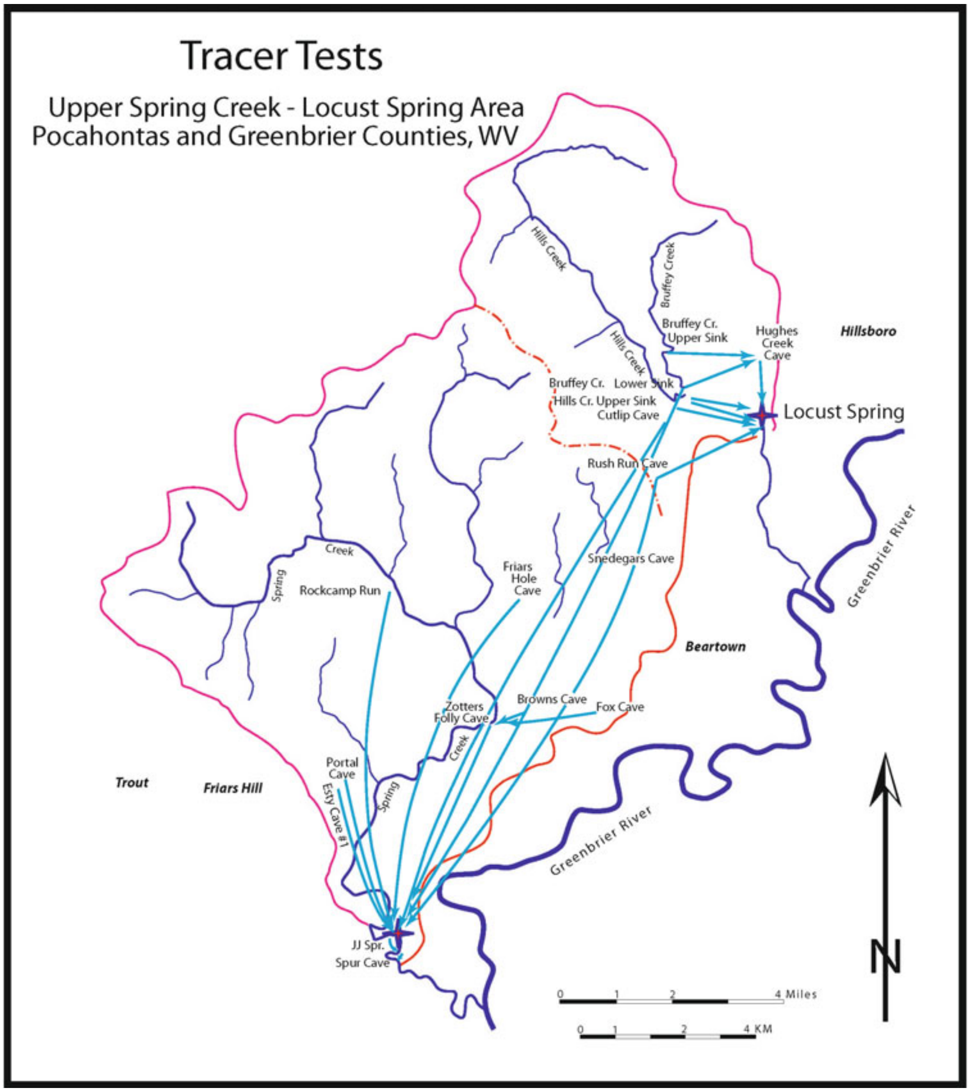

Fig. 3.9 Sketch map showing principle tracer tests in the Upper Spring Creek drainage Basin. Note that flow from Bruffey Creek, Hills Creek, and Rush Run goes to both Spring Creek and Locust Creek. All of the water from Clyde Cochran and Snedegars (Friars Hole) Caves goes to JJ Spring on Spring Creek (after Jones 1997; Dasher and Boyer 2000)

3.6.4 Davis Spring Basin

Davis Spring discharges into the Greenbrier River at Fort Spring southwest of Lewisburg. The spring rises at the base of a cliff (Fig. 3.18) and flows on the surface for about 1000 ft (300 m) to the river. About 80% of the Davis Spring drainage basin (Fig. 3.13) is underlain by carbonates of the Greenbrier Group. Subsurface flow in the basin is to the southwest, and much of it may be along the troughs of several synclines (Tudek 2009). All of the drainage (with the possible exception of extreme flood events) from the basin is through Davis Spring with surface streams flowing off the surrounding clastic highlands and sinking at ponors and blind valleys in the limestone to create a significant volume of allogenic recharge to the spring. Additional recharge to the karst aquifer system occurs through the extensive doline plain developed on the outcrop area of the limestone. Much of the flow through the aquifer occurs through cave passages. Over one hundred caves are known from this basin, and several of these are over 10 miles (16 km) long. Several large contact caves including Ludington, McClung, and Wades caves drain to Davis Spring along with the subsurface runoff from the Lewisburg area. The northern boundary is defined by the Higginbotham Caves west of Frankford. Water from the Higginbotham Number 1 cave reappears in a karst window at Coffman Cave, flows through Savannah Lane Cave, and flows southwest for about 15 miles (24 km) to Davis Spring. Milligan Creek drains the western part of the basin. Milligan Creek is a classic example of an “interrupted” karst stream that sinks and rises at five different points with the final rise at Davis Spring. Most of the tracer tests within this catchment were qualitative, but the straight-line travel velocities for the Lewisburg tests were about 1900 ft (580 m) per day. This basin is described in reports by Jones (1973, 1997) and Tudek (2009).

Fig. 3.10 Barbra Turner Williams adding fluorescein dye to the Cannon Hole estevale at low flow

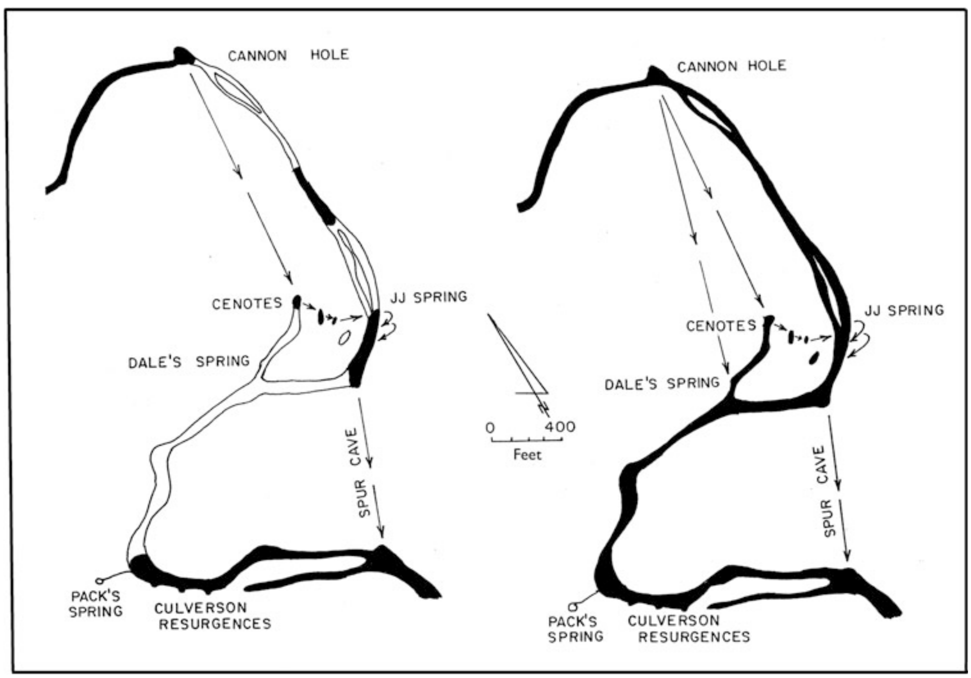

Fig. 3.11 Sketch map of the Spring Creek drainage from the Cannon Hole to JJ Spring (after Williams and Jones 1983; Jones 1997)

The Rockland Indian Spring basin is a small karst catchment that discharges in an alleviated spring alcove directly to the Greenbrier River at Rockland. The basin drains about 3.3 mi^2 (8.5 km^2) and is described by Jones (1973) and Tudek (2009). Watters Cave drains a narrow strip along the southeast edge of the Davis Spring Basin. The cave entrance is just south of the Greenbrier East High School in Fairlea, and the water resurges in a spring above the town of Ronceverte. The catchment is about 1 mi^2 (2.6 km^2).

One full water year (October 1–September 30, 1973) of daily discharge record is available for Davis Spring (Fig. 3.19) and nearby Howards Creek. The record for Davis Spring may slightly underrepresent total discharge because the record is truncated at a maximum flow of 1000 cfs (28.3 cm). The rating curve becomes undefined at high water levels due to backwater from the Greenbrier River at the gauge. Data from the Davis Spring gauging station was used by White (1977) in her study of the effect of karst on flows in Appalachian watersheds.

Fig. 3.12 Aerial photo showing Spring Creek sinking at the Cannon Hole (*far left*), the floodplain cenotes (*center*) and JJ Spring (*far right*)

Paired watershed studies are a common tool in forest and agricultural hydrology for studying the effects of different harvesting or agricultural systems on runoff and water quality from physically similar basins (Woolhiser et al. 1980; Zhang et al. 2001; Schilling 2013). Paired watersheds were also used in a study of the Mammoth Cave area in Kentucky by Hess and White (1989). The Howard Creek catchment is underlain by clastic rocks and is situated just to the east of the Greenbrier River. The Davis Spring catchment lies just to the west of the river. Both Basins are tributary to the Greenbrier River (Fig. 3.20), and the main difference between the basins is bedrock geology. This should appear as differences between the water budgets for the two basins. The basic annual water budget equation is:

$$P = R_s + R_{gw} + E_t \pm \Delta S$$

where P is the annual total precipitation, R_s is surface water runoff, R_{gw} is groundwater runoff, E_t is evapotranspiration, and S is the change in storage between the start and end of the water year. It is generally difficult to separate the surface and groundwater components, and the reported annual runoff for a gauging station is the sum of these components. A rough estimate for gauging stations in West Virginia with significant contributions from karst aquifers is that 85% of the annual runoff is from ground water (Shultz et al. 1995). Note that the water year runs from October 1 to September 30 in an attempt to minimize the changes in storage.

The drainage area for Davis Spring was defined from a series of water-tracing tests (Jones 1997; Tudek 2009). The Davis Spring (USGS Station ID number 03183200) catchment is about 74 mi^2 (192 km^2), and Howard Creek (USGS Station ID Number 03182950) drains an area of 84.5 mi^2 (219 km^2). Precipitation data from the NOAA weather station at Lewisburg, West Virginia, was used to calculate the water budgets for both catchments. The 1973 water year was wetter than the normal precipitation of 40 in. (1030 mm) with 45 in. (1150 mm) of precipitation for the study period. This year had a total of 47 in. (1190 mm) precipitation for the period so runoff from both basins was above the long-term averages. However, this should affect both basins equally so the comparison of water budgets for the two basins (Fig. 3.21 and Table 3.1) should be valid.

The total runoff for the 1973 water year was 21.2 in. (538 mm) from Howard Creek and 25.9 in. (658 mm) from Davis Spring. Runoff at Howard Creek exceeded that from Davis Spring during November (Fig. 3.22) when evapotranspiration was near a minimum but fell below that of the spring during August (Fig. 3.23) when evapotranspiration rates reach a maximum. It is interesting that in terms of area-runoff for November 1972, there was no significant difference between the two basins with Howard Creek having 3.1 in. (78 mm) of runoff and Davis Spring 3.0 in. (76 mm) of runoff. During August 1973, Howard Creek had 0.25 in. (6.3 mm) of runoff compared to 0.47 in. (11.9 mm) at Davis Spring. Adjusted for drainage area, Davis Spring was discharging almost twice as much water as Howard Creek in August.

The paired catchments have the same climate and annual rainfall. The two drainage basins differ fundamentally in only two respects: geology and vegetative cover, so differences in the water budgets should be due to these factors. In terms of the annual water budget, runoff was 22% higher, and evapotranspiration was 18% lower for the karstic catchment. The difference in vegetative cover between the catchment accounts for less than half of the observed differences between the paired basins. The rapid infiltration rates of water from precipitation into the karst aquifer alter the water budget to a certain extent by decreasing evapotranspiration with a corresponding increase in runoff from the karstic catchment.

The analysis of storm hydrographs can also be used to quantify differences between drainage basins (White and White 1974; Mangin 1975; Kresic and Bonacci 2009). A comparison of the storm response for the two basins

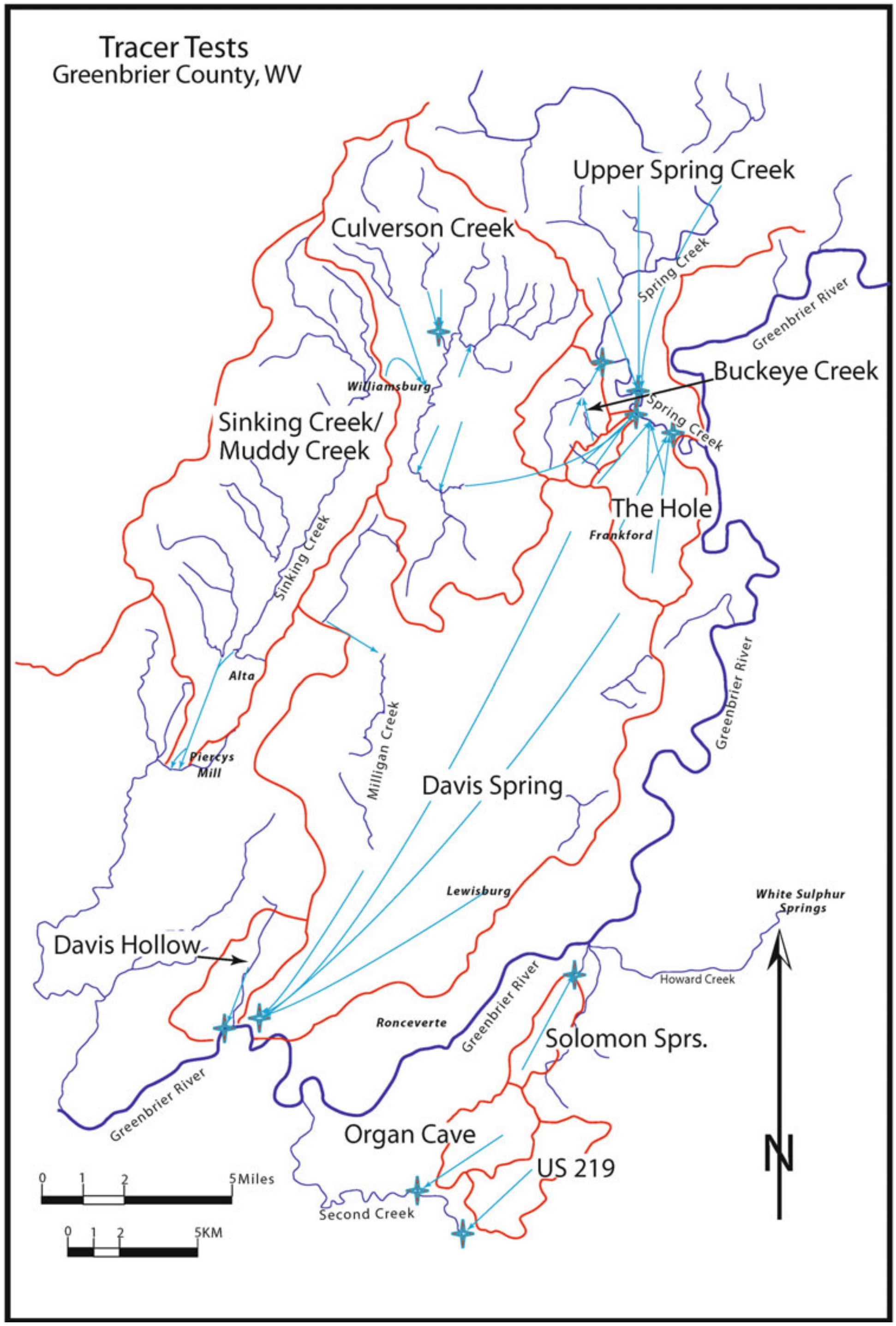

Fig. 3.13 Sketch map of the southern part of the Spring Creek basin (after Jones 1997; Dasher and Boyer 2000), the Davis Spring basin (after Jones 1973; Tudek 2009) and the Organ Cave Basin (after Jones 1997)

shows that Davis Spring typically lags Howard Creek by about a day. The slope of the recession curve following storm events can also show the nature of the release of water from groundwater storage. The recession coefficient (also called the coefficient of discharge) is the slope of the line of the recession hydrograph plotted on semilog paper. Most storm recession hydrographs have distinct periods with different slopes, so this type of analysis is usually conducted on the middle portion of the recession line. To compare recession hydrographs from different catchments, the same time period must be used. A recession analysis from a storm event in August 1973 showed a recession coefficient of 0.0428 for Davis Spring and 0.0625 for Howard Creek (Fig. 3.24). This suggests greater groundwater storage for the Davis Spring Basin or at least a more gradual release of water from storage.

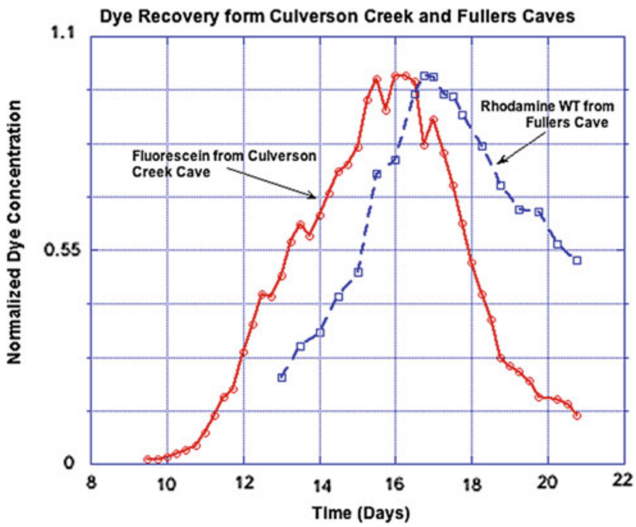

Fig. 3.14 Graph showing recovery of two tracers from the Culverson Creek Cave System at the lower Culverson resurgence on Spring Creek. The trace was started under low-flow conditions on August 24, 2009

3.6.5 Davis Hollow Basin

Davis hollow is underdrained by water sinking at Sinks of the Run Cave, then flowing through General Davis Cave, and resurging in a spring at base level with the Greenbrier River. This 4.4 mi^2 (11.4 km^2) basin is just to the west of Davis Spring at Fort Spring.

3.6.6 Sinking Creek/Muddy Creek Basin

The Sinking Creek/Muddy Creek Basin is northwest of the Davis Spring basin. It is described separately in Chap. 12.

3.6.7 Organ Cave Basin, Soloman Springs, and Route 219 Bridge Basins

These relatively compact catchments are on the southeast side of the Greenbrier River south of Caldwell (Fig. 3.13).

Fig. 3.15 Rader Valley is a large dry valley that was the former course of Culverson Creek when it flowed south to the Greenbrier River near Alderson

Fig. 3.16 Photo of the "Race Track," a polje-like closed depression extending south from the entrance to Buckeye Creek Cave

Fig. 3.17 Blue Hole, one of the two resurgences for the Hole

Water at Soloman Springs flows to the Greenbrier River and is about a mile (1.6 km) from caves.

The headwaters of Second Creek are on Ordovician carbonates on the western base of Peters Mountain in Monroe County. The creek flows to the west across the Greenbrier limestones from about the community of Second Creek to the Greenbrier River, and numerous springs may be found on both banks. A spring on the north bank of Second Creek just upstream of the Route 219 Bridge is fed by water from Helms Cave to the northeast of the spring.

Organ Cave is one of the longest caves in West Virginia and is developed in a synclinal trough at the contact of the Hillsdale limestone and the Maccrady Shale. The catchment for this large cave is only 3.6 mi^2 (9.3 km^2). The cave has a dendritic pattern of passages that drain toward and along the axis of the syncline, and the water resurges at a base-level spring (Fig. 3.25) on Second Creek. The hydrology and ecology of the basin are described in Jones (1988), Culver et al. (1994).

3.6.8 Dickson (Dixon) Spring Basin

Dickson Spring (Fig. 3.26) drains a 27.7 mi^2 (71.7 km^2) catchment in Monroe County on the south side of Second Creek (Fig. 3.27). The water rises at the base of a limestone cliff (Patton limestone) on the western limb of the Hurricane Ridge syncline. During high flow, water also boils up in the spring run about 100 ft (33 m) downstream from the cliff. Flow monitoring reported by Ogden (1976) for the 1974 water year showed a minimum discharge of 6.6 cfs (0.19 cm) and a maximum of 107 cfs (3.0 cm). This represents a high- to low-flow ratio of 16:1 and is much lower than most of the

Fig. 3.18 Aerial photo of Davis Spring

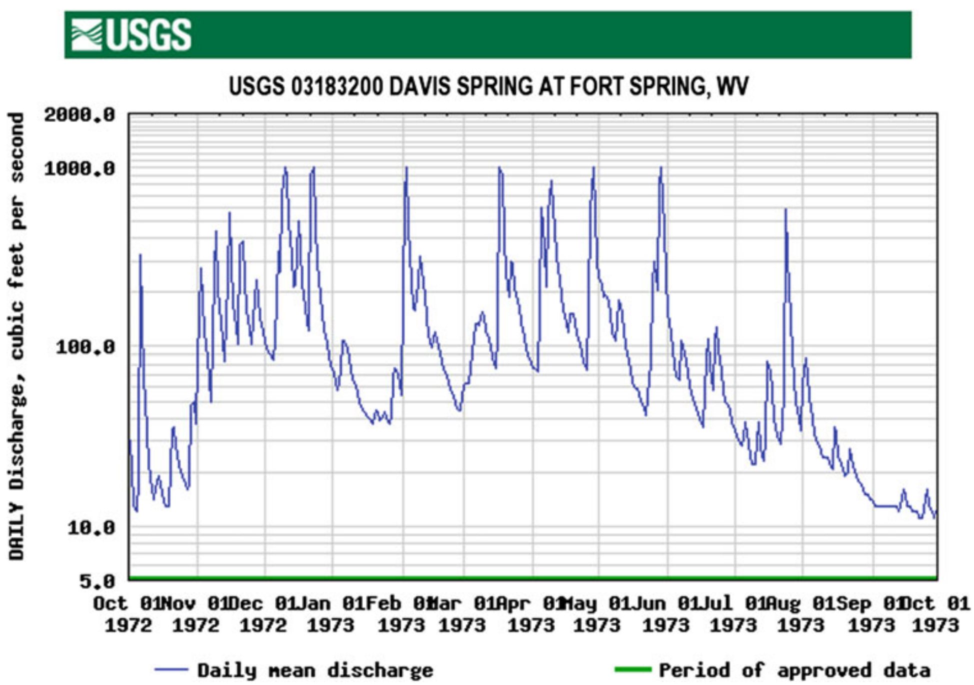

Fig. 3.19 Hydrograph for Davis Spring for the 1973 water year (US Geological Survey)

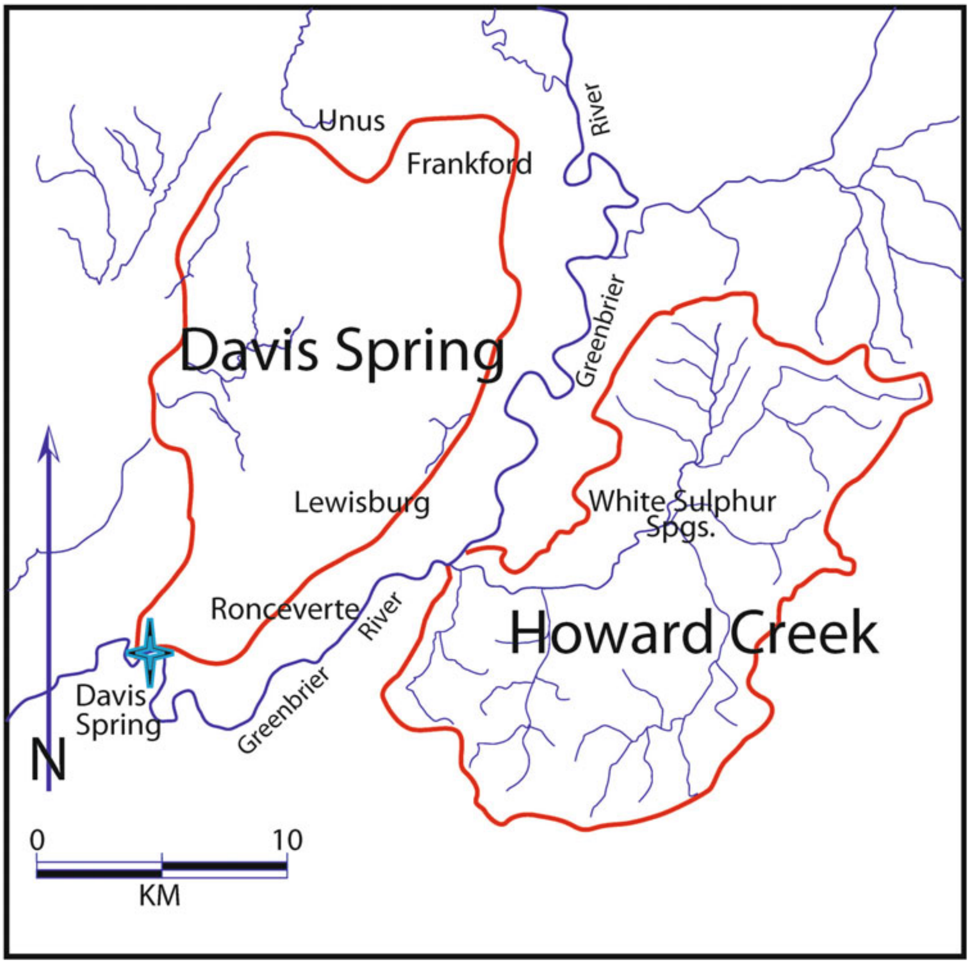

Fig. 3.20 Sketch map showing the Davis Spring and Howard Creek catchments

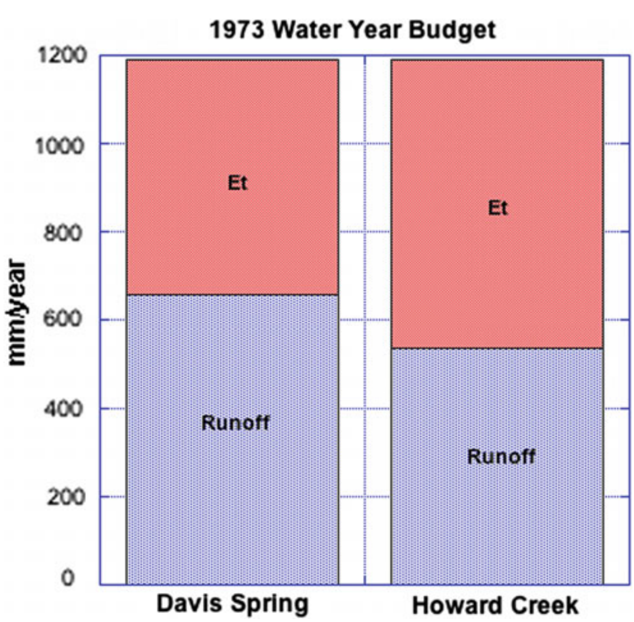

Fig. 3.21 Graph showing water budgets for the 1973 water year

springs draining the Greenbrier karst. The response to individual storm events was also slower than observed for most of the springs in this region. Jones (1997) reported seven tracer tests within this basin. The longest was from Ash Cave about 9 miles (14.5 km) south of the spring. Water sinking in the upper reaches of Burnside Branch goes to Dickson Spring while water from the lower part flows west to Indian Creek through Steel Cave. Straight-line travel velocities for tracer tests in this basin averaged about 2300 ft (700 m) per day.

3.6.9 Scott Hollow Basin

The area around Sinks Grove drains to the north through Scott Hollow Cave and then to several springs in the bed of the Greenbrier River downstream from Fort Spring. Only one small spring (Gloria's Spring) has been observed above normal river pool elevation. The master drain in Scott Hollow Cave is "Mystic River," and most of the underground flow in this basin is channeled through this large passage. The Scott Hollow catchment is about 18.6 mi^2 (48 km^2).

Table 3.1 Comparison of paired karst and non-karst catchments for the 1973 water year

	Davis Spring	Howard Creek
Drainage area	187 km^2	219 km^2
Mean discharge	3.89 cm	3.74 cm
Annual runoff	658 mm	538 mm
Total evapotranspiration	532 mm	652 mm
Q max	28.3 cm[a]	77.0 cm
Q min	0.311 cm	0.207 cm
August runoff	11.9 mm	6.35 mm
Recession coefficient	0.0428	0.0625

[a]Actual maximum exceeds this value

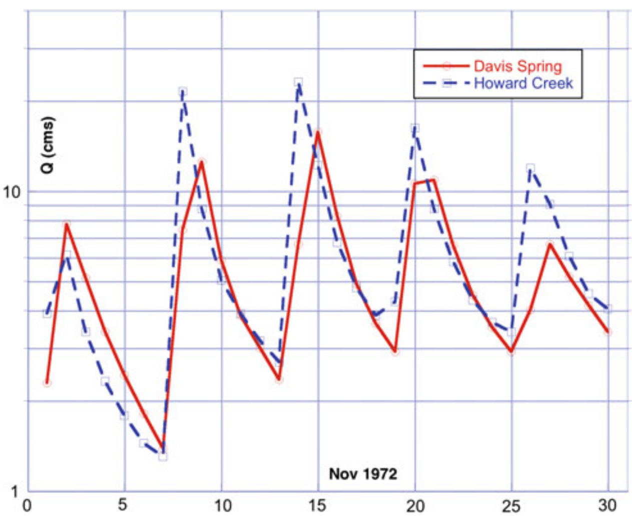

Fig. 3.22 Davis Spring and Howard Creek, November 1972

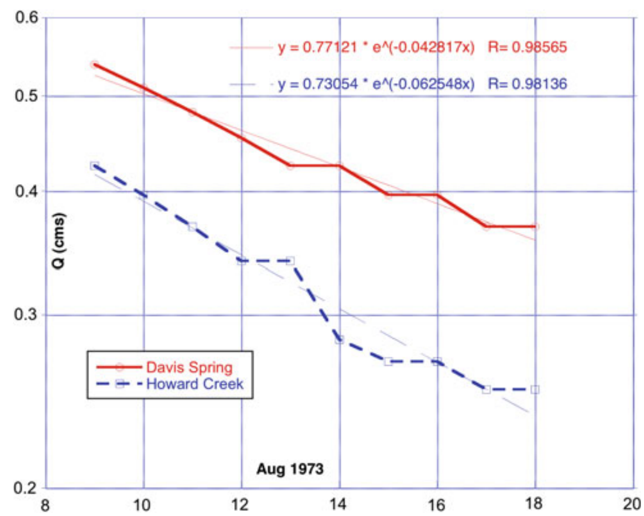

Fig. 3.24 Flow recession curves for Davis Spring and Howard Creek, August, 1973

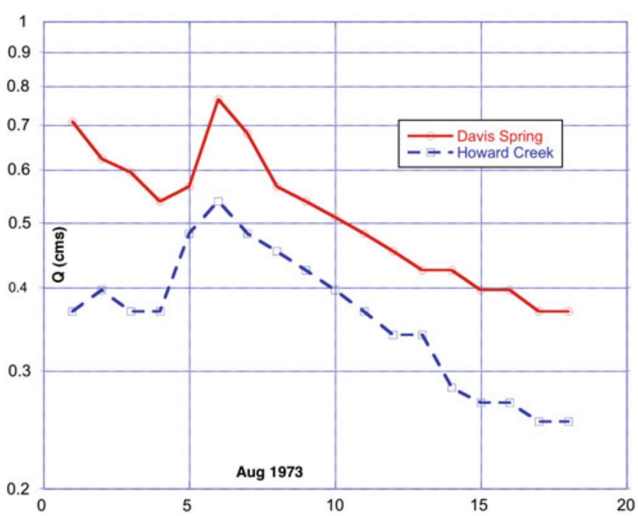

Fig. 3.23 Davis Spring and Howard Creek, August 1973

This basin has been described in reports by Jones (1997), Davis (1999), Bishop et al. (2009), and Demrovsky (2003).

Wolf Creek flows west to the Greenbrier from just to the west of the Scott Hollow basin, but little information is available on the karst of this area. Two large springs are reported near the junction of Broad Run and Wolf Creek at Wolf Creek.

3.6.10 Indian Creek (New River) Basins

Indian Creek rises at a spring (Fig. 3.28) southeast of Union in Monroe County and flows south and then west to join the New River at Indian Mills. Waters sinking in the lower reaches of Burnside Branch and Taggart Branch flow through Steels Cave and then reappear at the head spring of Indian Creek.

Fig. 3.25 Photo showing the Organ Cave resurgence on the north bank of Second Creek

Fig. 3.26 Aerial photo of Dickson Spring

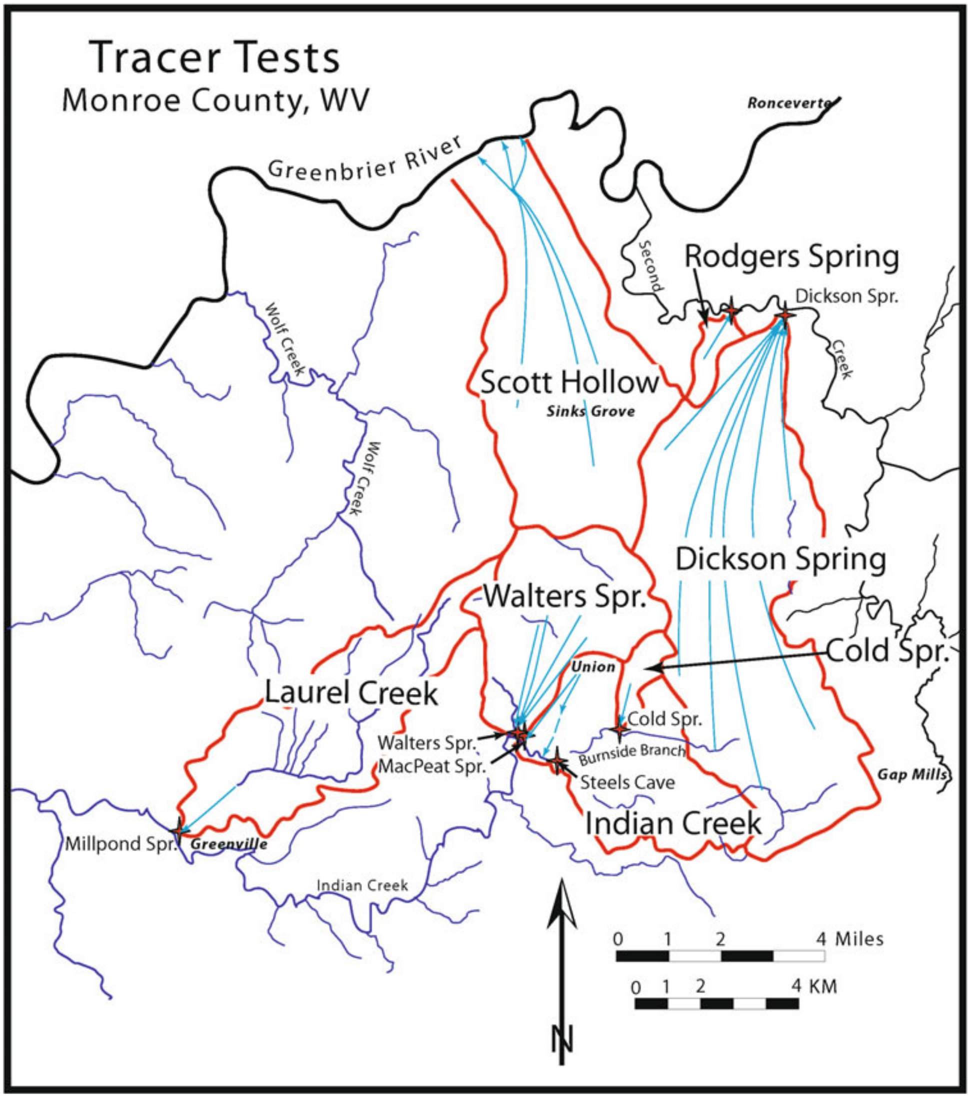

Fig. 3.27 Sketch map showing the Dickson Spring, Scott Hollow, and Indian Creek drainage (after Jones 1997)

MacPeat (McPeak) Spring and Broyles Cave receive water from the town of Union. Water sinking in the town flows partly to Broyles Cave (Broyles Sulfur Spring) and then to a spring on Indian Creek just below the head spring. The rest of the flow from Union reappears at MacPeat Spring on Indian Creek near Salt Sulphur Springs 1.7 miles (2.75 km) southwest of Union (Fig. 3.29). Flow from Union to Broyles probably only occurs during high-flow conditions. This water was contaminated by raw sewage from the town of Union until a treatment plant was constructed 1 mile (1.6 km) south of Union in 1984. The plant discharges into a sinkhole so the (now treated) water still goes to the same springs. The catchment for MacPeat Spring is 2.2 mi^2 (5.7 km^2). Ogden (1976) reported an average flow of 0.2 cfs (5.7 L per second).

Walters Spring is in the next valley about 1000 ft (320 m) to the west of MacPeat Spring but there does not appear to be any connection between the two springs with the MacPeat basin possibly overlying part of the Walters catchment. Both springs (Fig. 3.29) are on the north side of US Route 219. Walters Spring drains an area of 9.6 mi^2 (24.9 km^2) from northwest of Union. Walters Spring is situated at a thrust fault that brings the Greenville Shale to the surface. Ogden (1976) reported discharge ranging from 1 to 125 cfs (0.03–3.54 cm) with an average of about 5–9 cfs (0.14–0.25 cm).

Laurel Creek flows southwest from clastic rocks to the west of the Walters Spring Basin. It sinks at Laurel Creek Cave (Fig. 3.30), flows 0.8 (1.3 km) miles to the southwest to reappear in a karst window at the water entrance to Greenville Saltpeter Cave (Fig. 3.31) and then another 0.5 (0.8 km) miles to reappear at the head of a millpond, and flows to Indian Creek at Greenville. The catchment is 11.9 mi^2 (31 km^2). The Laurel Creek Cave entrance is in a blind valley and is subject to flooding. A study by Groves (1992) traced the changes in the chemistry of the water after the limestone contact was reached and at various locations within the cave and at the spring. Upstream from the limestone contact the Laurel Creek water had a pH of 7.31,

Fig. 3.28 Aerial photo showing the head spring of Indian Creek south of Union

Fig. 3.29 Aerial photo showing Walters and MacPeat Springs near Salt Sulphur Springs

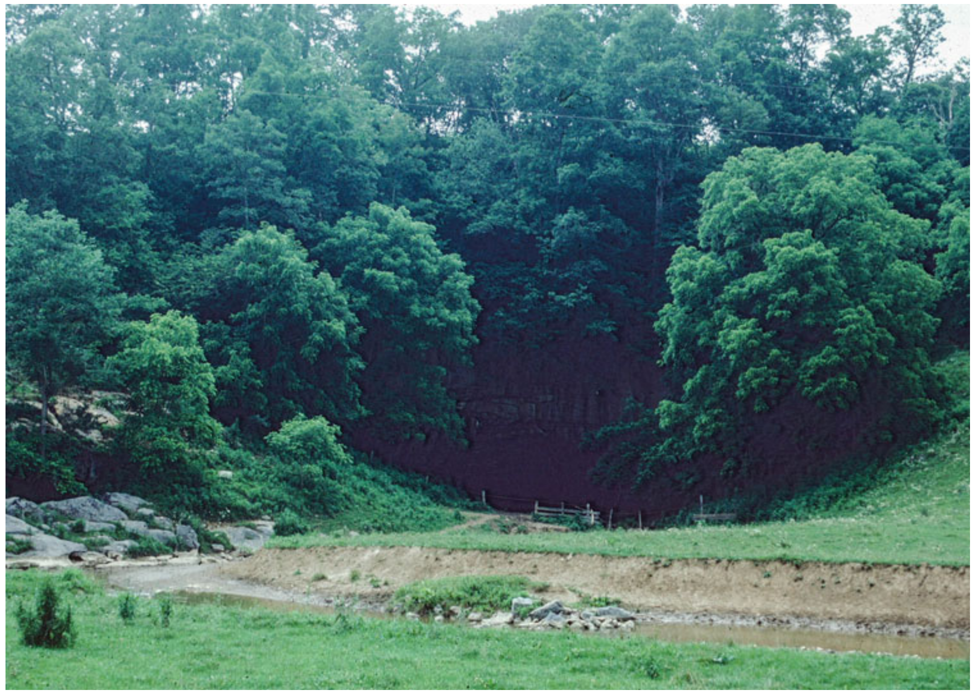

Fig. 3.30 Photo of the blind valley entrance to Laurel Creek Cave

Fig. 3.31 Photo of the water entrance to Greenville Saltpeter Cave

Fig. 3.32 Beginning the tracer test in McClung Cave that established the connection to Sweetwater River, the large stream passage in Maxwelton Cave, May, 2017. Photo by Clifford Lindsay. Used with permission

specific conductance of 62 μS/cm, and hardness of 9 mg/L. Values at the spring were a pH of 7.71, specific conductance of 129 μS/cm, and hardness of 18 mg/L. Discharge was 35 cfs (0.99 cm) at the time of the study.

Indian Draft Cave (Spring) is situated to the northwest of Greenville, and water from Greens Cave has been traced to this tributary of Indian Creek. The catchment is not defined at this time but is probably about 2 mi^2 (5 km^2).

3.7 Closing Comments

In many respects, West Virginia was the caving area of choice for the first caving organization in the Washington DC area. This caving group then became the National Speleological Society, and many of the first scientific investigations of caves and karst features in the USA were conducted in West Virginia during the 1940s. Work on the hydrology of the karst areas began in earnest in the early 1960s with the introduction of passive detectors (carbon traps) to monitor springs for the passage of fluorescent dyes. The area of the Greenbrier limestone karst covered in this book has now been extensively studied by cavers, professional hydrologists, and graduate students from various universities (especially McMaster University in Canada, West Virginia University, and the University of Akron). The area is currently being actively explored by cavers and scientists from many disciplines, and a more detailed picture of the hydrology will continue to evolve. The discovery of a large stream passage in the western end of Maxwelton Cave has once again provided incentive to continue the search for the elusive master drain that must underlie the lower portion of the Davis Spring Basin. A tracer test (Fig. 3.32) from McClung Cave to Maxwelton Cave conducted in May, 2017, established the potential for much additional cave passage and a more detailed picture of the conduit flow system on at least the eastern side of the Davis Spring Basin. If a connection between the western drainage (Higginbotham and Savannah Lane Caves) and the eastern contact caves can be established, a sizable underground river passage should be the reward.

Acknowledgements Many cavers and researchers have contributed to our present understanding of the hydrology of this complex karst area. The references should lead to more detailed information on the tracer tests and hydrogeologic studies summarized in this chapter. Special thanks are due to Roger Baroody for drafting the maps and illustration in this chapter.

References

Bishop, M.R., I.D. Sasowsky, and W.K. Jones. 2009. Preliminary water quality and boundary redefinition of the Scott Hollow drainage basin, Appalachian Plateau. *Proceedings of the 15th International Congress of Speleology* 3: 1422–1427.

Coward, J.H.C. 1975. Paleohydrology and streamflow simulation of three karst basins in southereastern West Virginia, USA, 303pp. PhD Thesis, McMaster University.

Culver, D.C., W.K. Jones, D.W. Fong, and T.C. Kane. 1994. Organ Cave karst basin. *Groundwater Ecology*, 451–473. San Diego: Academic Press.

Dasher, G.R., and D.G. Boyer. 2000. Recent Spring Creek area dye tracings Greenbrier County, West Virginia. *WV Speleological Survey Monograph* 2: 28pp.

Davis, S.B. 1999. Aquifer development in folded and fractured limestone: The Scott Hollow drainage basin, Monroe County, West Virginia, 210pp. MS Thesis, University of Akron.

Demrovsky, N. 2003. The effects of well drilling on karst aquifers in Monroe and Greenbrier County, West Virginia, 198pp. MS Thesis, University of Akron.

Dunn, J.R. 1957. Stream tracing. *National Speleological Society Mid-Appalachian Region Bulletin* 2: 3–9.

Groves, C.G. 1992. Geochemical and kinetic evolution of a karst flow system: Laurel Creek, West Virginia. *Ground Water* 30 (2): 186–191.

Heller, S.A. 1980. A hydrogeologic study of the Greenbrier limestone karst of central Greenbrier County, West Virginia, 421pp. PhD Thesis, West Virginia University.

Heller, S.A. 1991. An attempt to model an Appalachian karst aquifer using MODFLOW. Appalachian Karst. *Proceedings of the Appalachian Karst Symposium*, 177–185. Huntsville: National Speleological Society.

Hess, J.W. and W.B. White. 1989. Water budget and physical hydrology. In *Karst Hydrology*, ed. W.B. White and E.L. White, 105–126. New York: Van Nostrand Reinhold.

Jones, W.K. 1984. Analysis and interpretation of data from tracer tests in karst areas. *National Speleological Society Bulletin* 46 (2): 41–47.

Jones, W.K. 1973. *Hydrology of Limestone Karst in Greenbrier County, West Virginia*, vol. 36, 49pp. West Virginia Geological and Economic Survey Bulletin.

Jones, W.K. 1981. A karst hydrology study in Monroe County, West Virginia. *Eighth International Congress of Speleology Proceedings* 1: 345–347.

Jones, W.K. 1988. Hydrology. In *Caves of the Organ Cave Plateau*, ed. P.J. Stevens, vol. 9, pp. 40–48. Barrickville, WV: West Virginia Speleological Survey Bulletin.

Jones, W.K. 1997. *Karst Hydrology Atlas of West Virginia*, vol. 4, 111pp. Karst Waters Institute, Special Publication.

Jones, W.K., and H.W. Rauch. 1977. Discharge characteristics of karst drainage basins in Greenbrier County, WV. In *Karst hydrogeology*, ed. J.S. Tolson, and F.L. Doyle, 391–398. Huntsville: University of Alabama Press.

Jones, W.K., and J.A. Tudek, Jr. 2012. Hydrology of the Greenbrier Valley karst region. In *Maya Con 2012, Greenbrier Valley, West Virginia*, ed. M. Adler, 94–103. Huntsville: National Speleological Society.

Kresic, N., and O. Bonacci. 2009. Spring discharge hydrograph. In *Groundwater hydrology of Springs*, ed. N. Kresic, and Z. Stevanovic, 129–163. Burlington, MA: Butterworth-Heinemann.

Mangin, A. 1975. Contribution a l'etude hydrodynamique des aquifers karstiques. DES Thesis, University of Dijon, France.

Ogden, A.E. 1976. The hydrogeology of the central Monroe County karst, 263pp. Ph.D. Thesis, West Virginia University.

Schilling, K.E., C.S. Jones, and A.J. Seeman. 2013. How paired is paired? *Journal of Environmental Quality* 42: 1412–1421.

Shultz, R.A., W.A. Hobba, Jr., and M.H. Kozar. 1995. Geohydrology, water availability, and water quality of Berkeley County, WV, with emphasis on the carbonate area. *U.S. Geological Survey, Water Resources Investigations Report* 93: 88pp.

Springer, G.S. 2002. Caves and their potential use in paleoflood studies. In *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology,* ed. K. House, R. Webb, V.R. Baker, D.R. Levish. Water Sciences and Applications, vol. 5, pp. 329–344. Washington, DC. American Geophysical Union.

Springer, G.S. 2004. A pipe-based, first approach to modeling closed conduit flow in caves. *Journal of Hydrology* 289: 178–189.

Tudek, J.K. Jr. 2009. Further hydrogeologic investigations in the Davis Spring drainage basin, Greenbrier County, West Virginia. MS Thesis, West Virginia University.

White, E.L. 1977. Sustained flow in small Appalachian watersheds underlain by carbonate rocks. *Journal of Hydrology* 32: 71–86.

White, W.B., and V.A. Schmidt. 1966. Hydrology of a karst area in east-central West Virginia. *Water Resources Research* 2 (3): 549–560.

White, E.L., and W.B. White. 1974. Analysis of spring hydrographs as a characterization tool for karst aquifers. P*roceedings of the 4th Conference on Karst Geology and Hydrology*, 103–106. Morgantown, WV: West Virginia Geological Survey.

Woolhiser, D.A., D.C. Goodrich, W.E. Emmerich, and T.O. Keefer. 1980. Hydrologic effects of brush to grass conversion. In *Symposium Proceedings of IR Conference Watershed Management*, 293–302. Durango, CO. www.tucson.ars.ag.gov/unit/Publications/PDFfiles/809.pdf.

Williams, C.F., and W.K. Jones. 1983. Hydrology of the upper Spring Creek karst drainage basin, West Virginia (abstract). Huntsville, AL: Proceedings of the 1983 National Speleological Society Convention.

Wolfe, T.E. 1973. Sedimentation in karst drainage basins along the Allegheny escarpment in southeastern West Virginia. PhD Thesis, Geography, McMaster University.

Zhang, L., W.R. Dawes, and G.R. Walker. 2001. Response of mean annual evapotranspiration to vegetation changes at catchment scale. *Water Resources Research* 37 (3): 701–708.

Zotter, H. 1963. Stream tracing techniques and results—Pocahontas and Greenbrier Counties, West Virginia. *NSS News* 21 (10): 136–142.

Zotter, H. 1965. Stream tracing techniques and results—Pocahontas and Greenbrier Counties, West Virginia—Part 2. *NSS News* 23 (12): 169–177.