# Hydrogeologic Characterization and Methods Used in the Investigation of Karst Hydrology

By Charles J. Taylor and Earl A. Greene

Chapter 3 of

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## Introduction

Recharge to and discharge from ground water can be measured or estimated over a wide range of spatial and temporal scales in any hydrogeologic setting (National Academy of Sciences, 2004). Difficulties often arise in making these measurements or estimates because of insufficient knowledge of the processes involved in the transfer of water fluxes, inadequate characterization of the hydrogeologic framework in which they occur, and uncertainties in the measurements or estimates themselves. These difficulties may be magnified considerably in complex hydrogeological settings such as karst.

Karst is a unique hydrogeologic terrane in which the surface water and ground water regimes are highly interconnected and often constitute a single, dynamic flow system (White, 1993). The presence of karst usually is indicated by the occurrence of distinctive physiographic features that develop as a result of the dissolution of soluble bedrock such as limestone or dolostone (Field, 2002a). In well-developed karst, these physiographic features may include sinkholes, sinking (or disappearing) streams, caves, and karst springs. The hydrologic characteristics associated with the presence of karst also are distinctive and generally include: (1) internal drainage of surface runoff through sinkholes; (2) underground diversion or partial subsurface piracy of surface streams (that is sinking streams and losing streams); (3) temporary storage of ground water within a shallow, perched *epikarst* zone; (4) rapid, turbulent flow through subsurface pipelike or channellike solutional openings called *conduits*; and (5) discharge of subsurface water from conduits by way of one or more large perennial springs (fig. 1).

A karst aquifer can be conceptualized as an open hydrologic system having a variety of surface and subsurface input, throughput, and output flows, and boundaries defined by the catchment limits and geometry of conduits (Ford and Williams, 1989). The hydrogeologic characteristics of karst aquifers are largely controlled by the structure and organization of the conduits, the development of which generally acts to short-circuit surface drainage by providing alternative subsurface flow paths that have lower hydraulic gradients and resistance (White, 1999). Conduits are a third (tertiary) form of permeability that is distinctive from, yet interconnected with, the permeability provided by intergranular pores (bedrock matrix) and fractures. Because of the interconnection of matrix, fracture, and conduit permeability, karst aquifers are extremely heterogeneous compared to most granular and many fractured-rock aquifers and have hydraulic properties that are highly scale dependent and temporally variable (table 1).

Because of these unique hydrogeologic characteristics, data requirements for the hydrogeologic characterization of karst aquifers are somewhat more intensive and difficult to obtain than those for aquifers in most other types of hydrogeologic settings (Teutsch and Sauter, 1991). Wherever karst features are present, the water-resources investigator must anticipate the presence of a flow system that cannot be completely characterized by using conventional hydrogeologic methods such as potentiometric mapping or hydraulic tests of observation wells, by numerical modeling, or by using a study approach that treats ground water and surface water as separate hydrologic regimes (White, 1993). In karst terranes, a greater emphasis must generally be placed on the identification of hydrologic boundaries and subsurface flow paths, contributions of water from various recharge sources, and the structural and hydraulic properties of conduits. The acquisition of these data typically requires a multidisciplinary study approach that includes using more specialized investigation methods such as water-tracing tests and the analysis of variations in spring discharge and water chemistry (White, 1993; Ford and Williams, 1989).

This chapter presents an overview of methods that are commonly used in the hydrogeologic investigation and characterization of karst aquifers and in the study of water fluxes in karst terranes. Special emphasis is given to describing the techniques involved in conducting water-tracer tests using fluorescent dyes. Dye-tracer testing is a method successfully used in the study of karst aquifers in the United States and elsewhere for more than 30 years (Käss, 1998). However, dye-tracing techniques generally are not taught at the collegiate undergraduate or graduate level, lack a set of formalized peer-reviewed procedures, and sometimes are difficult to research because case studies often are reported in lesser-known publication venues outside the realm of mainstream professional journals (Beck, 2002). Dye-tracer test procedures described herein represent commonly accepted practices derived from a variety of published and previously unpublished sources. Methods that are commonly applied to the analysis of karst spring discharge (both flow and water chemistry) also are reviewed and summarized.

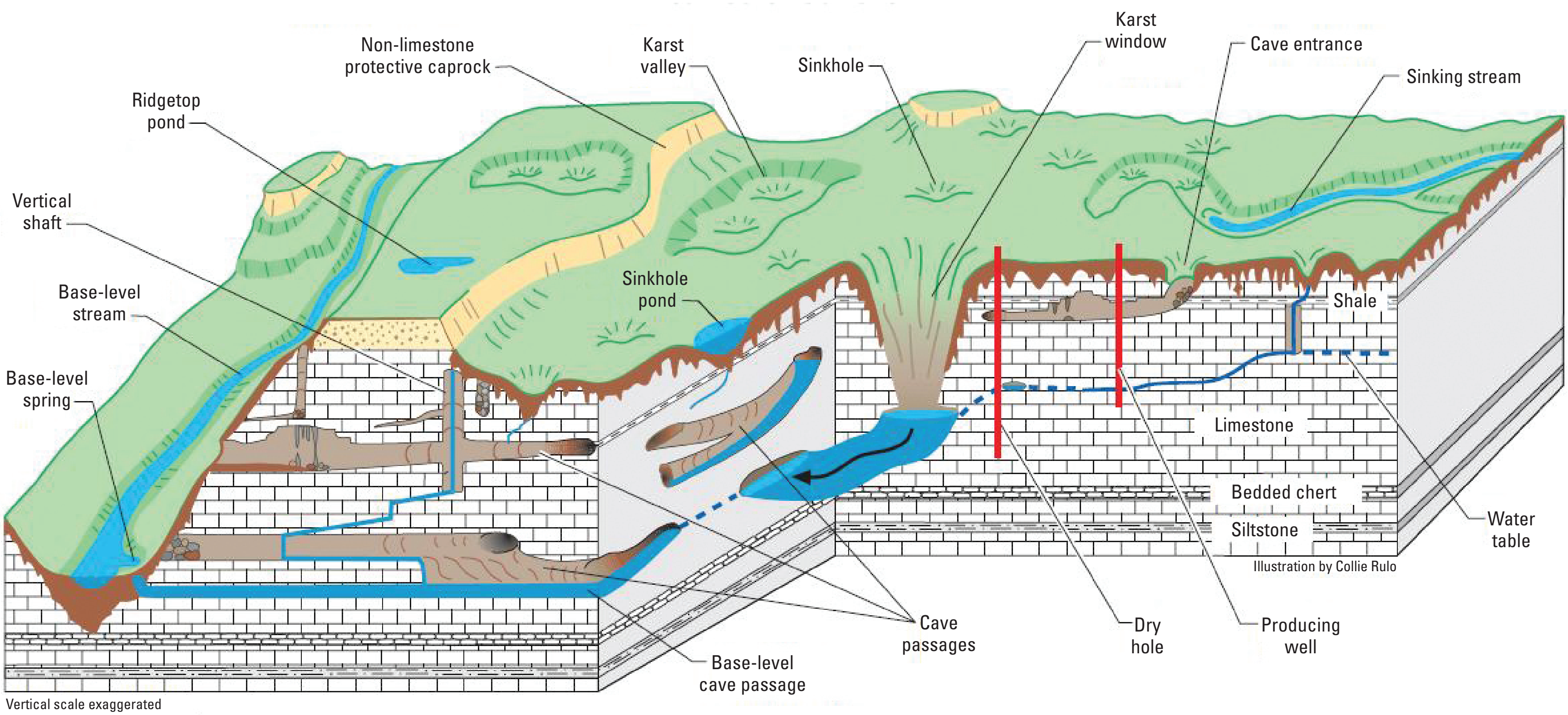



**Figure 1.** Physiographic and hydrologic features typical of a well-developed karst terrane (modified from Currens, 2001, Kentucky Geological Survey, used with permission).

**Table 1.** Comparison of various hydrogeologic properties for granular, fractured rock, and karst aquifers (ASTM, 2002).

| Aquifer characteristics | Aquifer type | | |
|---|---|---|---|
| | Granular | Fractured rock | Karst |
| Effective porosity | Mostly primary, through intergranular pores | Mostly secondary, through joints, fractures, and bedding plane partings | Mostly tertiary (secondary porosity modified by dissolution); through pores, bedding planes, fractures, conduits, and caves |
| Isotropy | More isotropic | Probably anisotropic | Highly anisotropic |
| Homogeneity | More homogeneous | Less homogeneous | Non-homogeneous |
| Flow | Slow, laminar | Possibly rapid and possibly turbulent | Likely rapid and likely turbulent |
| Flow predictions | Darcy's law usually applies | Darcy's law may not apply | Darcy's law rarely applies |
| Storage | Within saturated zone | Within saturated zone | Within both saturated zone and epikarst |
| Recharge | Dispersed | Primarily dispersed, with some point recharge | Ranges from almost completely dispersed- to almost completely point-recharge |
| Temporal head variation | Minimal variation | Moderate variation | Moderate to extreme variation |
| Temporal water chemistry variation | Minimal variation | Minimal to moderate variation | Moderate to extreme variation |



## Hydrogeologic Characteristics of Karst

A number of important characteristics of the physical hydrogeology of karst are summarized here for the benefit of readers less familiar with karst and with the differences between karst aquifers and aquifers in other hydrogeologic settings. The subject of karst hydrogeology involves a wide variety of geomorphologic, geologic, hydrologic, and geochemical topics that are beyond the scope of this report. White (1993, 1999) provides good overviews of karst hydrology and the methods typically used in its study. Other good sources of information about karst include textbooks written by Bogli (1980), White (1988), and Ford and Williams (1989); compendiums edited by Klimchouk and others (2000), and Culver and White (2004); and the proceedings of various karst conferences held in the United States from 1986 to 2005 (National Water Well Association, 1986, 1988; National Ground Water Association, 1991; Beck, 1995, 2003; Beck and Stephenson, 1997; Beck and others, 1999; and Kuniansky, 2001, 2002, 2005).

Many geological and hydrologic factors influence the development of karst, and not all karst features are present or developed to the same extent in every karst terrane. The information presented in this report best describes the hydrogeologic characteristics of fluviokarst and doline karst, which are common and widespread types of karst terranes in the United States (White, 1999). The term *fluviokarst* is used to describe a karst landscape in which the dominant physical landforms are valleys initially cut by surface streams that have been partly or completely diverted underground by subsurface conduit piracy (Field, 2002a). This type of karst is often typified by carbonate rocks that have low intrinsic permeability and is common of karst developed in Paleozoic limestones in the Interior Low Plateaus and Appalachian regions of the Eastern United States. The term *doline* karst describes karst landscape in which surface streams are almost entirely absent, and almost all surface drainage is captured and drained internally by closed sinkhole depressions. This type of karst is typical of carbonate rocks that have high intrinsic permeability, such as the Cenozoic limestones in the Atlantic coastal regions, and includes the well-known Floridan aquifer system. In reality, the physical and hydrologic distinctions between fluviokarst and doline karst are not always clearly defined, and many karst terranes have characteristics common to each.

## Conduits and Springs

The most distinctive feature of karst aquifers are the typically dendritic or branching networks of conduits that meander among bedding units, join together as tributaries, and increase in size and order in the downstream direction (Palmer, 1991). In the simplest terms, these conduit networks grow by way of a complex hydraulic-and-chemical feedback loop, in which the basic steps are: conduit growth and enlargement → increased hydraulic capacity → increased discharge → enhanced dissolution and physical corrosion → additional conduit enlargement → subsurface piracy of flows in smaller conduits by the larger conduits. In this process, the largest conduits act as master drains that locally alter the hydraulic flow (or equipotential) field so as to capture ground water from the surrounding aquifer matrix, the adjoining fractures, and the smaller nearby conduits (Palmer, 1991, 1999; White and White, 1989) (fig. 2). Depending on their sizes (hydraulic capacity) and organization (interconnection), conduit networks are capable of discharging large volumes of water and sediment rapidly through a karst aquifer (White, 1993). Flow velocities in well-developed and well-integrated conduit networks that range on the order of hundreds to thousands of feet per day are not uncommon (White, 1988).

Karst springs are the natural outlets for water discharging from conduit networks (fig. 3). They typically are developed at a local or regional ground-water discharge boundary—that is, at a location of minimum hydraulic head in the aquifer—often at or near the elevation of a nearby base-level surface stream (White, 1988). The tributary system of conduit drainage typically

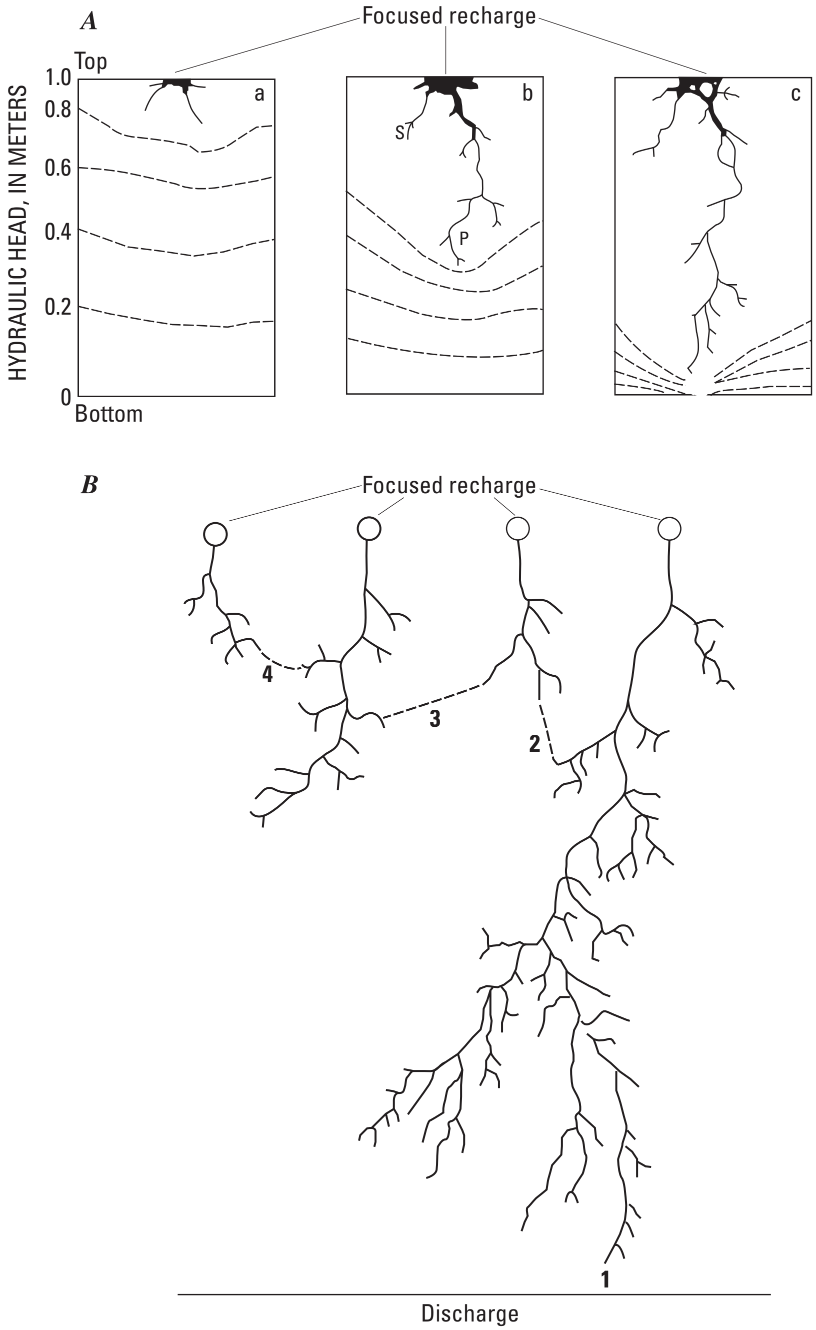



**Figure 2.** *A*, Diagram showing competitive growth of conduits and distortion of hydraulic flow field: (a) initiation of recharge, (b) change in hydraulic gradient in response to propagation of faster growing primary (P) conduit and slower growing secondary (S) conduit, (c) primary conduit breaks through to discharge boundary, slowing or inhibiting growth of secondary conduit. *B*, Sequence of development of integrated drainage network due to faster growth and breakthrough by primary conduit (1) and subsequent capture of flow and linking of secondary conduits (2–4). (Modified from Ford, 1999, fig. 8.) (Copyright Karst Waters Institute and Dr. Derek Ford, used with permission.)


**Figure 3.** Photographs showing a variety of physical outlets for karst springs: *A*, Orangeville Rise, southern Indiana; *B*, Whistling Cave Spring, southern Indiana; *C*, Rocky Spring, central Kentucky; *D*, Head-of-Doe-Run Spring, central Kentucky. (Photographs by Charles J. Taylor, U.S. Geological Survey.)

developed in most karst aquifers yields convergent flow to a trunk conduit that discharges through a single large spring (White, 1999). Many karst aquifers, however, have a distributary flow pattern where discharge occurs through multiple spring outlets. This distributary flow pattern generally occurs where there has been enlargement of fractures and smaller conduits located near a stream discharge boundary, where collapse or blockage of an existing trunk conduit or spring has resulted in shifting of flow and development of alternative flow paths and outlets, or where subsurface conduit piracy has rerouted preexisting conduit flow (Quinlan and Ewers, 1989).

Traditionally, springs are classified on the basis of discharge per Meinzer's scale (Meinzer, 1927) and are otherwise characterized on the basis of physical appearances and whether or not the discharge occurs under artesian or gravity flow (open-channel) conditions (U.S. Geological Survey, 2005). From a flow-system perspective, it may be more useful to classify karst springs according to their hydrologic function as outlets for conduit networks (Worthington, 1991, 1999). In most karst aquifers, one or a few perennial springs, called *underflow springs*, carry the base-flow discharge of conduits (Worthington, 1991). The elevation of the underflow springs exerts much control on the elevation of the water table at the output boundary of the karst aquifer, whereas the matrix hydraulic conductivity and the conduit hydraulic capacity determines the slope of the water table upstream and its fluctuation under differing hydrologic conditions (Ford and Williams, 1989). Other intermittent springs, called *overflow springs*, function as spillover outlets during periods of high discharge. Overflow springs are essentially a temporal form of distributary discharge. As conduits evolve through time and as base levels and water tables are lowered, the upper parts of the karst aquifer may be progressively drained and higher level conduits abandoned (Hess and White, 1989). During high-flow conditions, these higher level conduits may be reactivated and discharge through overflow springs now located at the outlets of former underflow springs.

## Karst Recharge

Karst terrane is unique in having multiple sources of recharge that vary considerably in terms of water residence time and in the timing and amounts of water contributed to the conduit network. Sources of karst recharge are categorized as *concentrated* or *diffuse*, and as either *autogenic* or *allogenic* depending, respectively, on whether the recharge originates as precipitation falling on karstic or nonkarstic terrane (Gunn, 1983). These distinctions are important because the relative proportion of concentrated to diffuse recharge generally dictates the distribution and linking together of conduits, and the timing and relative contributions of water fluxes from allogenic and autogenic sources significantly affects the variability in spring discharge and water chemistry (Ford and Williams, 1989).

A cross-sectional diagram of the major sources of recharge that contribute to a typical karst flow system is shown in figure 4. A major source of *concentrated allogenic* recharge to many karst aquifers is water contributed by sinking or losing streams that originate as normal gaining streams in nonkarstic borderlands. A major source of *concentrated autogenic* recharge is surface runoff funneled into sinkhole depressions, which may drain rapidly to the subsurface through throatlike openings called *swallets* or may drain relatively slowly by percolation through a mantle of soil or alluvium. *Diffuse allogenic* recharge may be contributed by interaquifer transfer of water from nonkarstic aquifers, but a more common source is water that drains down the walls of unsaturated (vadose) zone shafts—vertical or near-vertical conduit passages—where karstic rocks are overlain by nonsoluble caprocks such as sandstone (Gunn, 1983).

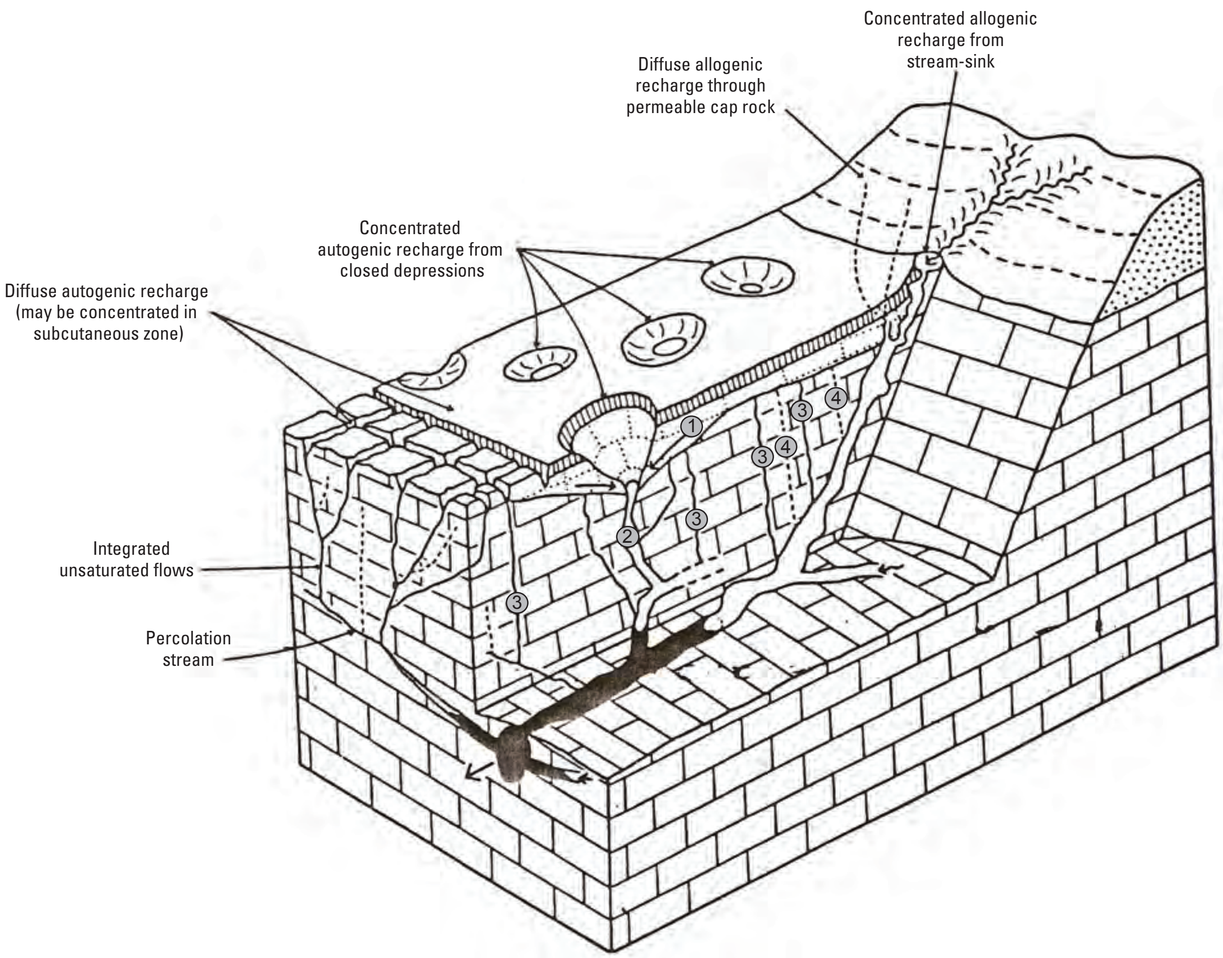



**Figure 4.** Geologic cross section of a karst basin showing various types of recharge sources: concentrated versus diffuse, and autogenic (recharge that originates as precipitation falling directly on karstic rocks) versus allogenic (recharge that originates as precipitation falling on nonkarstic rocks). Water flows through the unsaturated zone via (1) diffuse flow through soil or unconsolidated surface materials, (2) concentrated flow through solution-enlarged sinkhole drains, (3) diffuse infiltration through vertical fractures, and (4) diffuse infiltration through permeable rock matrix. Subterranean conduits shown as solid black are filled with ground water. (Modified from Gunn, 1986, used with permission.)

In typical studies of karst hydrology, the understandable focus placed on the characterization of concentrated recharge tends to overshadow the fact that most recharge to karst aquifers is contributed by diffuse autogenic recharge—that is, by infiltration through soil—as it is in most other hydrogeologic settings. In a study in Missouri, Aley (1977) estimated that the quantity of water contributed to a karst aquifer from diffuse areal recharge was approximately four times greater than that contributed by all concentrated recharge sources and almost twice that contributed by sinkholes and losing streams combined. Most sinkhole swallets have active inflow only during periods of heavy surface runoff when soil and macropore infiltration capacity is exceeded and, depending on antecedent moisture conditions, the inflow of concentrated recharge by way of swallets may not occur during many storms.

A particularly important source of recharge and storage in most karst aquifers is the *epikarst*—a zone of intensely weathered, fractured, and solution-modified bedrock located near the soil-bedrock contact (Williams, 1983). The thickness and physical hydrogeologic properties of the epikarst are highly variable within and among karst terranes because epikarst development is dependent on stratigraphic variability; bedrock porosity, permeability, and solubility; fracture density; and intensity of weathering. In terms of hydrology, the epikarst functions generally as a leaky perched aquifer zone, providing relatively long-term, diffuse autogenic recharge to conduits (Klimchouk, 2004). Much of the base-flow discharge from karst aquifers to springs and surface streams is water contributed from storage in the epikarst. Chemical hydrograph separation studies have indicated that flushing of water from the epikarst may contribute as much as 50 percent of the water discharging from springs during storms (Trćek and Krothe, 2002). Much research has been devoted to the development and hydrologic functioning of the epikarst; however, it remains one of the more poorly understood recharge components of karst aquifers (Aley, 1997; Jones and others, 2004).

## Karst Drainage Basins

In typical hydrogeologic studies, a fundamental mapping unit—usually defined by the ground-water basin—is used to characterize the spatial and temporal properties of the aquifer and to construct a conceptual model. For a karst aquifer, the traditional concept of the term "ground-water basin" is somewhat of a misnomer in that it minimizes the highly interconnected nature of surface and subsurface waters and the role of concentrated stormwater runoff as a significant source of recharge. A more appropriate term, and conceptual model, for most karst aquifers is the karst drainage basin (or karst basin)—a mapping unit defined by the total area of surface and subsurface drainage that contributes water to a conduit network and its outlet spring or springs (Quinlan and Ewers, 1989; Ray, 2001). Karst basins differ from conventionally defined ground-water basins—that is, the local ground-water basins described by Toth (1963)—in the following respects:

- Karst basin boundaries do not always coincide with topographic drainage divides, and discharge may or may not always be to the nearest surface stream.
- Recharge near the basin boundaries may flow in a radial or semiradial direction into adjacent basins drained by other underflow springs, and the divides between basins may be indistinct and may shift with changing hydrologic conditions.
- Direct injection of concentrated stormwater runoff and subsurface piracy of surface streamflows constitute a significant portion of the recharge to the basin.
- Most of the active flow is concentrated in the core of the basin, which consists of the conduit network, and is characterized by pipe-full or open-channel hydraulics. Vertical or cascading flow may be significant (Thrailkill, 1985).
- Hydraulic gradients, the number of active conduit flow routes, and directions of ground-water flow may change rapidly with changing hydrologic conditions.
- Directions of ground-water flow do not always conform with the maximum hydraulic gradient inferred by water-level measurements in wells.
- The contributing area and volume of discharged subsurface water changes over time as conduit development, hydraulic capacity, and subsurface piracy increases. In addition, the aquifer carries a substantial sediment load that is constantly changing and can alter flow routes and hydraulic properties of conduits (White, 1988; Dogwiler and Wicks, 2004).

Ray (1999, 2001) proposed that karst basins can be broadly categorized into three functional hydrologic groups on the basis of the hydraulic capacity of their conduit networks and their dominant recharge source (allogenic or autogenic). The three basin groups are defined as:

- Overflow allogenic basins—basins in which the trunk (master) conduit draining the basin is recharged mostly by subsurface piracy of a surface stream(s), but because of limited hydraulic capacity, the surface channel is maintained as a losing stream reach or as an intermittent, storm-overflow route.
- Underflow allogenic basins—basins in which the hydraulic capacity of the trunk conduits has increased to the point that the surface flow is completely diverted underground through streambed swallets, and the surface valley becomes blind.
- Local autogenic basins—basins in which all surface flow has been captured by subsurface piracy, and the trunk conduit is recharged almost exclusively by infiltration through the soil and internal sinkhole drainage.

These basin categorizations apply best to shallow, unconfined karst aquifers in fluviokarst settings, but also describe basins in doline karst and in deeper partly confined karst aquifers such as the Madison Limestone aquifer (Greene, 1997) in South Dakota, which is characterized by overflow allogenic basins recharged by sinking streams draining a structurally uplifted recharge area. A progressive sequence of karst basin development—from overflow allogenic to underflow allogenic to local autogenic—may occur in many karst terranes over geologic time as karstification and subsurface piracy of surface streams increases (Smart, 1988; Ray, 2001).

# Hydrogeologic Characterization

As in other complex hydrogeologic settings, a proper hydrogeologic characterization of karst drainage basins is the key to understanding and estimating water fluxes. As applied within the framework of a karst conceptual model, this requires the acquisition of data needed to characterize the extent and overall effects of conduit-dominated flow, multiple discrete inputs and outputs for water, and spatial and temporal variability in recharge, storage, and flow. Water-tracing tests, typically done using fluorescent dyes, are the most effective means of determining subsurface conduit connections between karst drainage features such as sinkholes and springs, directions of ground-water flow in the karst aquifer, boundaries of karst ground-water basins, and the hydraulic properties of conduits (Mull and others, 1988; White, 1993). The analysis of spring discharge hydrographs and temporal variations in the chemical or isotopic composition of spring water provide data needed to characterize the recharge, and storage and discharge functions occurring in karst aquifers and to provide additional insights into the structure of conduits at basin-to-regional field scales (Ford and Williams, 1989; White, 1993).

Other, more conventional hydrogeologic data-collection methods—including those described in other chapters of this report—also may be used in the study of karst aquifers if these methods are applied within the framework of a karst conceptual model. Careful consideration must be given to the field scale of collected hydrologic measurements and to whether the measurements obtained by use of a particular method are representative of the conduit-dominated flow components of the aquifer, the aquifer matrix or nonconduit flow component, or a composite of both.

In addition to the topographic, structural, and stratigraphic characteristics that are necessary to define the physical hydrogeologic framework, White (1999) proposes six basic hydrologic properties needed for the evaluation of karst basins: (1) the area of the karst basin, (2) allogenic recharge, (3) conduit carrying capacity, (4) matrix and fracture system hydraulic conductivity, (5) conduit system response, and (6) conduit/fracture coupling. A water budget is suggested here as a seventh additional characteristic for evaluation. Information collected about each of these seven karst basin features will contribute to the identification and estimation of fluxes between surface water and ground water in karst terranes.

## Area of the Karst Drainage Basin

Various methods have been used to estimate the recharge or contributing areas of karst springs (Ginsberg and Palmer, 2002), but dye-tracer tests provide the most effective means of identifying the point-to-point connections between flow inputs (sinkholes or sinking streams) and outputs (springs) needed to actually define the boundaries of karst drainage basins (White, 1993; Ray, 2001). Dye-tracer tests can be done at multiple input sites by injecting different fluorescent dyes either simultaneously or sequentially. As tracer-inferred ground-water flow directions are determined and the number and distribution of tracer-determined flow paths increase, the boundaries, approximate size, and shape of the basin under study can be delineated with increasing levels of confidence. To fully delineate the boundaries of the area contributing recharge to a particular spring, dye-tracer tests need to be planned and conducted in strategic locations so that the results obtained "push" the point-to-point connections established between the spring and its contributing inputs (for example, sinkholes) toward the anticipated locations of subsurface drainage divides. The presence of these drainage divides are inferred where the trajectories of plotted dye-tracer flow paths indicate a divergence in subsurface flow directions, that is, identify areas where subsurface flows are being routed to springs draining other adjacent karst basins. The geographic distribution of these inferred subsurface drainage divides constrains the boundaries of the karst basin under study (fig. 5).

Tracer-inferred flow paths can be plotted as straight lines between input and resurgence sites, or preferably, as curvilinear vectors that depict a tributary drainage system more visually representative of the natural conduit network (Ray, 2001). Other hydrogeologic mapping data such as cave surveys or contoured water-level maps can be used as an aid in the planning and interpretation of dye-tracer tests; for example, the locations of major ground-water conduits often are correlated with the positions of apparent troughs in the potentiometric surface or water table, which are thought to represent a locus of maximum ground-water flow (Quinlan and Ewers, 1989). Karst mapping studies that illustrate various applications of these techniques include those of Crawford (1987), Mull and others (1987), Vandike (1992), Bayless and others (1994), Schindel and others (1995), Imes and others (1996), Jones (1997), Taylor and McCombs (1998), and Currens and Ray (1999).

Dye-tracer tests have routinely shown that conduit flow paths commonly extend beneath topographic drainage divides and, in some places, beneath perennial streams, and that surface runoff draining into sinkholes or sinking streams in one topographic basin (watershed) may be transferred via subsurface flow routes into adjacent topographic basins (Ray, 2001) (fig. 6). In karst terranes, mapping of the contributing areas of springs and surface streams, identification and estimation of water fluxes and, in particular, estimation of water budgets for either surface or subsurface drainage basins, are critically dependent on identifying and delineating the areas that indicate this "misbehaved drainage" (White and Schmidt, 1966; Ray, 2001). Dye-tracer tests are the most reliable method of obtaining this information. For example, dye-tracer tests were used to conclusively demonstrate that the USGS Hydrologic Unit (watershed) boundaries delineated for the Barren River basin in central Kentucky using topographic drainage divides encompass approximately 220 square kilometers (85 square miles) of surface drainage that actually contributes water to the adjacent Green River basin via subsurface conduits (Ray, 2001) (fig. 6).

## Allogenic Recharge and Conduit Carrying Capacity

As previously noted, a significant component of recharge to underflow and overflow allogenic karst basins is the water contributed by subsurface piracy of surface streams, and it is this concentrated allogenic recharge that largely influences the discharge and water-chemistry changes indicated by karst springs during and after storms. Quantifying the allogenic recharge subbasin area and the sum of the inputs from individual sinking or losing streams defines an important characteristic of the hydrology of a karst basin. Geographic Information System (GIS) technology provides a convenient way of delineating the catchment areas of all sinking or losing streams that contribute to a karst basin and of estimating the relative proportion of allogenic recharge subbasin area to autogenic recharge (sinkhole-dominated) subbasin area (Taylor and others, 2005). In theory, all of the allogenic recharge contributed to a karst basin can be measured by synoptic gaging of discharge in the stream channels directly above the locations of terminal swallow holes. When evaluated with discharge measurements from the basin's outlet springs, the measured allogenic inputs provided by each sinking or losing stream can be used to evaluate conduit-carrying capacity (White, 1999) in the following manner:

In underflow allogenic basins, the hydraulic capacities of the conduits are defined by the following relation:

$$Q_c > Q_{a(max)} \tag{1}$$

where

$Q_c$ is the carrying capacity of the conduits

and

$Q_{a\,(max)}$ is the maximum discharge of the surface stream(s) contributing recharge to the conduits.

In this particular instance, the carrying capacity of the conduit network always exceeds the maximum input contributed by the allogenic stream recharge, and surface flows are completely diverted underground by one or more swallow holes shortly after crossing onto karstic bedrock. This case describes a classic sinking stream.

In overflow allogenic basins, the carrying capacities of the conduits are defined by one of two relations (eqs. 2 or 3):

$$Q_{a(base)} > Q_c, \tag{2}$$

where

$Q_{a\,(base)}$ is the base-flow discharge of the allogenic surface stream.

In this case, the carrying capacity of the conduits cannot accommodate the base-flow discharge of the allogenic stream, and perennial surface flow occurs in the channel despite flow losses through streambed swallow holes. This case describes a classic losing stream.

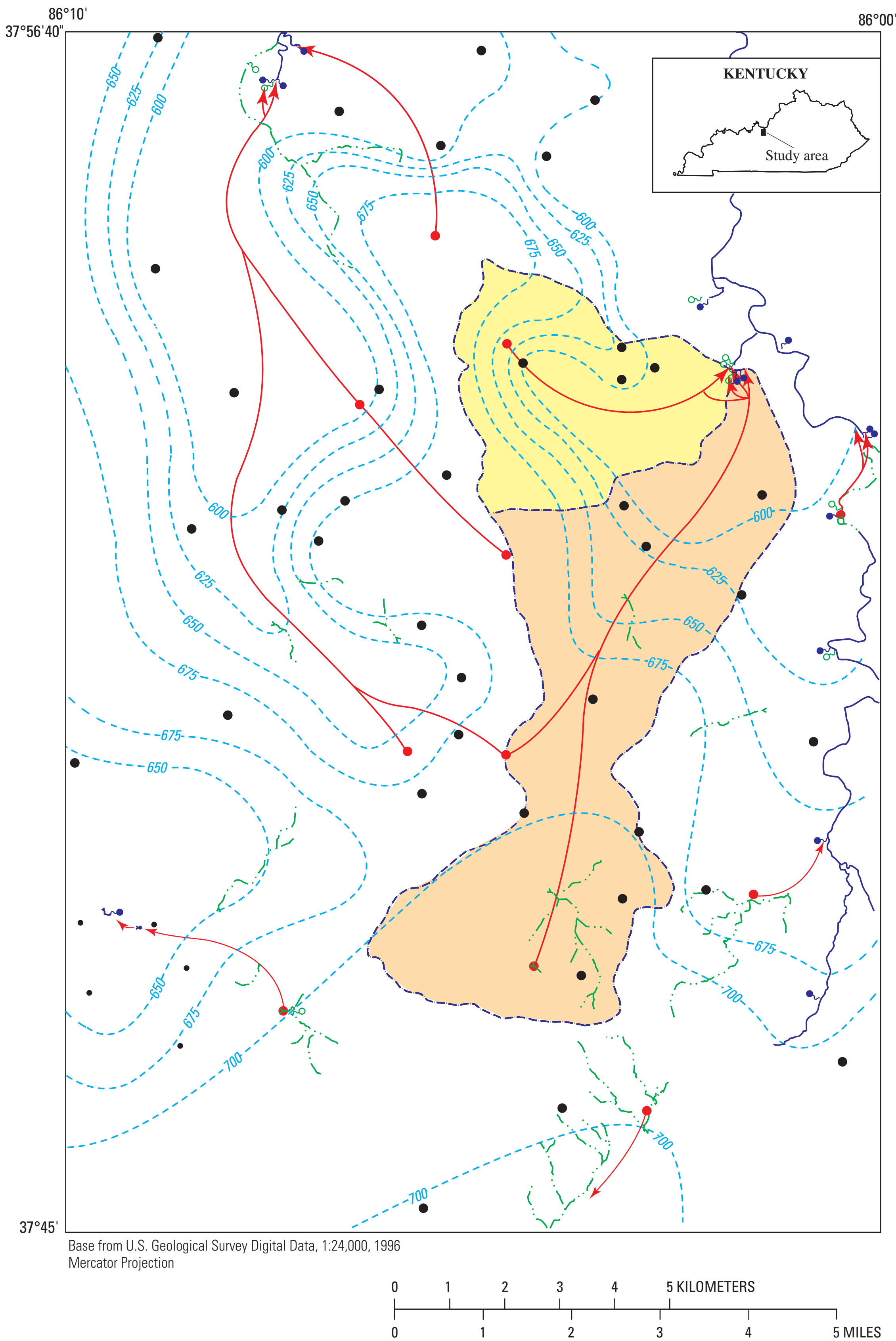
$$Q_{a(max)} > Q_c > Q_{a(base)}. \tag{3}$$

In this case, the carrying capacity of the conduits can accommodate all of the base-flow discharge from the allogenic stream, but stormflow discharge often exceeds the capacity of the conduits, overtops swallow holes, and results in continuation of flow down the channel. This case describes an intermittent sinking stream, often characterized as a "dry-bed stream" (Brahana and Hollyday, 1988). The reactivation of swallow holes as sink points often occurs in a successive manner as surface flow overtops upstream swallow holes first and reaches or overtops the farthest downstream swallow holes only during the largest storms (George, 1989).

White (1999) makes the interesting suggestion that determining the critical flow threshold when $Q_a = Q_c$ would be a meaningful way of characterizing conduit permeability; however, it would require gaging the discharge in sinking streams above the terminal swallow holes at the exact time that the swallow holes are filled and overtopped. There are practical difficulties involved in obtaining such measurements, not only with regard to the timing of the measurements, but because flow in the channels of many sinking streams often is lost progressively through a series of swallow holes; for example, the Lost River basin of southern Indiana (Bayless and others, 1994), or because clogging of the swallow holes with sediment or debris is a factor that controls the rate of inflow (Currens and Graham, 1993).

## Matrix and Fracture System Hydraulic Conductivity

Because of combined permeability provided by matrix, fracture, and conduit-flow components, the timing and amount of response to hydraulic stresses varies greatly from place to place within a karst aquifer. Investigation of the hydraulic conductivity of the matrix and fracture components is typically performed with conventional hydrogeologic tools. Matrix permeability can be determined using laboratory permeability tests done on representative rock core samples. Fracture hydraulic conductivity is best determined using


86°10'
37°56'40"
86°00'
37°45'
KENTUCKY
Study area
600
625
650
675
700
Base from U.S. Geological Survey Digital Data, 1:24,000, 1996
Mercator Projection
0 1 2 3 4 5 KILOMETERS
0 1 2 3 4 5 MILES

**EXPLANATION**

**Northern ground-water basin**

**Southern ground-water basin**

600 **Inferred potentiometric-surface contour**—Shows altitude at which water level is expected to stand in tightly cased wells completed exclusively in the St. Louis Limestone. Contour interval 25 feet. Datum is North American Vertical Datum of 1988 (NAVD 88).

**Ground-water basin boundary**—Appropriate location of ground-water divide defined by topographic, geologic, and hydrologic features that influence the direction of ground-water flow.

**Dye flow path**—Shows inferred route of dye tracer in karst aquifer and confirmed hydraulic connection between dye-injection site and dye-recovery site. Dashed line indicates intermittent flow route to an overflow spring. Number indicates dye-tracing test.

**Intermittent stream and terminal sink point (swallow hole)**

**Well**

**Dye-injection site**

**Dye-recovery site**

**Perennial (underflow) spring**

**Intermittent (overflow) spring**

**Karst-window**—Perennial spring and sinking stream

**Figure 5 (above and facing page).** Part of map showing dye-tracing flow paths (red curvilinear vectors) used to constrain the boundaries for two karst spring subbasins (orange, yellow shading). Dashed blue lines are water-table contour lines, which provide additional information useful in mapping the basin boundaries and interpreting subsurface flow paths (modified from Taylor and McCombs, 1998).

straddle-packer hydraulic tests and borehole flow meters (Sauter, 1991). Conventional aquifer tests (time-drawdown, distance-drawdown, or slug tests) provide a measurement of the integrated local matrix and fracture system transmissivity. Borehole geophysical methods, including cross-borehole tests, also provide valuable data to assist with permeability and flow characterization at local to subbasin scales (Paillet, 2001).

Analysis of karst aquifer test data using conventional Darcian analytical methods may provide erroneous results, and special consideration should be given to the possible effects of slow-flow and quick-flow karst components on the hydraulic responses represented by the well-hydraulic test data. Streltsova (1988) reviews aquifer-test methods best suited to investigations of heterogeneous aquifers such as karst. If the test well penetrates large solutional openings or conduits, the hydraulic conductivity (or transmissivity) and storage coefficients of these should be evaluated separately from those of fractures (Greene and others, 1999).

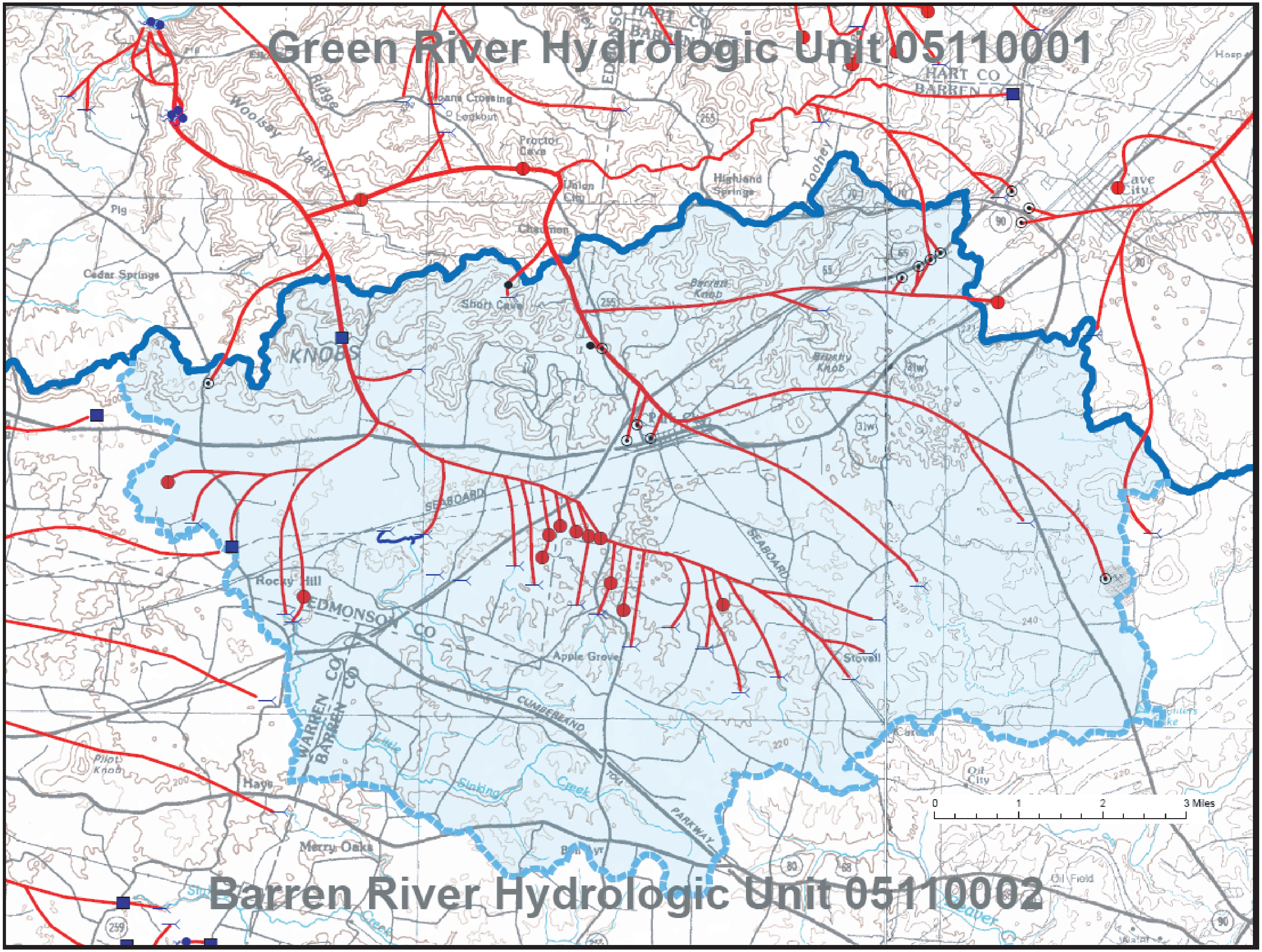
Comparative studies of hydraulic properties measured in different karst aquifers have shown that, regardless of the range of porosity measured in the aquifer matrix, conduits typically account for less than 1 percent of the porosity of the aquifer, but more than 95 percent of the permeability (table 2) (Worthington and others, 2000). As in studies of many fractured rock aquifers, there is a general tendency for measured hydraulic conductivities to increase with increasing field scale (Sauter, 1991). Typically, the distribution of hydraulic conductivity and other properties is related to lithostratigraphic facies changes or other physical changes in the characteristics of the bedrock matrix (Rovey and Cherkauer, 1994).

## Conduit System Response

Conduit system response may be evaluated using: (1) quantitative water-tracing tests to determine traveltime and tracer-breakthrough characteristics, (2) recession analysis of spring discharge hydrographs (White, 1999), (3) evaluation of the ratio between peak storm discharge and base-flow discharge ($Q_{max}/Q_{base}$) of karst springs, (4) chemical hydrograph separation, and (5) hydrologic pulse analysis—analysis of changes in spring discharge and water-quality constituents in response to storms (Ryan and Meiman, 1996; Katz and others, 1997).

Early studies of the variation in spring discharge and water chemistry led to the suggestion that karst springs and aquifers could be categorized along a hydrologic continuum defined by conduit-dominated and diffuse-dominated end members (Shuster and White, 1971; Atkinson, 1977; Scanlon and Thrailkill, 1987) with the observed hydrologic response differing according to the proportion of conduit-to-nonconduit permeability (fig. 7). So-called conduit-dominated karst springs typically exhibit rapid changes in discharge and wide-ranging changes in water chemistry in response to precipitation input (fig. 8). In contrast, so-called diffuse-dominated karst springs respond more slowly to precipitation input and exhibit more buffered, gradual changes in discharge and water chemistry. These distinctions seem to be applicable in a broadly descriptive context and are still used as a convenient way of characterizing karst flow systems.

More recent studies have indicated that karst spring discharge and water chemistry responses are influenced by temporal variability in the proportion of recharge contributed from diffuse and concentrated sources (White, 1999), and by the timing and volume of water contributed from conduit, fracture, and matrix flow components that reflect the range of transmissivities present in the karst basin or aquifer (Doctor and Alexander, 2005). Many karst springs and aquifers are observed to exhibit a dual or triple hydrologic response to precipitation defined by: (1) an initial rapid flow response created by water transmission in conduits greater than 5 to 10 millimeters in diameter where velocities generally exceed 0.001 meter per second, followed by (2) a secondary, slower flow response created by water transmission in intergranular pore spaces, smaller aperture fractures, and solutional openings within the aquifer matrix where velocities are less than 0.001 meter per second (Worthington, Davies, and Ford, 2000), and (3) a transitional response period between these


Green River Hydrologic Unit 05110001
Barren River Hydrologic Unit 05110002
Woolsey Valley
Proctor Cave
Highland Springs
Cave City
Cedar Springs
Short Cave
Barrett Knob
Brushy Knob
KNOBS
Rocky Hill
EDMONSON CO
WARREN CO
BARREN CO
Apple Grove
Stovall
SEABOARD
CUMBERLAND
PARKWAY
Little Sinking Creek
Hays
Merry Oaks
Pilot Knob
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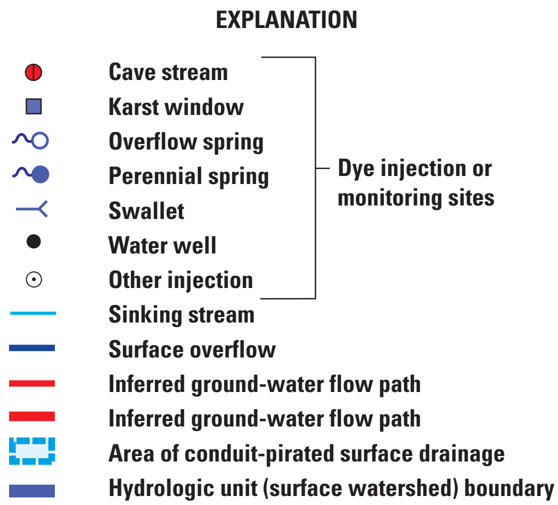


**Figure 6 (above and facing page).** Subsurface conduit piracy of surface drainage from part of the Barren River watershed to springs discharging to the Green River watershed. (Courtesy of Joe Ray, Kentucky Division of Water, map modified from Ray and Currens, 1998.)

two. Accordingly, the alternative terms "quick flow," "slow flow," and "mixed or intermediate flow" now are used often to describe the range of hydrologic responses exhibited by a karst spring or aquifer (White, 1993). Various methods of spring hydrograph analysis, summarized later in this chapter, may be applied to investigate and quantify these changes in karst hydrologic responses.

One simple method of quantifying and evaluating the "flashiness" of the conduit system response is to determine the ratio of maximum peak-flow to base-flow spring discharge ($Q_{max}/Q_{base}$); it is a function of storm intensity and conduit organization or interconnectivity (White, 1993). Springs dominated by a quick-flow response typically exhibit $Q_{max}/Q_{base}$ ratios in the range of 40 to 100, whereas ratios of about 1 to 3 and 7 to 10, respectively, are exhibited by springs dominated by a slow flow response and by intermediate or mixed flow response (White, 1993). The timing of these changes in hydrologic response depends on the size of a karst basin, the distances between flow inputs and outputs, and on the internal organization of its conduit network (White, 1993). The response time, $t_r$, determined by fitting an exponential function to the recession limb of the spring hydrograph, also seems to indicate a wide range in values that cluster into distinctive groups characteristic of each hydrologic response type.

## Conduit/Fracture Coupling

Under normal base-flow conditions, conduits act as low-hydraulic resistance drains that locally alter the hydraulic flow (or equipotential) field so as to capture ground water from the surrounding aquifer matrix and adjoining fractures (White, 1999). The flux of water between conduit-flow and nonconduit-flow components is a complex head-dependent process and may be reversible when conduits fill completely and pressurize under certain storm-flow conditions. Water flux reversal also can be induced by backflooding of surface streams, wherein surface water enters conduit passages by way of underflow and overflow springs and results in hydraulic damming. In either instance, the injection of water from the conduits back into the aquifer matrix constitutes an unusual type of aquifer recharge and bank storage, which has been well documented, for example, in the Green River-Mammoth Cave karst aquifer system in Kentucky (Quinlan and Ewers, 1989). As stormwater or flood pulses are drained rapidly through the conduits, spring discharge returns to base-flow conditions, and the normal flux resumes as the dominant source of recharge shifts to water contributed from longer term storage in the epikarst, bedrock matrix, fractures, and smaller tributary conduits.

The effectiveness of the coupling between conduit and fracture components, combined with the hydraulic conductivity of the matrix/fracture system, control the rate of movement of water into and out of storage after storms or floods and during base-flow conditions (White, 1999). The conduit/fracture coupling can be evaluated by: (1) deconvolution of spring hydrographs, (2) comparisons of storm-related hydrograph response in springs or observation wells in the manner described by Shevenell (1996), and (3) evaluation of unit base flow.

The unit base flow (UBF), or base-flow discharge per unit area, is a particularly useful measurement derived from the concept that surface-stream watersheds of similar size (area) located in similar hydrogeologic settings and climates will generate approximately equal quantities of base-flow runoff (Quinlan and Ray, 1995). Applied to karst basins, the UBF represents the amount of water discharged from long-term ground-water storage, as controlled by the coupling between the conduits and the diffuse-flow component. Its value is best calculated by using dry-season, base-flow spring discharge measurements (Quinlan and Ray, 1995). Table 3 lists the range of UBF values calculated for several spring basins in Kentucky. UBF values are useful in estimating the basin areas of springs in similar hydrogeologic settings whose basin boundaries are unknown or untraced, and help identify anomalous recharge or storage characteristics for a spring basin under study (White, 1993; Quinlan and Ray, 1995).

**Table 2.** Comparison of porosity and permeability measurements in various karst aquifers (after Worthington, 1999).

[%, percent; m/s, meter per second]

| Karst area | Porosity (%) | | | Hydraulic conductivity (m/s) | |
|---|---|---|---|---|---|
| | Matrix | Fracture | Conduit | Matrix | Fracture |
| Smithville, Ontario | 6.6 | 0.02 | 0.003 | $1\times10^{-10}$ | $1\times10^{-5}$ |
| Mammoth Cave, Kentucky | 2.4 | 0.03 | 0.06 | $2\times10^{-11}$ | $1\times10^{-5}$ |
| Devonian Chalk, England | 30 | 0.01 | 0.02 | $1\times10^{-8}$ | $4\times10^{-6}$ |
| Nohoch Nah Chich, Yucatan, Mexico | 17 | 0.1 | 0.5 | $7\times10^{-5}$ | $1\times10^{-3}$ |

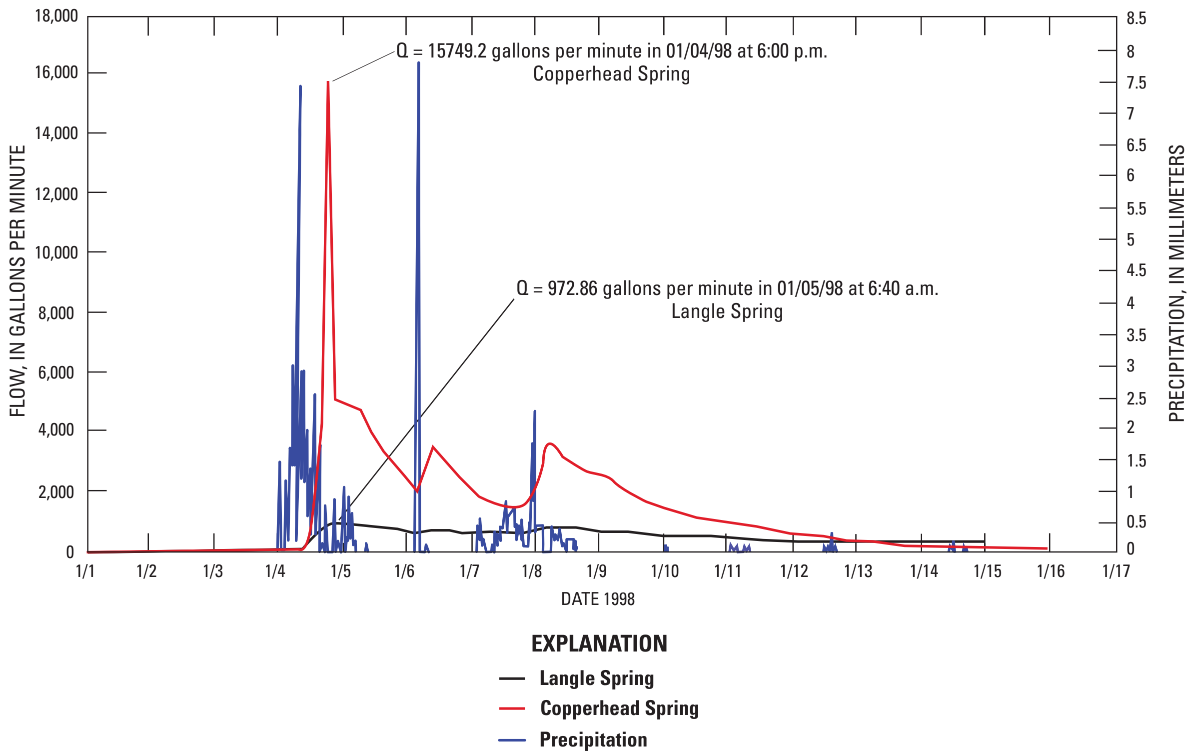


**Figure 7.** Variable response of springs to precipitation. Copperhead Spring hydrograph shows rapid conduit-dominated flow response. Langle Spring hydrograph shows slow diffuse-dominated flow response. These are related to the relative proportion of conduit permeability to nonconduit permeability (courtesy of Van Brahana, University of Arkansas).

## Water Budget

Water budgets typically are written with the instantaneous flows integrated over a specified period of time, which can be a water year (season to season), a season, the duration of a single storm, or any other period (White, 1988). Published examples of water-budget calculations for karst aquifers include Bassett's (1976) study of the Orangeville Rise spring basin in south-central Indiana, and Hess and White's (1989) study of the spring-fed Green River within the boundaries of Mammoth Cave National Park in Kentucky. Other examples are discussed by Milanovic´ (1981a, b) and Padilla and others (1994). Equations may take a variety of forms depending on the purpose and the hydrologic terms that can be estimated or must be evaluated. A simple, conventional water-budget equation for a karst basin or aquifer may be written in the form:

$$I = O + ET + \Delta S, \qquad (4)$$

where

$I$ is precipitation,

$O$ is basin or spring discharge,

$ET$ is evapotranspiration,

and

$\Delta S$ is change in ground-water storage (Bassett, 1976).

Using this equation, a water balance can be obtained by summing the values for $O$, $ET$, and $\Delta S$ and subtracting the resulting value from $I$. The results are expressed as the percentage of rainfall unaccounted for (positive $I$ values), or in excess of the balance (negative $I$ values) (table 4).

A water-budget equation also can be written to express the change in storage occurring as a result of a storm:

$$Q_i - Q_o = \pm \Delta V / \Delta t, \qquad (5)$$

where

$Q_i$ is the total inflow or recharge contributed by the storm,

$Q_o$ is the outflow discharge,

$\Delta V$ is the change in storage,

and

$\Delta t$ is the time period of the storm (Ford and Williams, 1989).

Antecedent precipitation and soil-moisture conditions are influential in determining the magnitude of $Q_i$ and $Q_o$.

More complex water-budget equations can be developed to include additional karst hydrologic factors. White (1988), for example, describes how a water budget developed for an allogenic overflow karst basin might include terms for the input by sinking streams (the allogenic recharge), internal runoff (sinkhole drainage), diffuse infiltration (through soil, epikarst, and bedrock matrix), and positive or negative changes in ground-water storage. In these types of calculations, allogenic

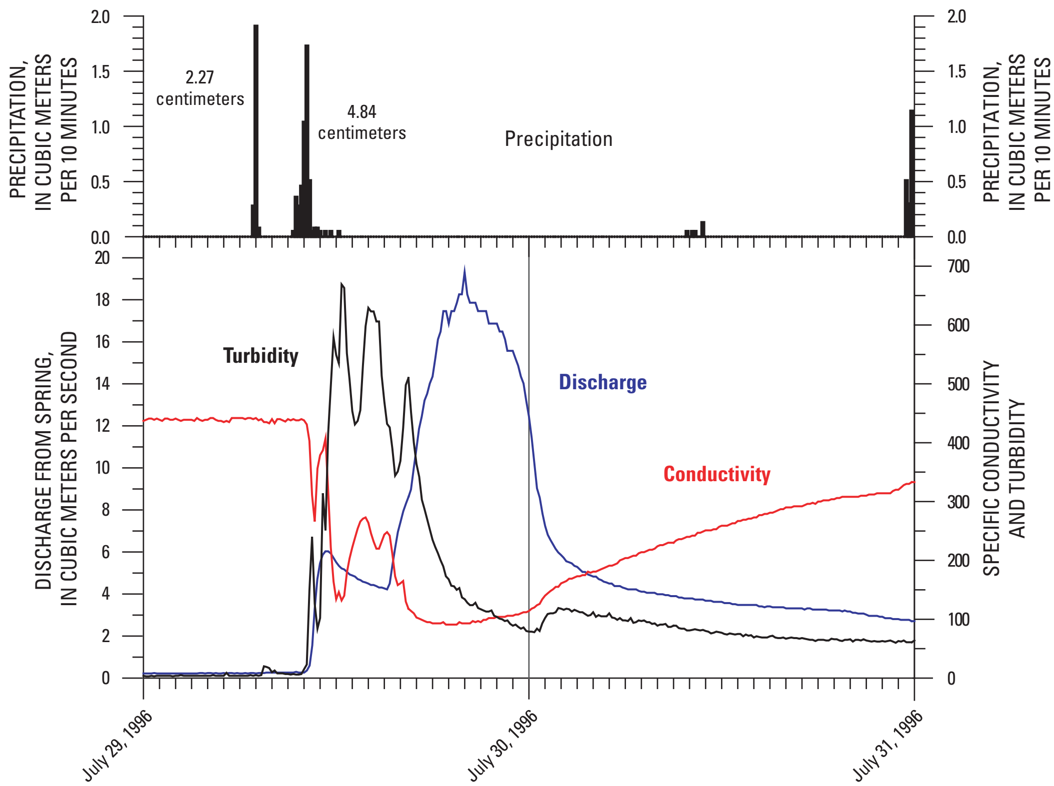


**Figure 8.** Water-quality changes in a karst spring related to allogenic recharge: Precipitation-stormwater runoff causes a rise in turbidity and a decrease in specific conductance prior to and during peak spring discharge. After passage of the stormwater recharge pulse, conductivity increases as the spring discharge returns to base flow (discharge of water contributed from storage by the slow diffuse-flow karst component). (Courtesy of James Currens, Kentucky Geological Survey.)

recharge from sinking or losing streams can be directly measured, at least in theory, and estimation of the contribution of autogenic (sinkhole) recharge is more problematic. Change in storage typically is estimated from analysis of spring recession hydrographs using the methods described by Kresic (1997), or estimated in terms of net head change in the aquifer on the basis of water-level measurements from observation wells.

## Spring-Discharge Hydrograph Analysis

Spring discharge represents an integration of the various processes that govern recharge, storage, and throughflow in a karst basin upstream from its outlet (Kresic, 1997). Analysis of the spring-discharge hydrograph makes it possible to obtain valuable insights into hydraulic stresses acting on the basin, to evaluate basin flow characteristics, and to estimate average basin hydraulic properties (Bonacci, 1993; Baedke and Krothe, 2001; Pinault and others, 2001). A wide variety of graphical, time-series, and spectral analysis techniques have been applied that are beyond the scope of discussion of this chapter. Many of these techniques are reviewed by White (1988) and Ford and Williams (1989).

Analysis of the recession period of the spring discharge hydrograph is one of the simpler and more useful methods to apply to karst studies because it provides information about the volume of water drained from the karst basin over time after peak flows and changes in the rate of discharge that may indicate thresholds and limits in aquifer flow regimes (Doctor and Alexander, 2005). A step-by-step review of the recession analysis technique is presented by Kresic (1997). Its application to the determination of karst basin flow and hydraulic characteristics is summarized here.

### Basin Flow Characteristics

Interpretation and analysis of a spring hydrograph assumes that: (1) the discharge of the spring is controlled by input events such as a high-intensity precipitation event or a recharge event at a sinking stream, and (2) the shape of the hydrograph is controlled by flow through various pathways that have different conductivities and velocities (Milanović, 1981b). By using recession analysis, it is possible to identify whether the overall basin flow characteristics are dominated by quick flow (conduit-dominated flow), slow flow

**Table 3.** Range of Unit Base Flow (UBF) values (designated here as normalized base flow or NBF) in spring basins in Kentucky (modified from Quinlan and Ray, 1995).

[cfs, cubic feet per second]

| Autogenic recharge group | Spring name | Spring identification number | Basin area (miles$^2$) | Summer base flow (cfs) | Normalized base flow, NBF (cfs/mile$^2$) | Geometric mean NBF (cfs/mile$^2$) |
|---|---|---|---|---|---|---|
| 1. With up to 25 percent allogenic recharge from sandstone-capped ridgetops or near-surface, leaky, chert aquitard | Lavier Blue Hole | 21 | 10.2 | 2.1 | 0.21 | 0.20 (0.21) |
| | Garvin-Beaver | 22 | 7.2 | 1.7 | 0.24 | |
| | Echo River | -- | 8.8 | 1.8 | 0.21 | |
| | Lost River | -- | 55.2 | 12.0 | 0.22 | |
| | Pleasant Grove | -- | 16.1 | 2.5 | 0.16 | |
| | Shakertown | -- | 19.0 | 3.6 | 0.19 | |
| 2. With significant allogenic recharge from carbonate terrane | Gorin Mill | 23 | 152 | 25.1 | 0.17 | 0.17 |
| | Turnhole | 3 | 90.4 | 14.3 | 0.16 | |
| | Graham | 21 | 122 | 20.8 | 0.17 | |
| 3. With locally thick, areally significant sand and gravel cover | Rio | 13 | 5.2 | 4.7 | 0.91 | 0.70 (0.75) |
| | Rio | 13 | 6.5 | 4.7 | 0.72 | |
| | McCoy Blue | 1 | 36.1 | 12.3 | 0.34 | |
| | Roaring | 6 | 10.8 | 11.9±1 | 1.19 | |
| | Roaring | | | | 1.01 | |
| | Johnson | 7 | 17.5 | 11.0 | 0.63 | |
| | Jones School | 19 | 3.9 | 2.3 | 0.59 | |
| | Jones School | 19 | 2.9 | 2.3 | 0.79 | |
| 4. With much interbedded shale, Bluegrass Region | Royal | -- | 25.0 | 2.8 | 0.11 | 0.11 |
| | Russell Cave | -- | 6.4 | 1.0 | 0.15 | |
| | Garretts | -- | 7.4 | 0.5 | 0.07 | |

(diffuse-dominated flow), or mixed flow, and to evaluate the timing and magnitude of changes in spring discharge that correspond to changes between these flow regimes (fig. 9).

Analysis of a spring discharge hydrograph to determine the flow regimes of the karst basin is done through methods presented by Rorabaugh (1964) and Milanović (1981a,b). Even though these methods are based on Darcian theory, the hydrograph analysis methods have been successfully applied to many studies of karst basins (Baedke and Krothe, 2001; Shevenell, 1996; Padilla and others, 1994, Sauter, 1992; and Milanović, 1981a). The method of characterizing karst flow regimes is based on the equation below, whereby the recession curves of spring hydrographs are analyzed to calculate the value of $\alpha$, the recession slope:

$$Q_t = Q_o\, e^{-\alpha(t-t_o)} \qquad (6)$$

where

$t$ is any time since the beginning of the recession for which discharge is calculated,

$t_o$ is the time at the beginning of the recession, usually set equal to zero,

$Q_t$ is spring discharge at time t,

$Q_o$ is spring discharge at the start of the recession ($t_o$),

and

$\alpha$ defines the slope, or recession constant, that expresses both the storage and transmissivity properties of the aquifer.

By using a semilog plot of discharge and time during a spring's recession curve, one can easily determine a characteristic $\alpha$ value that defines the recession curve slope. For some hydrographs, one $\alpha$ value may be obtained that is sufficient to describe the slope of the recession curve. It is not uncommon, however, for karst springs to exhibit two to three major changes in slope on a single hydrograph recession limb, and here it is advantageous to evaluate each slope change and its corresponding $\alpha$ value. A common interpretation of these changes is that the first and steepest slope represents the transmission of the

**Table 4.** Water budget calculations for Orangeville Rise spring, southern Indiana (modified from Bassett, 1976). Used with permission of the National Speleological Society (*www.caves.org*).

| Interval | Duration (days) | I* | O* | O/I (%) | ET* | ΔS* | Bal.** (%) |
|---|---|---|---|---|---|---|---|
| June 1 August 20, 1972 | 81 | 26.4 | 3.8 | 14 | 26.4 | –2.8 | –3.8 |
| August 21 October 29 | 70 | 20.7 | 2.9 | 14 | 20.7 | +1.4 | –21 |
| October 30 December 5 | 37 | 15.3 | 5.5 | 36 | 3.2 | +2.3 | 28 |
| December 6 January 18 | 44 | 11.4 | 10.8 | 95 | 2.8 | +0.1 | –18 |
| January 19 March 2 | 43 | 6.6 | 7.9 | 120 | 0.8 | –0.4 | –26 |
| March 3 May 6 | 65 | 32.4 | 19.7 | 61 | 3.7 | +1.3 | 24 |
| May 7 July 5, 1973 | 61 | 20.1 | 10.0 | 50 | 7.2 | –1.6 | 6 |

*Units are acre-feet × $10^3$.

**Bal. = I – (0 + ET + ΔS), expressed as a percentage of I.

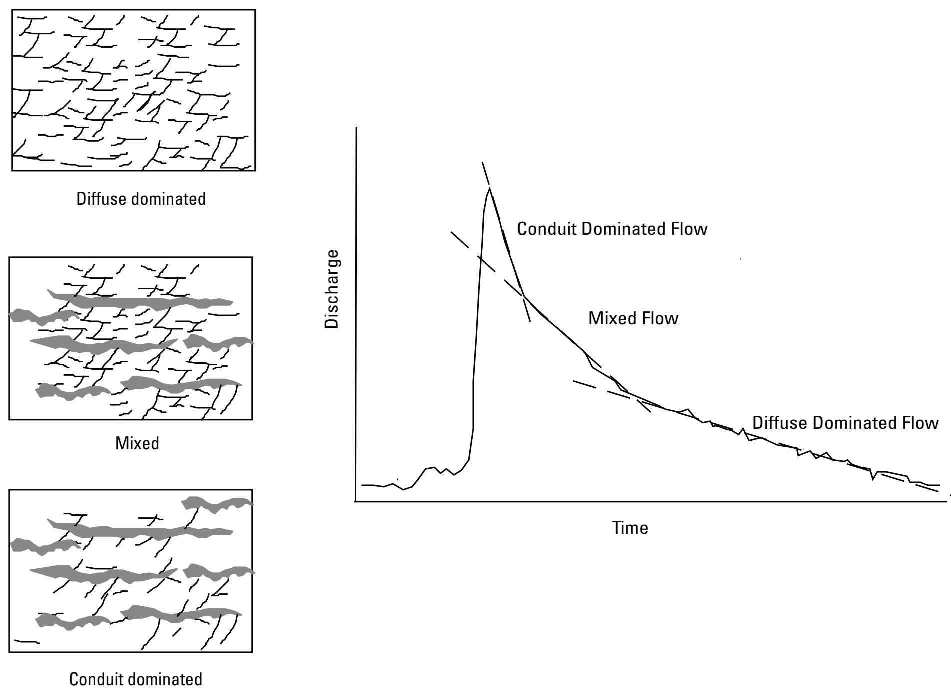


**Figure 9.** Conceptual spring hydrograph showing changes in slope and dominant flow regime (conduit, mixed, diffuse) due to differing hydraulic responses (artwork by Earl Greene, U.S. Geological Survey).

main stormwater runoff pulse through the largest conduits. This often is followed by a change to a less steep, intermediate recession slope interpreted as marking the beginning of depletion of the stormwater pulse and (or) the spring discharge being composed of a mixture of stormwater and stored ground water discharging from smaller conduits and larger fractures. The final change in slope on the recession curve signals the return to base-flow conditions wherein the spring discharge is composed of ground-water stores discharging from a network of smaller fractures and bedrock matrix.

As noted previously, these differences in spring flow characteristics sometimes are referred to as the quick-flow (or conduit-dominated) response, the intermediate flow response, and the slow-flow (or diffuse-dominated) response. The $\alpha$ value calculated for a spring discharge recession curve, or for each "slice" of a multisloped recession curve, typically takes on a characteristic value or range indicative of each type of flow regime (table 5). For example, all three karst flow regimes (quick-flow or conduit-dominated-flow, mixed flow, and slow-flow or diffuse-dominated flow) are evident in the discharge hydrograph for San Marcos Springs from the Edwards aquifer in Texas (fig. 10).

## Basin Hydraulic Properties

Bonacci (1993) and Baedke and Kroethe (2001) have suggested that it is possible to estimate the average transmissivity of the karst basin by using spring-discharge hydrograph analysis, again following the methods of Rorabaugh (1964) and Milanović (1981a), by applying the equation

$$\frac{T}{S_y} = \frac{\log\left[\frac{Q_1}{Q_2}\right]}{(t_1 - t_2)} \frac{L^2}{1.071} \qquad (7)$$

where

$T$ is aquifer transmissivity,

$S_y$ is specific yield, $Q$ is discharge,

$t$ is time,

and

$L$ is the effective karst basin length.

Results obtained from aquifer (well hydraulic) test analysis may be used to estimate the storage ($S_y$) parameter, and Shevenell (1996) and Teutsch (1992) measured the linear distance from the karst spring to the farthest basin drainage divide to obtain a value for $L$. The transmissivity estimate obtained using this method needs to be compared to values determined from aquifer tests and quantitative dye-tracer tests.

## Chemical Hydrograph Separation

Analysis of the flux of dissolved ionic species or isotopes in spring discharge during storms provides a useful means of identifying water fluxes contributed by different sources of recharge and quantifying their proportions in spring discharge. Although a variety of naturally occurring isotope and geochemical tracers may be used (Katz and others, 1997; Katz, 2005), the method requires that there be a distinctive difference in isotope or geochemical composition between water discharged at base flow and that discharged during storm-pulse

**Table 5.** Characteristic values for the slope of the recession curve (α) to determine flow regimes in a karst aquifer.

| α[1] | Prevailing flow regime |
|---|---|
| 0.0018 | Slow-flow or diffuse-flow |
| 0.0058, 0.006 | Mixed (intermediate) flow |
| 0.25, 0.13, 0.038 | Quick-flow or conduit-dominated flow |

[1]Range of characteristic α's from literature (Baedke and Krothe, 2001; Shevenell, 1996; Padilla and others, 1994, Sauter, 1992; and Milanović, 1981a, b).

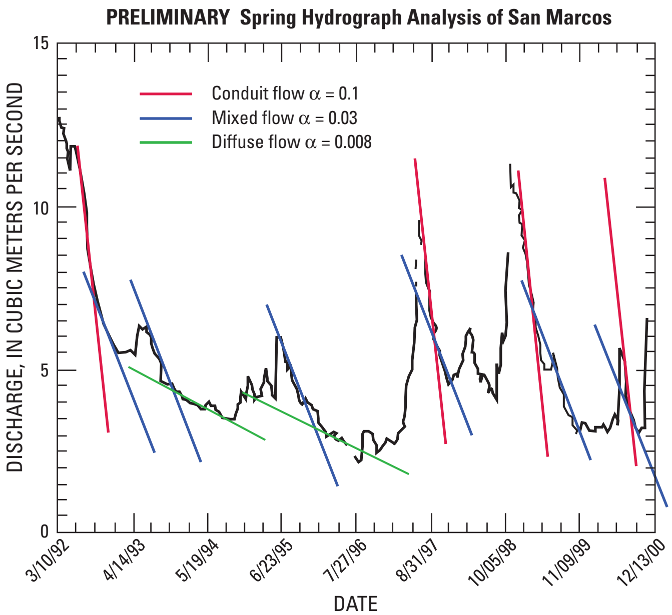


**Figure 10.** Analysis of spring hydrograph of San Marcos Springs in Texas identifying the conduit, mixed, and diffuse flow regimes of the karst aquifer. (Analysis by Earl Greene, U.S. Geological Survey.)

events, and(or) between waters contributed from the various recharge sources under study, in order to determine mixing proportions (Clark and Fritz, 1997). For example, Lakey and Krothe (1996) used stable isotopes of oxygen ($\delta^{18}O$) and hydrogen ($\delta^{2}H$) to calculate the mixing proportions of fresh meteoric water and stored ground water discharging from the Orangeville Rise spring basin in south-central Indiana. Studies done by Lee and Krothe (2001) and Trćek and Krothe (2002) used sulfate, dissolved inorganic carbon (DIC), $\delta^{13}C$ obtained from DIC, $\delta^{2}H$, and $\delta^{18}O$ as natural tracers of recharge contributed by matrix ground water, soil water, epikarst water, and fresh meteoric water, and developed three- and four-component mixing models of spring discharge by hydrograph separation (fig. 11). More recently, Doctor and Alexander (2005) used hydrograph recession analysis to identify the flow regimes contributing to spring discharge and then grouped the sampled chemical and isotopic data according to when they were collected with each flow regime. From analysis of these data, water chemistry patterns were identified that were distinctive of each hydrograph-defined flow regime (flood flow, high flow, moderate flow, and base flow).

# Precipitation Response Analysis

In well-developed karst aquifers, large springs will act as outlets or drains to the system. The rate of ground-water flow and chemical composition of the spring water is directly related to the basin-scale hydraulic and transport properties of the karst aquifer. Because of the direct connection to surface recharge (sinkholes, sinking streams), karst springs have a wide range of physical and chemical response to precipitation events. Depending on the degree of conduit-to-fracture/matrix coupling, spring hydrographs may show a variable response to recharge events. If there is a high degree of conduit-to-fracture/matrix coupling, the spring will respond in a relatively short time (hours to weeks) to a recharge event, whereas, if this coupling is low, the spring response may take many days or weeks. Knowing how the spring response is related to the recharge events is so important in karst hydrology that much research has been directed toward methods of simulating or predicting this response. Three approaches, linear systems analysis, lumped parameter (statistical modeling), and numerical deterministic modeling, commonly are used to simulate or predict the output function (spring discharge) of a karst system on the basis of the known or measured input function (precipitation pulse).

## Linear Systems Analysis

Linear systems analysis has been used in the hydrological sciences for many years to characterize rainfall-runoff relations (Dooge, 1973; Neuman and de Marsily, 1976) and has been used to describe rainfall (recharge)-spring discharge relations in karst systems (Dreiss, 1982; 1983; 1989). The use of a linear method to characterize a nonlinear system (karst ground-water flow) has been justified on a practical basis. First, it is difficult if not impractical to develop a detailed deterministic (numerical) model of ground-water flow in a karst basin because of the difficulty in physically modeling fluid movement in pores, fractures, and conduits. Secondly, the discharge hydrographs of large resurgent springs, like surface-runoff hydrographs, show a response that is directly related to recharge provided by rainfall events. Linear systems modeling will lump many of the complex processes and is useful for describing the karst aquifer.

If a karst system can be conceptualized to act as a linear, time-variant filter, the relation of continuous input (sinkholes, sinking streams, precipitation) can be transformed as continuous output, usually spring discharge (Dreiss, 1982) (fig. 12). The convolution integral below can be used to describe the relation between the output, or spring discharge $y(t)$, and the input, or ground-water recharge $x(\tau)$, and $h(t-\tau)$ is the kernel function (Dreiss, 1982),

$$y(t) = \int_{-\infty}^{\infty} h(t-\tau)x(\tau)d\tau \qquad (8)$$

For two discrete finite series that are causally related, the form of the convolution equation above becomes

$$y_i = \Delta t \sum_{j=0}^{i} x_j h_{i-j} \quad i = 0,1,2,.....N \qquad (9)$$

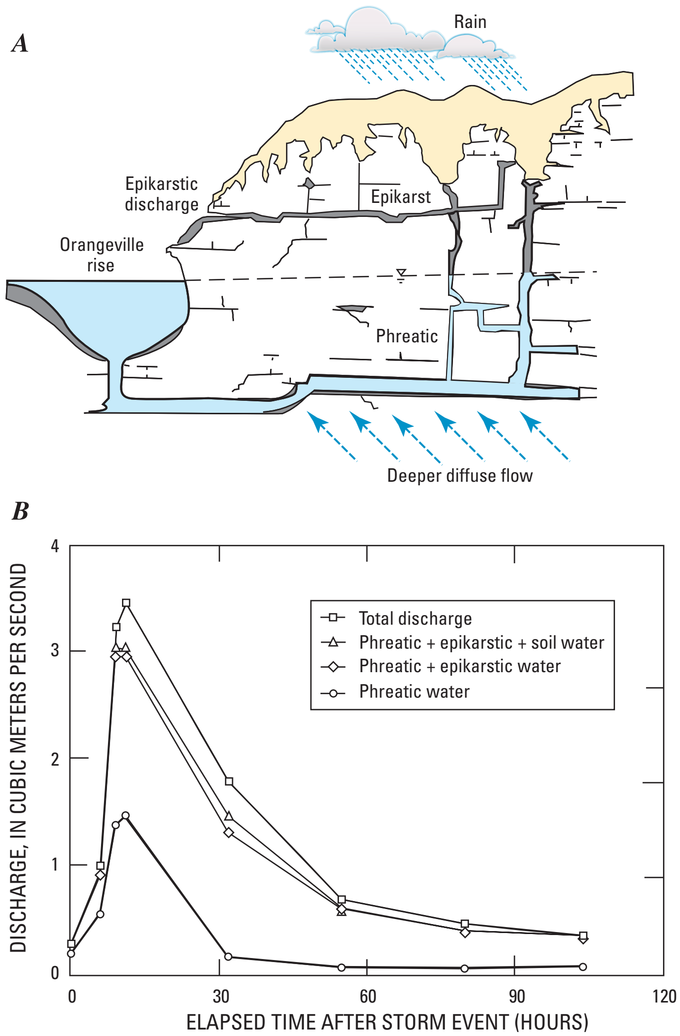


**Figure 11.** *A*, Conceptual model of the hydrologic components of the upper Lost River drainage basin in south-central Indiana; *B*, Four-component hydrograph separation curves at Orangeville Rise (from Lee and Krothe, 2001, reprinted from Chemical Geology, copyright 2001, with permission from Elsevier).

for N +1 sampling intervals of equal length $\Delta t$, where $y_i$ is the mean value of the output during the interval i; $x_j$ is the mean value of the input during the interval j, and $h_{i-j}$ is the kernel function during the interval i-j. Thus, if $x_j$ and $h_{i-j}$ are known, then $y_i$ can be determined directly by convolution. If $x_j$ and $y_i$ can be identified, then $h_{i-j}$ (kernel function) can be determined through deconvolution (Dooge, 1973; Dreiss, 1982, 1989).

Identification of the kernel function from hydrological data is difficult because of the nonlinearities in the hydrological system and errors in the measured data. Then the convolution relation becomes:

$$y_i = \Delta t \sum_{j=0}^{i} x_j h_{i-j} + \varepsilon_i \qquad i = 0,1,2,.....N \qquad (10)$$

where $\sum_i$ are the sum of the residual errors. In this case, the identification of the kernel function $h_{i-j}$ is more of an optimization problem and is found by minimizing the sum of the square errors. Methods of identifying the kernel functions in karst aquifers, in addition to issues involved in defining and working with these functions to predict flow and nonpoint source contamination, are presented in Dreiss (1982, 1989) and Wicks and Hoke (2000).

Wicks and Hoke (2000) applied and expanded the application of linear systems analysis to predict the changes in quantity and quality of water from a large karstic basin. Wicks and Hoke (2000) were able to predict the first arrival time and dispersion of solute discharged from a spring when injected into a specific point (fig. 13).

Long and Derickson (1999) applied a linear systems analysis approach to the karstic Madison aquifer in the Black Hills, South Dakota, to investigate the aquifer's response (head) to an input function. In this instance, stream loss (recharge), which was modeled by using a transfer function, could be related to the total memory length of the karst system (fig. 14). This method could be used as a response-to-recharge event-prediction tool in karst aquifers.

## Lumped-Parameter Models

In some karst basins, a linear response (kernel function) cannot adequately simulate the spring outflow. The purpose of lumped-parameter models is to simulate the temporal variations in discharge from springs. When the discharge rate varies continuously and depends on hydrologic input processes of precipitation, sinking streams, evapotranspiration, and infiltration, a model can be developed that produces the output based on some or all of the inputs (Zhang and Bai, 1996). One of the most common nonlinear, lumped-parameter models

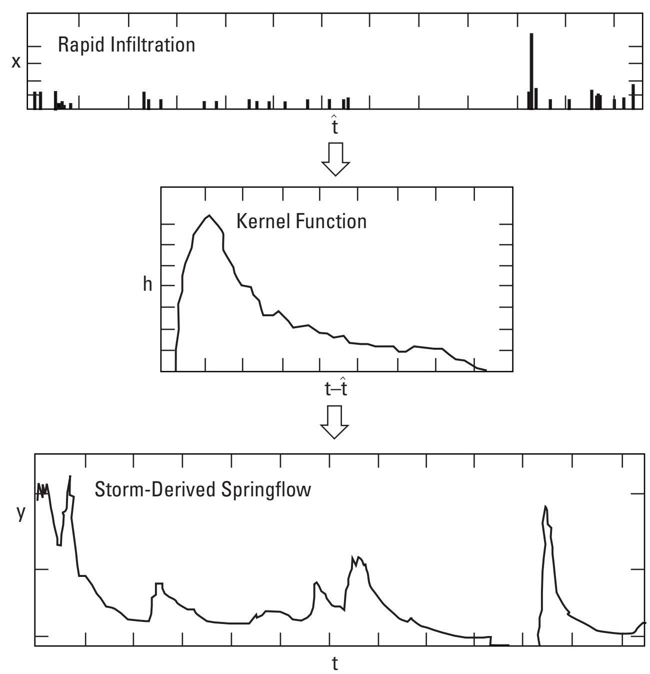


**Figure 12.** Linear system analysis of a karst conduit spring showing the recharge-discharge relation (after Dreiss, 1989).

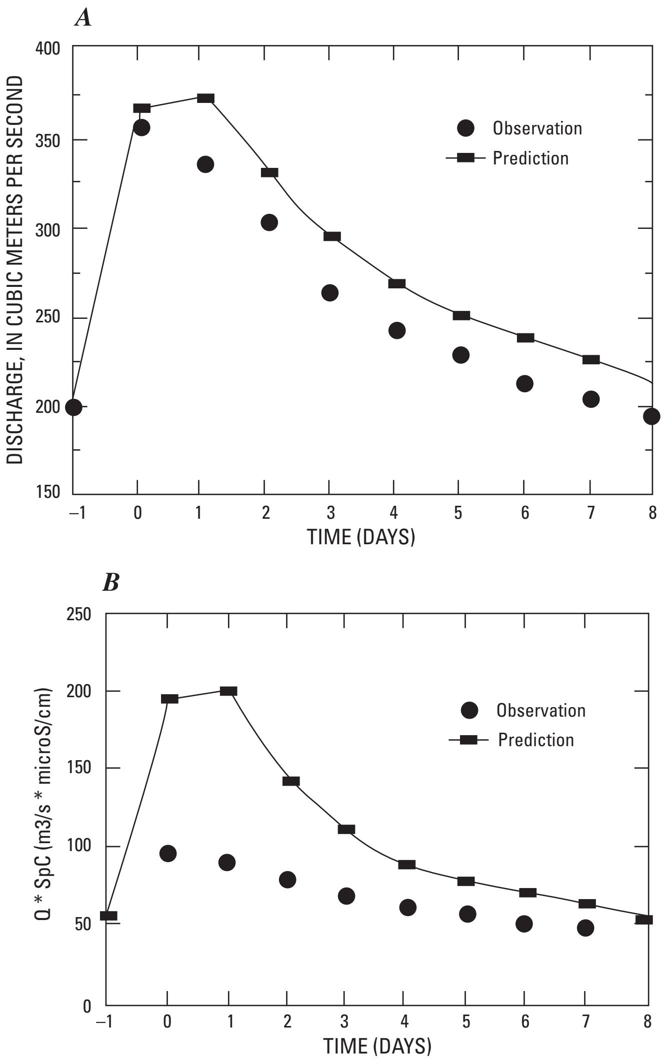


**Figure 13.** Predicted and observed (*A*) discharge at Maramec Spring, Missouri; and (*B*) specific conductance times spring discharge water (after Wicks and Hoke, 2000). Reprinted from Ground Water with permission from the National Ground Water Association, copyright 2000.

is the Hammerstein Model, and its use is demonstrated by Zhang and Bai (1996) and Stoica and Soderstrom (1982). The Hammerstein Model is a particularly good, general method for developing a lumped-parameter model for a karst basin. The model uses a least-squares approach to solve for coefficients in the Auto-Regressive Moving Average (ARMA) model and then is used to simulate spring discharge.

Zhang and others (1995) developed a lumped-parameter model to simulate the temporal variations in discharge from Big Spring, Iowa. When precipitation is assumed to be the sole input, the simulated spring discharge matched poorly with the observed spring discharge. The match improved significantly when other variables were added to the model, such as evapotranspiration, infiltration, and snowmelt. Approaches using lumped-parameter models as demonstrated by these authors can be successfully used to simulate spring discharge.

# Deterministic (Numerical) Modeling

Numerical modeling has become an important, commonly applied tool for investigating and quantifying many complex hydrogeological relations. However, many technical and conceptual difficulties remain to be solved to facilitate the discretization of conduit geometry or karst basin boundaries, parameterization of rapid- and slow-flow karst components, and simulation of temporal or spatial changes in saturation and flow conditions.

The use of deterministic models is most problematic in quick-flow or conduit-dominated karst systems. Data requirements for parameterization and proper model calibration of conduit-dominated flow are difficult to meet (Teutsch and Sauter, 1991; White, 1999). At present, the technical modeling capabilities and experience base needed to support such applications typically are lacking. Conduit-flow modeling codes are under development that may be of use in studies where the geometry of the conduit system can be fairly accurately mapped (Liedl and others, 2003). Some successes in simulating the effects of conduit flow have been achieved using a modified double-porosity modeling approach (Teutsch and Sauter, 1991) and by embedding high transmissivity zones within the grids of finite-difference or finite-element models (Worthington, 2003; Kuniansky and others, 2001).

Some of the more successful applications of numerical modeling have been in the simulation of spring discharge. Scanlon and others (2003) evaluated two different equivalent porous-media approaches (lumped and distributed parameter) to simulate regional ground-water flow to Barton Springs in the Edwards aquifer, Texas. Both methods worked fairly well to simulate the temporal variability in spring flow

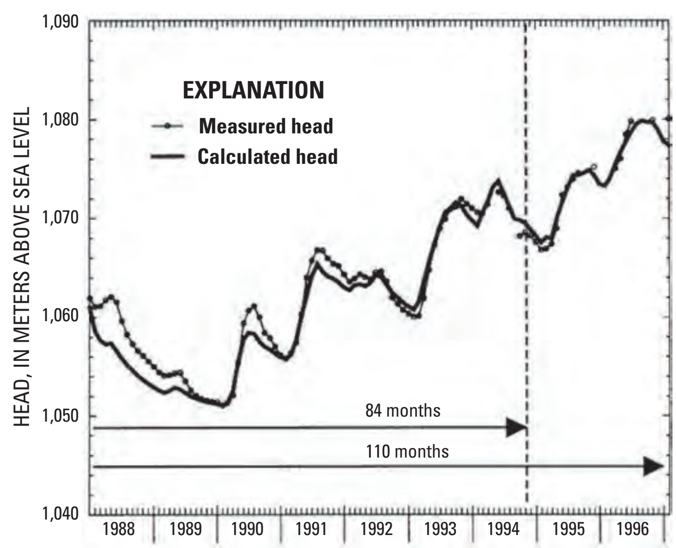


**Figure 14.** Calculated and measured head in an observation well, analysis is based on an 84-month time period that was used to predict a 110-month time period (after Long and Derickson, 1999). Reprinted from Journal of Hydrology, copyright 1999, with permission from Elsevier.

(fig. 15). The effect of pumping at a regional scale on spring discharge was best simulated by using a lumped-parameter distribution approach; however, a detailed evaluation of the effect of pumping on water levels and spring discharge required a distributed-parameter approach (Scanlon and others, 2003). Other successful results have been achieved in simulating karst aquifers dominated by slow-flow (diffuse-dominated) components, or in regional-scale studies where the effects of conduit-related heterogeneity can be minimized or neglected.

Deterministic rainfall-runoff models have been used successfully to estimate ground-water recharge and to simulate the hydrologic responses of watersheds in many non-karstic terranes (Beven, 2001; Cherkauer, 2004). Their possible application to karst hydrology studies seems promising but has received little attention thus far. As with deterministic ground-water models, a variety of technical and conceptual difficulties currently limit the use of these models. Larger, regional-scale modeling may be less problematic (Arikan, 1988). Available rainfall-runoff modeling codes such as TOPMODEL (Beven and Kirkby, 1979) and PRMS (Leavesley and others, 1983) are not well suited to dealing with issues related to internal drainage by sinkholes or routing of subsurface flow through conduits. Various “workarounds” such as filling and smoothing sinkhole depressions, or artificially inflating the volume or storage capacities of sinkholes, have been used experimentally to achieve model calibration (Campbell and others, 2003). These approaches have not been very successful, however, and have resulted mostly in models that do not accurately represent the physical hydrogeologic conditions in the karst basin or simulate the full range of observed flow conditions and hydrologic responses. Additional research aimed at improving the conceptualization and parameterization of karst flow systems in rainfall-runoff models is needed and would be beneficial.

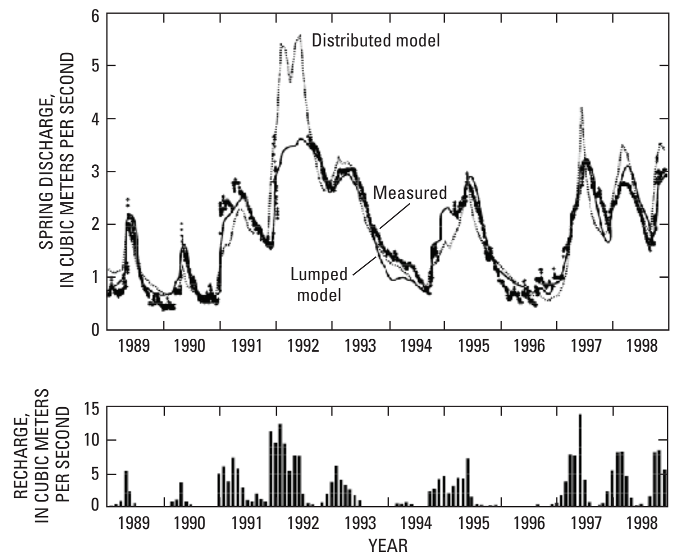


**Figure 15.** Simulation of Barton Spring discharge in the Edwards aquifer, Texas, using a lumped and distributed parameter approach (after Scanlon and others, 2003). Reprinted from Journal of Hydrology, copyright 2003, with permission from Elsevier.

## Water Tracing with Fluorescent Dyes

Water tracing with fluorescent dyes is a particularly useful tool for investigating water fluxes in karst flow systems because dye-tracer tests can be used to obtain direct information about flow direction, velocity, and other hydraulic characteristics in conduits between specific points of focused recharge and discharge. Fluorescent dyes are organic chemicals that absorb light from the ultraviolet part of the spectrum, are molecularly energized, and emit light at a longer wavelength range (Käss, 1998). As described by Smart and Laidlaw (1977) and Field and others (1995), an ideal water tracer is one that (1) is easy to introduce into the aquifer or flow system; (2) travels at or near the flow rate of water; (3) is relatively conservative—that is, not easily lost through sorption; (4) is stable with regard to water chemistry; (5) is easily detectable at low concentrations; and (6) has little or no toxicity to humans or aquatic organisms and poses no long-term intrinsic threat to the environment. As a group, fluorescent dyes possess almost all of these characteristics and, as such, they are widely used and popular choices for artificial tracer tests in karst studies.

Several of the fluorescent dyes most commonly used in water-tracing investigations in karst are listed in table 6. The fluorescent characteristics, detection limits in water, and sorption tendencies of these dyes are provided in table 7. Most of the individual dyes listed are members of a family or group of dyes that vary slightly in chemical structure and have overall similar fluorescent properties. The xanthene dyes are a large group that exhibit fluorescence in the green to orange wavelengths of the visible light spectrum (Käss, 1998) and includes such well-known tracer dyes as sodium fluorescein (also known as uranine), which fluoresces in the green wavelength band (500–570 nanometers [nm]), and Rhodamine WT, which fluoresces in the yellow-orange wavelength band (570–590 nm). Another large group of tracers is the stilbenes, or optical brighteners—compounds that technically are not dyes but are whitening agents—which fluoresce in the violet-blue wavelength range of the visible light spectrum (380–500 nm). Trade names of individual dyes may vary by manufacturer or supplier, so it is advisable to refer to a specific tracer dye using the Color Index (CI) generic name and the Chemical Abstracts Service (CAS) identification number (Field and others, 1995).

Successful use of fluorescent dyes in water-tracing studies requires at least a general working knowledge of the physical and chemical properties of individual dyes, and the conditions and limitations involved in their use. For example, fluorescence is pH and temperature sensitive; however, different dyes have different ranges of sensitivity and response to these properties. Certain dyes, such as sodium fluorescein, are particularly photosensitive, whereas others, such as Rhodamine WT, are not.

**Table 6.** Some commonly used fluorescent dye types, their dye names, and their respective Color Index and Chemical Abstracts Service (CAS) number (from Field, 2002b).

| Dye type and common name | Color index generic name | CAS No. |
|---|---|---|
| | Xanthenes | |
| sodium fluorescein | Acid Yellow 73 | 518-47-8 |
| eosin | Acid Red 87 | 17372-87-1 |
| | Rhodamines | |
| Rhodamine B | Basic Violet 10 | 81-88-9 |
| Rhodamine WT | Acid Red 388 | 37299-86-8 |
| Sulpho Rhodamine G | Acid Red 50 | 5873-16-5 |
| Sulpho Rhodamine B | Acid Red 52 | 3520-42-1 |
| | Stilbenes | |
| Tinopal CBS-X | Fluorescent Brightener 351 | 54351-85-8 |
| Tinopal 5BM GX | Fluorescent Brightener 22 | 12224-01-0 |
| Phorwite BBH pure | Fluorescent Brightener 28 | 4404-43-7 |
| Diphenyl Brilliant Flavine 7GFF | Direct Yellow 96 | 61725-08-4 |
| | Functionalized polycyclic aromatic hydrocarbons | |
| Lissamine Flavine FF | Acid Yellow 7 | 2391-30-2 |
| pyranine | Solvent Green 7 | 6358-69-6 |
| amino G acid | -- | 86-65-7 |

**Table 7.** Emission spectra and detection limits for dyes in water (modified from Field, 2002b).

[nm, nanometer; %, percent; µg/L, micrograms per liter]

| Dye name | Maximum excitation $\lambda$ (nm) | Maximum emission[1] $\lambda$ (nm) | Fluorescence intensity (%) | Detection limit[2] (µg/L) | Sorption tendency |
|---|---|---|---|---|---|
| sodium fluorescein | 492 | 513 | 100 | 0.002 | Very low |
| eosin | 515 | 535 | 18 | 0.01 | Low |
| Rhodamine B | 555 | 582 | 60 | 0.006 | Strong |
| Rhodamine WT | 558 | 583 | 25 | 0.006 | Moderate |
| Sulpho Rhodamine G | 535 | 555 | 14 | 0.005 | Moderate |
| Sulpho Rhodamine B | 560 | 584 | 30 | 0.007 | Moderate |
| Tinopal CBS-X | 355 | 435 | 60 | 0.01 | Moderate |
| Phorwite BBH Pure | 349 | 439 | 2 | -- | -- |
| Diphenyl Brilliant Flavine 7GFF | 415 | 489 | -- | -- | -- |
| Lissamine Flavine FF | 422 | 512 | 1.6 | -- | -- |
| pyranine | 460[3] | 512 | 18 | -- | -- |
| | 407[4] | 512 | 6 | -- | -- |
| amino G acid | 359 | 459 | 1.0 | -- | -- |
| sodium napthionate | 325 | 420 | 18 | 0.07 | Low |

[1]Approximate values only.

[2]Typical values for tracer detection in clean water using spectrofluorometric instrumentation.

[3]For pH greater than or equal to 10.

[4]For pH less than or equal to 4.5.

In addition, fluorescent dyes have varying ranges of reactivity with geological materials such as clays and other silicate minerals (Käss, 1998; Kasnavia and others, 1999). These and other important physiochemical characteristics always need to be considered prior to use. Useful references include Smart and Laidlaw (1977), Mull and others (1988), and Käss (1998). Although toxicity generally is not a great concern with most of the fluorescent dyes commonly used for water-tracing studies, this and other possible environmental concerns are reviewed by Smart and Laidlaw (1977) and Field and others (1995).

## Dye-Tracer Test Objectives and Design

Dye-tracer testing is a versatile method that can be employed in a number of ways by using various combinations of field and laboratory techniques that can be tailored to fit the specific objectives, context, and scale of the investigation (Smart, 2005). The basic goal of any dye-tracer test is to create a detectable fluorescent signal in water that can be positively identified as originating from the injected tracer dye and that can be interpreted in a manner needed to achieve the planned objectives of the test. Careful planning and execution of dye-tracer tests is essential so that positive, understandable results are obtained from each test—that is, from each injection of a tracer dye (Quinlan, 1989). If a dye is injected and not detected, the investigator may be faced with difficult and often costly decisions that may include whether or not to repeat the test, to change the type or amount of dye injected, to use additional monitoring sites, to conduct monitoring for a longer period of time, or to evaluate the sensitivity of the analytical method being used to detect the presence and (or) concentration of the injected dye.

In practice, dye-tracer tests generally are categorized as either quantitative or qualitative, depending largely on type of monitoring used and the data to be collected and interpreted (Jones, 1984a,b; Mull and others, 1988; Smart, 2005). Fully quantitative dye-tracer tests require accurate measurement of the amount (mass) of tracer dye injected, the discharge from the spring or aquifer during the test, and the concentration or total mass of tracer dye resurging from the aquifer. Quantitative dye-tracer tests primarily are used to obtain information about the time-of-travel and breakthrough characteristics of the tracer dye—which are important to contaminant-related studies—and to investigate karst conduit structure and flow properties (Field and Nash, 1997). Provided that discharge is measured simultaneously with tracer concentration at all dye-resurgence points, tracer mass recovery can be determined and used to make reliable estimates of conduit hydraulic properties including mean residence time, mean flow velocities, longitudinal dispersion, and storage (Field, 2002b).

In contrast to quantitative dye-tracer tests, qualitative dye-tracer tests are those that require only a determination of positive or negative resurgence of injected tracer dye at monitoring sites used for the test (Jones, 1984a). Qualitative dye-tracer tests are usually conducted to identify flow connections between focused points of recharge and discharge (for example, a sinkhole and a spring), thereby helping to delineate the trajectories of subsurface flow paths and to estimate an approximate maximum time-of-travel (based on the sampling interval used). Discharge data typically are not collected, and the actual concentrations of dye resurging in water at each monitoring site may or may not be determined, depending on the analytical methods used. Monitoring for these types of tracer tests often is accomplished by using passive detectors made of an adsorptive media such as granular activated charcoal to trap the tracer dye.

Planning required for a dye-tracer test typically involves a careful review of available hydrogeologic information, selection of dye-injection and dye-monitoring sites, an assessment of ambient fluorescence and hydrologic (flow) conditions, and selection of a method or methods to be used for dye monitoring and detection that is appropriate for the objectives and category of tracer test (that is, quantitative or qualitative). Qualitative dye-tracing and quantitative dye-tracing methods are not mutually exclusive, and the two methods often are used in combination in many karst studies (Quinlan, 1989). In common practice, quantitative dye-tracer tests often are conducted after subsurface flow routes have been identified between specific input points and discharge points by qualitative dye-tracer tests.

During any dye-tracer test, it is important that all potential dye-resurgence sites be identified and monitored to ensure that complete recovery of tracer dye is achieved. The results of previously conducted dye-tracer tests are very useful in the planning of subsequent tests, so every effort needs to be made during the planning phase to identify and review existing dye-tracer test information. For studies intended to delineate subsurface flow paths or karst basin boundaries, previous dye-tracer test results, estimates of unit-base flow of local springs, and other types of available hydrogeologic mapping data, are helpful in establishing the size and boundary of the study area required for monitoring. If few subsurface flow routes have been dye-traced and karst basin boundaries have been only partly delineated, or are not known, it may be necessary to monitor many springs in the study area, even those thought to be improbable resurgence sites, to ensure detection of the injected tracer dye.

Information about local ground-water flow directions and hydraulic gradients is extremely useful in the planning and the interpretation of dye-tracer tests. Therefore, if suitable water-level or potentiometric-surface maps are not available, it is wise to conduct an inventory and synoptic water-level survey of wells in the study area prior to initiation of a dye-tracer test. For many studies, selected wells need to be incorporated in addition to springs and streams as potential dye-monitoring sites. A field reconnaissance also needs to be done prior to implementation of any dye-tracer test. This is often a necessary and underappreciated aspect of the planning process. During the reconnaissance, potential dye-injection and dye-monitoring sites can be located and inspected to identify any logistical issues that may affect the implementation of the planned tracer test. Springs identified on published topographic maps often are inaccurately located, and the number of spring outlets plotted on a topographic map of a given area can be underrepresented as well (Quinlan, 1989). A thorough spring inventory needs to be conducted as part of the tracer-test planning process by searching existing databases; by walking, wading, or boating along surface stream reaches within the selected study area; by consulting aerial photographs; and by interviewing local landowners.

Because of the rapid temporal changes in hydraulic gradients, flow velocities, and flow directions typical of many conduit-dominated karst aquifers, the results obtained during a specific dye-tracer test are, strictly speaking, representative only of the flow conditions existing at the time of the test. For this reason, some consideration needs to be given during the planning phase as to whether additional dye-tracer tests need to be conducted during specific high- or low-flow conditions. For practical reasons, most dye-tracer tests are conducted during moderate- or base-flow conditions. During low-flow conditions, greater losses of tracer dye, and longer resurgence times, can be expected than at high-flow conditions, because of sorption, low-flow velocities, and storage of dye in hydraulic “dead-zones.” Different issues may occur during high- or flood-flow conditions. Injected tracer dyes may become too diluted and resurge in springs at concentrations below detection limits, the increased turbidity may interfere with dye monitoring and detection, physical access to dye-monitoring sites may be hindered, and in-situ dye-monitoring equipment may be damaged by flooding. In addition, hydraulic damming of conduits caused by flooded streams may temporarily halt or delay the resurgence of tracer dyes.

## Dye Injection

Dyes are typically injected as a “slug” of known weight, volume, or mass (fig. 16). A principal cause of negative or inconclusive dye-tracer test results is the injection of an insufficient quantity of dye into the aquifer (Quinlan, 1989; Field, 2003). Proper determination of the amount of dye to inject also is needed to ensure that dye resurges at detectable but not unacceptably high concentrations, particularly in public or private water supplies, and that residual storage of dye in the aquifer is minimized. Because of concerns about the possible formation of the carcinogen diethylnitrosamine resulting from the use of Rhodamine WT dye (Steinheimer and Johnson, 1986), the USGS adopted a policy that the concentrations of Rhodamine WT should not exceed 10 µg/L during tracing tests of surface waters near public water intakes (Water Resources Division Memorandum No. 66.90 and 85.82). In a review of toxicity and other environmental data, Field and others (1995) suggested that the resurgent concentration of most commonly used tracer dyes should not exceed 1 to 2 mg/L (1,000–2,000 µg/L) for more than 24 hours at a point of ground-water withdrawal or discharge. These concentration limits, while desirable, may not always be possible to achieve because of the unpredictability of subsurface flow routes and field variables that affect the rate of transport, dispersion, and subsequent concentration of dye discharged through conduits.

Historically, a variety of equations have been devised to estimate the quantity of dye needed for tracer test injections, based largely on distance to the anticipated resurgence point and or estimated ground-water flow velocities. Most of these are difficult to apply in practice and do not provide a means for the investigator to predict and manage the resurgent concentration of tracer dye. These shortcomings are addressed in methods devised by Field (2003) and by Worthington and Smart (2003).

The Efficient Hydrologic Tracer-Test Design (EHTD) method by Field (2003) includes a computer program that estimates the amount of dye needed for injection and provides forward modeling capability needed to predict tracer-breakthrough curve characteristics and the time intervals needed for effective sampling of the passage of the dye pulse—information that is important to planning quantitative tracer tests. The EHTD method calculates the amount of dye needed for injection by using various forms of the advection-dispersion equation for open-channel flow, closed-conduit flow, and flow through porous equivalent media. The program enables the user to designate the mass of tracer dye to be injected and an injection flow rate. For open-channel and closed-conduit flow conditions, the EHTD method requires the following input values: (1) discharge at the sampling station (spring), (2) estimated longitudinal distance from the dye-injection site to the anticipated resurgence site, (3) estimated cross-sectional area of the discharge point (that is, spring or stream cross-sectional area), and (4) a sinuosity factor applied to straight-line estimates of the distance between a dye injection and potential dye resurgence site.

The method proposed by Worthington and Smart (2003) relies upon the empirically derived equations:

$$M = 19\,(LQC)^{0.95} \tag{11}$$

and

$$M = 0.73\,(TQC)^{0.97} \tag{12}$$

where

$M$ is mass of tracer dye injected (grams/meter$^3$),

$L$ is anticipated distance between the injection site and the anticipated primary resurgence site (meters),

$Q$ is discharge at the anticipated resurgence (meters$^3$/second),

$C$ is peak tracer concentration at the anticipated resurgence (grams/meter$^3$),

and

$T$ is traveltime as determined from prior tracing-test results (seconds).

Using either equation, the investigator can select a target concentration desired for resurging tracer dye and solve to determine the required amount (mass) of dye needed for injection.

In practice, dye injection is best accomplished at locations that provide rapid, direct transport of the tracer into conduits, thus minimizing loss of dye through photochemical decay, sorption, or other field conditions. Open-throated swallets in sinkholes and the swallow holes of sinking streams are ideal sites. In the absence of naturally occurring runoff (inflow), dyes can be injected into a stream of potable water discharged from a tanker truck or large carboy. In general, 300 to 500 gallons of water are a minimum quantity needed for dye injection, and quantities of 1,000 gallons or more are preferable. Approximately one-half of the water is used to initiate flow into the swallet prior to dye injection. This is done to test the swallet’s drainage capacity, to initiate flow, and to flush the flow path to minimize losses to sorption. The remainder of the water is discharged after the dye is injected as a “chaser.” Under most conditions, this technique does not substantially change the naturally occurring flow conditions or alter hydraulic heads in the aquifer.

*A*

*B*

*C*

**Figure 16.** Dye injections: *A*, sodium fluorescein injection into collapse sinkhole formed in a pond (stream of water is outflow from a settling pond at a public supply water-treatment plant); *B*, Rhodamine WT injected into sinking stream; *C*, injection of Rhodamine WT into a water-level observation well (photographs by Charles J. Taylor, U.S. Geological Survey).

Several practical tips regarding dye injection that may be useful to consider during the planning of dye-tracer tests include the following:

- Open-borehole wells or screened wells can be used as dye injection sites. Although a pre-flush is not necessary prior to injecting the dye, it is advisable to conduct a falling-head slug test in order to test the hydraulic connection with the aquifer and to estimate the local hydraulic conductivity and rate of discharge of tracer from the well. After injecting dye, it is necessary to flush the dye from the well using several borehole volumes of potable water. It is important to control the volume and rate of water inflow during the flush to prevent the well from overflowing with dye-laden water.

- In locations where available sinkholes do not contain open swallets, dye injection may be accomplished by drilling a temporary injection well. Ideally, these injection wells would be drilled to intercept fractures or solutional openings in the bedrock; however, successful injection into the epikarst may be accomplished by completing the well at the top of the karstic bedrock. Aley (1997) discusses in detail the issues involved in conducting dye-tracer tests through the epikarst zone.

- Dye injections also can be made through the epikarst by excavating a pit into the soil; however, dye losses may be significant, and the quantity of dye used for the injection usually must be increased several times above the “normal” dosage amount.

- The injection of dye into flooded sinkhole depressions or swallets choked with sediment generally is not advisable, particularly for quantitative tracer tests. Unless there is evidence that drainage through the regolith into the subsurface is relatively rapid, excessive loss of the tracer dye may be incurred. If necessary, swallets that are partly choked with sediment may be cleared out with a shovel or backhoe and pre-tested for drainage capacity prior to an attempted injection of dye.

- A slug injection may not be an effective means of attempting to trace flow from a losing stream—that is, into a stream not fully diverted underground by swallow holes—because too much of the dye may be flushed downstream before it infiltrates the subsurface. Under such conditions, dye injection may be more effectively accomplished by using the continuous-injection technique described by Kilpatrick and Cobb (1984). A recent paper by Field (2006) specifically examines the problem of conducting dye-tracer tests from losing streams.

## Dye Monitoring and Detection

As dictated by the tracer-test objectives and the resources available, a variety of methods can be used to monitor for and detect the resurgence of injected tracer dyes, including direct visual observation, fluorometric analysis of discrete water samples or eluant obtained from granular activated charcoal detectors, and in-situ continuous-flow fluorometry. Dyes often can be visually detected in water at parts-per-million concentrations, whereas some method of fluorometric analysis is needed to detect dyes at subvisual concentrations. Three types of fluorometers are commercially available: scanning spectrofluorophotometers, filter fluorometers, and in-situ submersible fluorometers. Each of these instruments operates essentially by selectively measuring the fluorescent intensity of a sample (for example, water) that has been selectively excited (Duley, 1986). The selective range of light wavelengths used to excite the sample is called the excitation spectrum, and the selective range of light wavelengths that are measured by the fluorescent intensity is called the emission spectrum (Käss, 1998). Depending on which type of the three instruments is used, common tracer dyes can be detected in water at concentrations as low as parts per trillion—although environmental factors usually limit unequivocal detection to the range of parts per billion or greater (Smart and others, 1998).

Scanning spectrofluorophotometers are research-grade laboratory instruments that use a system of monochromators, diffraction gratings, and bandwidth slits to scan across user-selected excitation and(or) emission spectra at selected bandwidth intervals. These instruments are exceptionally sensitive—dyes often can be detected in the parts-per-trillion range—and enable precise characterization of the various sources of fluorescence in a sample. One advantage of these instruments is that they can be used to do synchronous scanning, a technique in which the excitation and emission monochromators are scanned together at a fixed wavelength difference determined by the separation (in nanometers) between the excitation and emission peaks for the dye(s) of interest (Duley, 1986; Rendell, 1987). For most xanthene dyes, this distance is approximately 20 to 25 nanometers (Käss, 1998). The synchronous scanning technique is useful for analyzing unknown mixtures of fluorescent solutes having various excitation wavelengths because it provides a spectral “fingerprint” for each solute present and therefore can be used to identify the presence of multiple tracer dyes in a single sample (Käss, 1998) (fig. 17). The tradeoff in using these instruments is that they require a good working knowledge of relatively specialized fluorescence spectroscopy methods and the use of rigorous quality-control methods, which can be quite time and labor intensive.

Filter fluorometers, such as the Turner Designs Model 11, are versatile instruments that can be set up to work under field or laboratory conditions, to analyze discrete samples of water or eluant or be used with flow-through cells and pumps to continuously monitor the change in fluorescence due to the passage of a dye pulse in water. These instruments have user-exchangeable glass filter kits that transmit light in the excitation and emission wavelengths of fluorescein or rhodamine dyes and are capable of identifying dyes at concentrations in the parts-per-billion range. Because the excitation and emission wavelengths are fixed by the set of filters installed, only one tracer dye at a time can be identified in a sample. Filter fluorometers are dependable “workhorse” instruments widely used for dye-tracer studies in karst. As discussed by Smart and others (1998), however, there are a number of practical constraints on the use of filter fluorometers, the most serious of which is that the sensitivity of these instruments may be adversely affected by ambient fluorescence so that it is possible to obtain apparently significant fluorescent readings—indicating positive detection of dye—with a particular filter set when the dye of interest is not actually present. Such false positive readings result from the presence of other fluorescent solutes having fluorescence spectral properties that overlap those of the tracer dye of interest (Smart and others, 1998).

Recently, manufacturers such as Turner Designs and Yellow Springs Instruments have introduced submersible fluorometers that can be used for in-situ continuous-flow monitoring of either sodium fluorescein or Rhodamine WT dyes. The filters needed for detection of each dye are preinstalled by the manufacturer in an optical probe assembly. These instruments include internal data loggers capable of recording thousands of data values in nonvolatile flash memory and simultaneously collect temperature and turbidity data as needed to correct the recorded fluorescent intensity values. Continuous-flow fluorometry conducted with either filter fluorometers or submersible fluorometers provides significant advantages for quantitative dye-tracer tests because highly resolved dye-breakthrough curves can be obtained whose properties are not as affected by sampling biases or by insufficient sampling frequency (Smart, 1998). Quality control also may be considerably improved because the water is analyzed directly without excessive sampling and handling activities that potentially increase the chances of sample contamination or degradation. These advantages sufficiently outweigh any potential loss of spectral precision (Smart and others, 1998).

For quantitative analysis, all fluorometers must be calibrated so that the concentration of dye is determined by the fluorescent intensity of the sample measured relative to that of dye-concentration standards. Standards are prepared from the tracer-dye stock solution by using gravimetric and serial dilution techniques in the manner described by Wilson and others (1986) or Mull and others (1988). Turner Designs offers a number of application notes that can be downloaded or ordered from their Internet website (*http://www.turnerdesigns.com*) that describe in detail the procedures involved in the preparation of dye standards and the calibration of filter fluorometers. The preparation of dye-concentration standards is a critical procedure and needs to be undertaken with great care. Errors introduced into the standards preparation process will adversely affect instrument calibration and, for quantitative tracer tests, may result in serious mass-balance errors (Field, 1999b). Adjustments to account for the percent actual tracer in powdered and liquid dyes, and also for specific gravity in liquid dyes (Field, 1999b, 2002b) (table 8) must be factored in during the calculation of dye standard concentrations. Dye-concentration calibration curves typically are made by using a logarithmic distribution of dye-concentration standards ranging over two or three orders of magnitude (Alexander, 2002).

In practice, error in dye detection and(or) determining dye concentration in water is dominated by issues of ambient (background) fluorescence, loss of tracer (that is, due to adsorption or photodegradation), and improper sampling frequency (Smart, 2005). Ambient (background) fluorescence is probably the largest single source of systematic error in dye

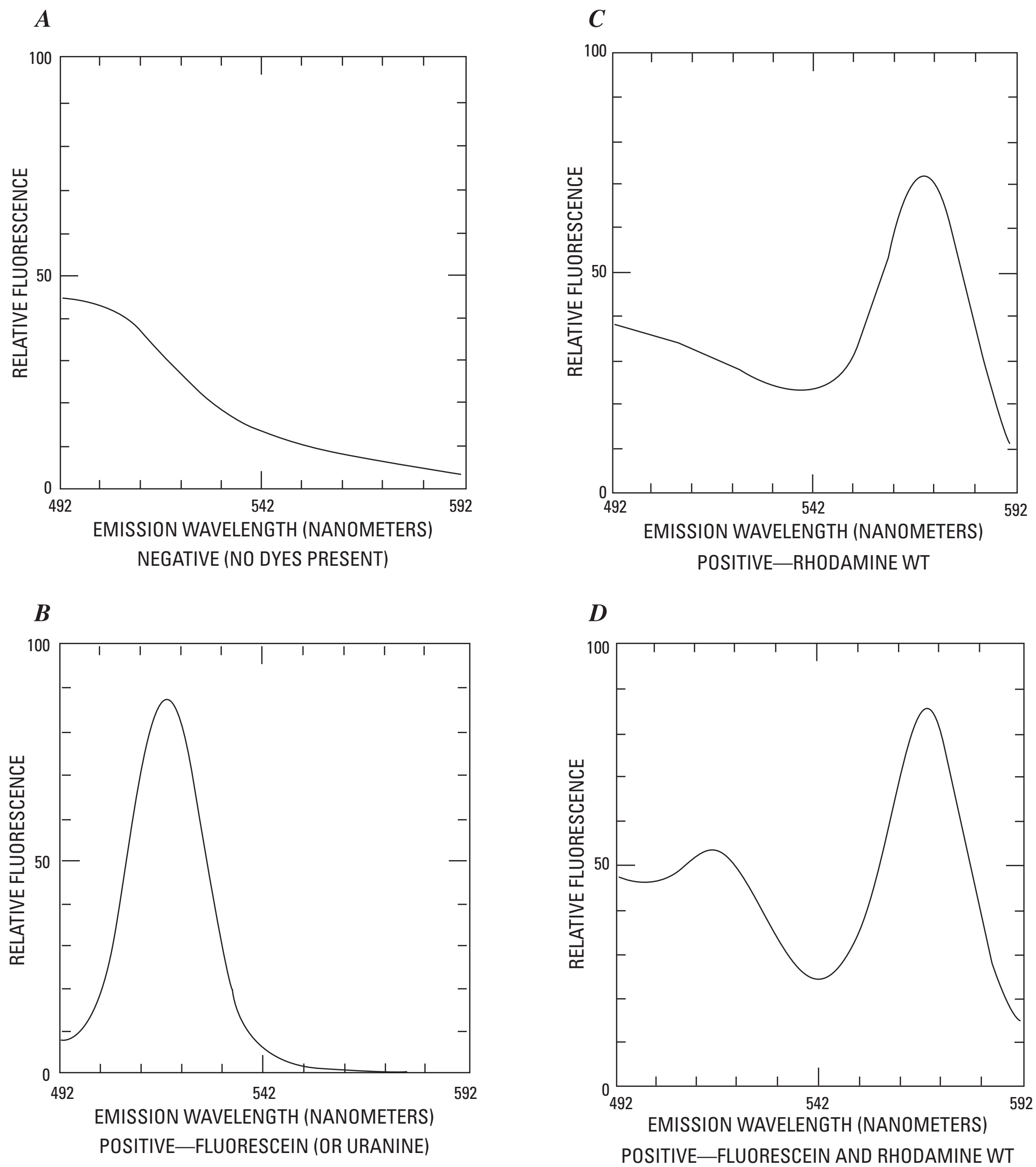


**Figure 17.** Dye spectral "fingerprints" obtained from use of the synchronous scanning method: *A*, no fluorescent tracer dye is present; *B*, sodium fluorescein (or uranine) tracer dye is present; *C*, Rhodamine WT tracer dye is present; *D*, sodium fluoresein and Rhodamine WT tracer dyes are present (after Vandike, 1992).

tracing and must be carefully assessed prior to initiation of any (qualitative or quantitative) dye-tracer test. For quantitative tests, ambient background levels occurring in the range of the emission peak of the tracer dye being used must be subtracted to accurately calculate dye concentrations. Often, the choice of dye selected for a particular tracer test is influenced by the presence and level of fluorescent intensity of ambient fluorescent interferences. Potential sources of interference with tracer dyes include naturally occurring humic and fulvic acids, certain species of algae, petroleum hydrocarbons, optical brighteners discharged in septic or treated waste-water effluent, automotive antifreeze chemicals (a widespread source of fluorescein), and hundreds of other dyes and organic chemicals used in industrial, commercial, and household products (Käss, 1998). In general, ambient background interferences typically are more problematic for optical brighteners and for xanthene dyes that fluoresce in the blue-green spectral wavelengths (Käss, 1998), and less problematic for xanthene dyes that fluoresce in the yellow-orange spectral wavelengths (Smart and Karunaratne, 2002).

One important point to consider is that the timing, duration, and intensity of fluorescence can vary considerably, depending on its sources, during the period over which ambient fluorescence is being monitored (Smart and Karunaratne, 2002). For this reason, it is advisable to conduct background monitoring for a period of at least several days or weeks immediately prior to initiating any dye-tracer test. It is also advisable to contact local, state, and Federal water-resources agencies at this time to determine whether or not other tracer tests are in progress or have recently been completed in order to be aware of, and avoid, interference and potential cross contamination with a previously injected tracer dye.

**Table 8.** Percent active tracer, and specific gravity, measured for some commonly used fluorescent dye tracers. These may vary from batch to batch and should be determined for the specific lot of dye being used for mass-balance calculations attempted during quantitative dye-tracing tests (Field, 2002b).

| Color index generic name | Powder dye (%) | Liquid dye (%) | Specific gravity (g cm$^{-3}$) |
|---|---|---|---|
| Acid Blue 9 | 74.0* | 37.0 | -- |
| Acid Red 52 | 90–90.2 | 18.0 | 1.175 |
| Acid Red 87 | 86.0 | 26.0 | -- |
| Acid Red 388 | 85.0** | 17.0 | 1.160 |
| Acid Yellow 73 | 60.0 | 30.0 | 1.190 |
| Basic Violet 10 | 90.0 | 45.0† | -- |
| Fluorescent Brightener 351 | 60.0 | -- | -- |

Values listed are equal to within 5.0 percent.

*Acid Blue also is sold with a Food, Drug and Cosmetic (FD&C) purity equal to 92.0%.

**Acid Red 388 is not commercially available in powder form.

†Basic Violet 10 as a liquid is mixed with glacial acetic acid.

Note: The values listed are specific to one manufacturer; crude dye stocks can and will vary significantly with manufacturer.

Because of ambient fluorescence and other analytical variables involved in fluorometry, there may be some subjectivity and difficulty in assessing the results of a single sample analysis—that is, the question might be asked “Is there enough of a change in fluorescent intensity to indicate that the tracer dye has been detected?” These decisions are made somewhat more objectively if minimum threshold concentrations or fluorescent intensity values are established by statistical methods or some other means to ensure that the fluorescence intensity or concentration measured in a sample is sufficiently higher than background to provide a high confidence level that dye was positively detected. From the literature, it appears that many researchers apply an arbitrary 10:1 signal-to-noise ratio for fluorescent intensity or dye concentration measured at the expected emission peak of the tracer dye as a minimal threshold for reporting the positive detection of dye in a sample (Smart and Simpson, 2001). A number of analytical and data post-processing techniques also have been devised in an attempt to enhance the detection of tracer dyes, particularly when working under “noisy” fluorescent background conditions (Smart and others, 1998; Smart and Smart, 1991; Lane and Smart, 1999; Tucker and Crawford, 1999). More recently, the use of advanced spectral analysis techniques has been explored as a means of better distinguishing tracer dyes from ambient background fluorescence (Alexander, 2005).

In general, caution needs to be used when making a determination that breakthrough and detection of an injected tracer dye has occurred based on only one “positive” sampling result. Evidence of dye breakthrough and detection is more conclusive if repeated positive detections are obtained, particularly where these results demonstrate a change in dye concentrations or fluorescent intensity that is indicative of the passage of a dye pulse and subsequent return to ambient fluorescent conditions. This is a principal reason that some researchers strongly recommend the use of quantitative dye-tracer tests methods, which include high-frequency sampling using automatic water-samplers or continuous-flow fluorometry, whenever possible (Field, 2002b; Kincaid and others, 2005). The passage of a dye pulse, however, also can be conclusively demonstrated by changes in fluorescent intensity or equivalent dye concentration obtained during qualitative dye-tracer tests using passive samplers, provided that a sufficiently high sampling frequency is used. Where questionable or inconclusive dye-tracer test results are obtained, it is advisable to review the tracer test design—particularly the methods used for monitoring and detection—and repeat the test using a different tracer dye.

## Use of Charcoal Detectors

As previously indicated, the use of passive detectors containing granular activated charcoal is a popular method of monitoring for dye resurgence during qualitative tracer tests. The detectors typically are constructed of fiberglass screen, nylon netting, or a similar material, fashioned into packets that contain several grams of charcoal sampling media. The size and shape of the packets and the amount of charcoal used in them is not particularly critical—the only requirements are that detectors be relatively durable, securely retain the activated granular charcoal, and allow water to flow easily and evenly through the packets. The detectors, or "bugs", generally are used in rapidly flowing water in streams and springs suspended above the substrate using a wire-and-concrete or wire-and-brick anchor. The detectors also can be staked or pinned directly into the streambed in very shallow water, and they can be easily suspended in monitoring or water-supply wells using, for example, monofilament line, snap-swivels, and steel shot-weights sold as fishing tackle. A common practice is to use detectors at all anticipated resurgences and exchange the detectors at 2 to 10 day intervals throughout the duration of the test (Quinlan, 1989). As a practical matter, it is generally inadvisable to leave detectors in the field longer than 10 days because of physical degradation that can diminish the adsorptive capability of the charcoal.

The principal advantage of using charcoal detectors is their economy and relative ease of use for ground-water reconnaissance studies, for simultaneous monitoring of many potential dye resurgence sites, and for mapping of conduit-flow paths or karst basin boundaries in areas where these are primarily or completely unknown. The detectors are relatively easy to conceal, thus minimizing the potential for disturbance and vandalism; and they are inexpensive, most of the cost being associated with their use onsite, collection, and analysis (Smart and Simpson, 2001). Handling, storage, and transportation requirements used in the exchange of detectors are not particularly critical with the exception of simple procedures needed to eliminate the potential for misidentification of detectors or cross contamination during handling and to prevent degradation of the dyes adsorbed by the charcoal (Jones, 1984a).

Another benefit of using charcoal detectors is their ability to concentrate dye at levels 100 to 400 times greater than the concentration of dye resurging in water, thereby helping to increase the probability of a positive detection of dye at distant monitoring sites (Smart and Simpson, 2001, 2002). To expel adsorbed dye, a few grams of charcoal are removed from a detector packet and eluted in an alkaline-alcohol solution. Two popular eluants include the so-called "Smart solution" (Smart, 1972), prepared by mixing 1-propanol, distilled water, and 28 to 30 percent ammonium hydroxide in a 5:3:2 ratio, and a solution of 70 percent 2-propanol and 30 percent deionized water saturated with sodium hydroxide (Alexander, 2002). Upon mixing, the solution separates into a lighter (saturated) and denser (supersaturated) liquid and it is the lighter phase that is decanted off and used as the actual eluant. Other eluant formulas may be chosen to enhance the elution of specific tracer dyes (Käss, 1998). Any prepared eluant always needs to be scanned as a blank before actual use (Alexander, 2002). Generally, 1 hour of elution is needed before fluorometric analysis can be done, although different dyes have different optimal elution times in various eluants (Smart and Simpson, 2001). Elution may be done with wet or dry charcoal; for longer term storage, however, charcoal samples need to be completely dried to prevent microbial degradation of the adsorbed dye.

If dye elutes from the charcoal below visible concentrations, an aliquot of the eluant can be removed for analysis by using either a filter fluorometer or a scanning spectrofluorophotometer (Smart and Simpson, 2002). As with water samples, the relative concentration of tracer dye in the eluant is determined by the fluorescent intensity, or the area of the spectral peak, measured at the emission wavelength of the dye (Jones, 1984b; Smart and Simpson, 2002). Because fluorescence for most tracer dyes is pH-dependent, the emission wavelengths for dyes in alkaline-alcohol eluants generally are shifted several nanometers relative to the emission wavelengths reported for dyes in water samples at or near neutral pH (Käss, 1998), and the emission peak characteristics and calibration curves obtained by fluorometric analysis may vary for different eluant formulations.

Although the use of charcoal detectors is relatively easy and has many potential benefits, the method is not without its shortcomings. Variables in the field, differences in the adsorptive efficiency of charcoal with various tracer dyes, complexities associated with the adsorption-desorption process of organic solutes on charcoal, and other variables introduced as a result of processing in the laboratory, preclude any determination of the actual concentrations of tracer that resurged in water and the replication of analytical results obtained from eluted charcoal (Smart and Simpson, 2001, 2002). The amount of dye concentrated on the detectors is a factor of the rate of flow through the detectors, the total surface area exposed to the dye, and of the length of time of the exposure. Dye concentrations measured in eluant also are affected by the time and method of elution (Smart and Simpson, 2001), therefore the concentrations of dye measured in eluant have a nonlinear, nonquantifiable relation to the concentrations of dye resurging in water. It is primarily because of these difficulties that some researchers, such as Field (2002b), have expressed reservations about the use of charcoal and strongly advocate the collection and quantitative analysis of water samples, or use of in-situ continuous-flow fluorometery during dye-tracer tests. Assuming that a qualitative tracer test design will meet the objectives for the study, many of these difficulties can be overcome by careful evaluation of ambient fluorescence, careful tracer-test design, proper application of analytical methods, and the application of rigorous QA/QC techniques during all field and laboratory activities. All of these issues deserve careful consideration during the planning phases of a dye-tracer test.

Proper evaluation of ambient fluorescence (background) is even more critical with activated charcoal than with water samples. When used in the field, activated charcoal captures a broad range of organic molecules, and a complex hierarchy of adsorption occurs based on the range of adsorptive sites, their accessibility, and the loading (composition and duration of flow) (Smart and Simpson, 2001). As with tracer dyes, these solutes will be recovered on the charcoal at substantially higher levels than the concentrations present in the water.

During spectrofluorometric analysis, the fluorescent signatures created by these solutes may be confused with, or mask, dye spectral peaks (Smart and others, 1998). High levels of organic solutes can foul the detectors because the solutes can consume the available adsorptive capacity of the carbon. Older charcoal tends to be less adsorptive than fresh charcoal, because of denaturing of the more energetic adsorption sites and capture of organic molecules from the surrounding atmosphere (Smart and Simpson, 2001, 2002).

Unfortunately, there is no ready means of distinguishing a genuine tracer recovery from accidental contamination of the charcoal detector (Smart and others, 1998). Wood charcoals, including coconut shell used to manufacture granular activated charcoal, can contain 10 to 20 percent fluorescein-type functional groups, which may create apparent false-positive peaks for sodium fluorescein dye when eluted (Alexander, 2002); however, this problem is usually manageable in that the compounds generally have a weak fluorescent intensity and seem to be flushed from charcoal by 1 to 2 days exposure to flowing water (Smart and Simpson, 2001).

## Dye-Breakthrough Curve Analysis

Analysis of dye-breakthrough curves (measured dye concentration over time) obtained via quantitative dye-tracer tests is an effective means of determining conduit-flow characteristics in karst aquifers (Smoot and others, 1987). Advantages provided by using this method, listed by Kincaid and others (2005), include:

- Plotting of the increase and decrease in fluorescence increases the confidence that tracer-test results are accurate and reflect the actual passage of the injected tracer dye through the aquifer.
- More accurate estimates of flow velocity can be calculated using time-to-peak concentrations.
- Integrating the area under the dye-breakthrough curve allows for estimation of the mass of tracer recovered at a sampling site and, therefore, the relative contribution of flow from the injection site to the tracer resurgence site.
- If it can be assumed that 100 percent of the tracer dye was recovered, evaluation of the shape of the dye-breakthrough curve provides data needed for estimation of hydraulic properties such as longitudinal dispersion, Reynolds and Peclet numbers, and discharge.

Important characteristics of the dye-breakthrough curve (fig. 18) include the first arrival or time to the leading edge of the dye pulse, time to peak concentration, elapsed time of passage of the dye pulse, and time to trailing edge or passage of the dye pulse. As Field (1999a) notes, these characteristics are not entirely objectively defined because they are dependent on sampling frequency and instrument sensitivity. Apart from sampling frequency bias, the shape and magnitude of the dye-breakthrough curve are most influenced by: (1) the amount of dye injected, (2) the velocity and magnitude of the flow, (3) internal structure and hydraulic properties of the conduit flow path taken by the tracer dye, and (4) other factors that affect mixing and dispersion of the tracer dye in the aquifer (Smart, 1998; Field, 1999a). Thus, the dye-breakthrough results obtained represent the transport characteristics of the tracer dye under the hydrologic conditions occurring during a particular test. Repeated quantitative tracer tests may be needed to characterize tracer dye characteristics under different flow conditions. Normalized dye-concentration and dye-load curves are used to compare and evaluate the transport characteristics of dye under different hydrologic conditions (Mull and others, 1988).

The physical properties of the dye-breakthrough curve provide information about conduit structure and organization (Smart, 1998). The dispersion of a dye plume increases with time and distance, and the pattern of dye recovery obtained reflects the effects of processes such as dilution, longitudinal dispersion, divergence, convergence, and storage, which are related to discharge and conduit geometry. The effects of longitudinal dispersion of the dye pulse usually are seen as a lengthening of the breakthrough curve (“tailing”), and the effects of tracer retardation usually are seen as multiple secondary peaks in dye concentration along the profile of the breakthrough curve. Interpretation of complex or multipeaked dye-breakthrough curves may be difficult because the factors contributing to tracer dispersion or retardation may include anastomosing (bifurcation or braiding) conduit-flow paths; flow reversal in eddies and variability in conduit cross-sectional areas (Hauns and others, 2001); intermittent storage and flushing of hydraulically stagnant zones (Smart, 1998); and interconnected zones of higher and lower fracture permeabilities (Shapiro, 2001). The potential effects of such factors on the shapes of dye-breakthrough curves under high-flow and low-flow conditions are illustrated in figure 19. Interpretation of the physical characteristics of the breakthrough curves usually cannot be based solely on the pattern of recovery of dye, but also on knowledge of the physical hydrogeology and conduit structure in the karst aquifer under study (fig. 20) (Jones, 1984b).

A variety of hydraulic properties, including the hydraulic radius or (assuming open-channel flow conditions) hydraulic depth, Peclet number, Reynolds number, Froude number, and hydraulic head loss can be estimated using dye-breakthrough curve data if it can be assumed that nearly 100 percent of the tracer dye was recovered (Field, 1999a; Mull and others, 1988; Field, 2002b). The computer program QTRACER2 (Field, 2002b), automates curve plotting and facilitates many of the calculations involved in the dye-breakthrough curve analysis obtained by analysis of dye-breakthrough curve data.

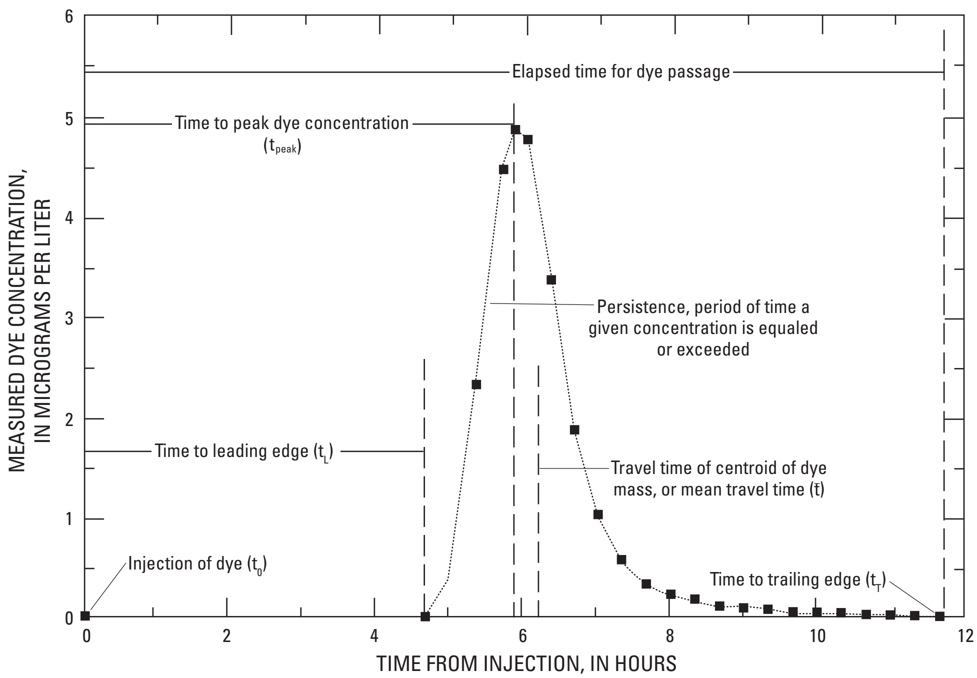


**Figure 18.** Some important physical characteristics of a dye-breakthrough curve (from Mull and others, 1988).

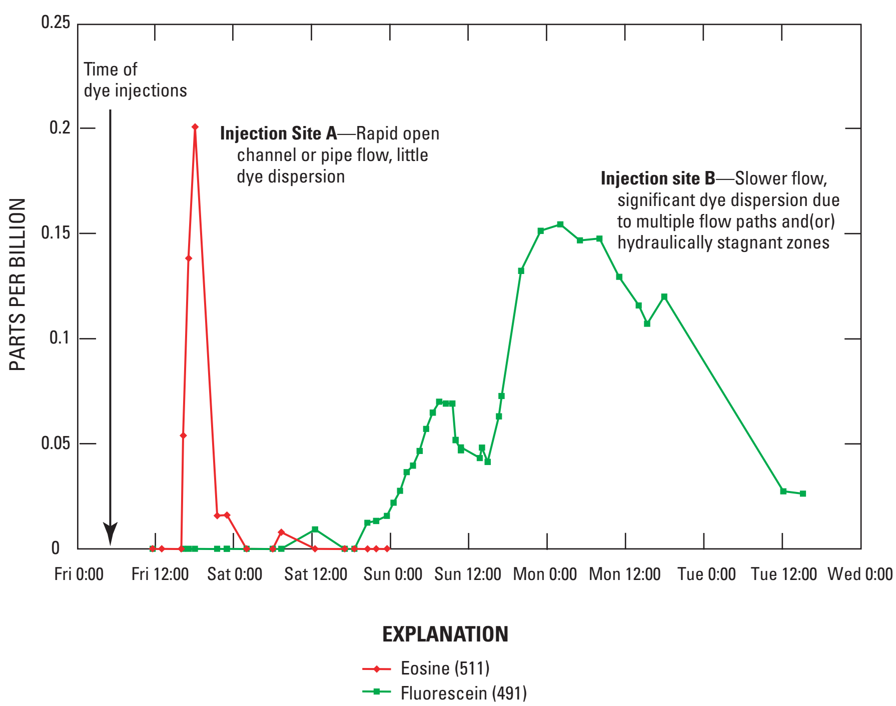


**Figure 19.** Example of dye-breakthrough curves for two dye-tracing tests conducted in the Edwards aquifer, Texas, showing a quick-flow response with little or no dispersion (Injection site A, left), and a slow-flow response showing the effects of dye dispersion (Injection site B, right) (courtesy of Geary Schindel, Edwards Aquifer Authority).

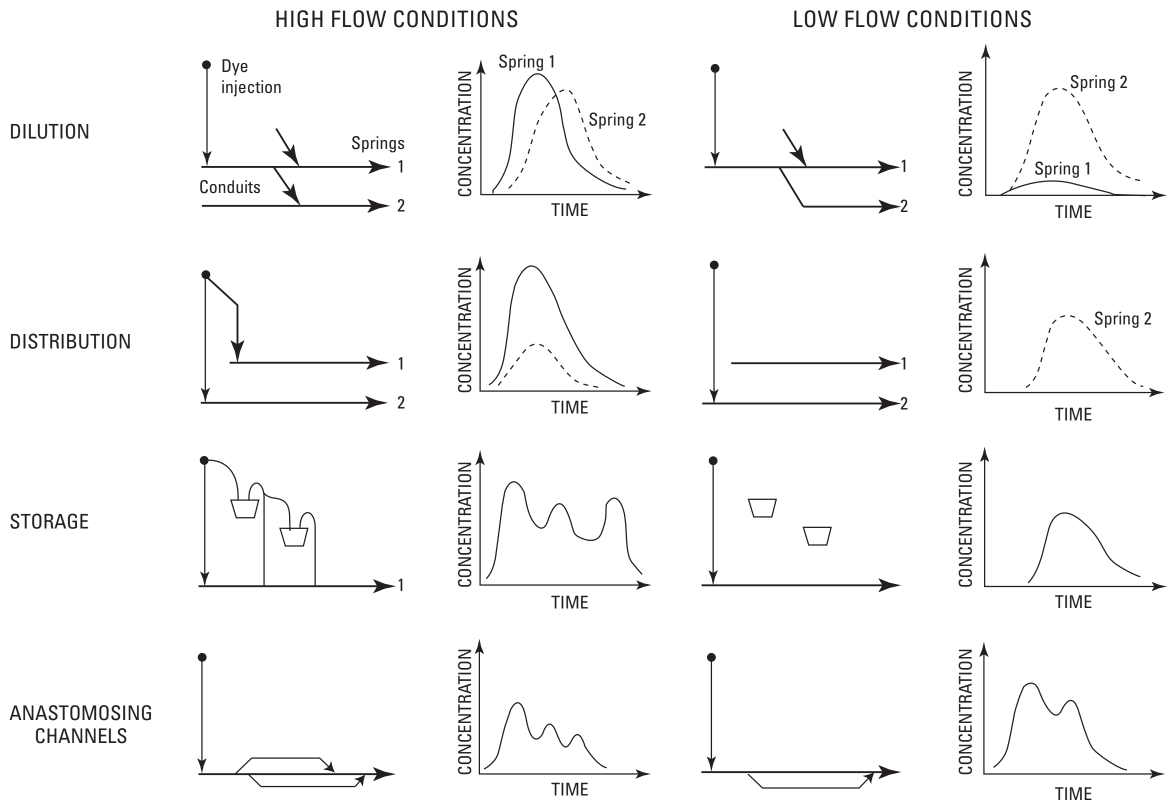


**Figure 20.** Shapes of hypothetical dye-breakthrough curves affected by changes in hydrologic conditions (high flow, low flow) and conduit geometry (modified from Jones, 1984b, after Smart and Ford, 1982). Used with permission from the National Speleological Society (www.caves.org).

## Mean Tracer-Dye Residence Time

Mean tracer-dye residence time is the length of time required for the centroid (gravity mass) of the tracer dye to traverse the entire length of the karst basin, thus representing the average time of flow through the basin. The centroid generally is not the same as the peak concentration of the tracer-dye mass in the tracer-breakthrough curve, but the more the dye plume conforms to Fick's law (the mass of the diffusing substance passing through a given cross section per unit time is proportional to the concentration gradient) the less obvious the difference between the dye centroid and peak concentration will be.

Mean tracer-dye residence time is estimated by the equation:

$$t_m = \int_0^\infty t\,C(t)Q(t)\,dt \Big/ \int_0^\infty C(t)Q(t)\,dt, \qquad (13)$$

where

$t$ is time of sample collection,

$C(t)$ is measured dye concentration of the sample,

and

$Q(t)$ is the discharge measured at the sampling location.

Tracer-dye residence time will vary from nearly zero for instantaneous transport to almost infinity where the tracer mass is mostly lost to dispersion or storage in the aquifer. If QTRACER2 or another suitable mathematical software program is not used, and the sampling frequency was done at regularly spaced intervals, the integration can be done by using a simple summation algorithm as detailed in Field (2002b) and by Mull and others (1988).

## Mean Dye Velocity

Mean tracer velocity (of the dye mass centroid) represents the average rate of travel of dye through the karst basin and is estimated by:

$$V_{(M)} = \int_0^\infty (1.5x/t)C(t)Q(t)\,dt \Big/ \int_0^\infty C(t)Q(t)\,dt\,, \qquad (14)$$

where

$x$ is straight-line distance between the dye injection and resurgence site,

and

1.5 is a constant representing the conduit sinuosity factor (Field, 1999a).

## Tracer Mass Recovery

The accuracy of calculations of mean tracer-dye residence time, flow velocities, and other conduit hydraulic properties from dye-breakthrough curve data is entirely dependent on tracer mass recovery. Few tracing tests result in 100 percent recovery of dye, but as the percentage of mass recovery decreases, the margin of error in the calculated hydraulic parameters increases and confidence in the values obtained declines. Tracer recovery may be affected by the internal structure of conduit networks (Brown and Ford, 1971; Atkinson and others, 1973). It therefore is important to assess tracer mass recovery as a starting point in the analysis of quantitative dye-tracing tests.

The quality of the tracer experiment may be quantified in terms of the relation between the mass of dye tracer injected ($M_{in}$) during the experiment and the total mass of dye tracer recovered ($M_r$). A test accuracy index proposed by Sukhodolov and others (1997) is calculated by:

$$A_I = M_{in} - M_r / M_{in} \tag{15}$$

This index provides a semiquantitative assessment of the quality of the test. A value $A_I = 0$ indicates a perfect tracing experiment with no loss of tracer dye mass. A positive $A_I$ value indicates that more tracer dye mass was injected than was recovered—a common result, whereas a negative value indicates more dye mass was recovered than was injected—an impossibility unless residual tracer dye is present in the aquifer, errors are made in determining the dye concentration in test samples, or initial calculations of the injected dye mass are in error.

In the previous equation, the value for $M_r$, the total mass of tracer dye recovered is given by the equation:

$$M_r = \int_0^\infty C(t)Q(t)\,dt. \tag{16}$$

A simple summation algorithm can be used to facilitate the calculations needed to obtain the value for $M_r$ as described by Field (2002b):

$$M_r = \int_0^\infty C(t)Q(t)\,dt \approx, \tag{17}$$

$$\sum_{i=1}^{n} C(i)Q(i)\,\Delta t_i \approx, \tag{18}$$

and

$$t_c \sum_{i=1}^{n} (C_i Q_i), \tag{19}$$

where

$t_c$ is a time conversion needed to obtain units of mass only.

The previous equations assume that the total dye mass is recovered at a single spring site. If dye has resurged at multiple spring outlets, these calculations are repeated for each site and the results are summed to obtain $M_r$.

# Summary

The hydrogeologic complexities presented by karst terranes often magnify the difficulties involved in identifying and measuring or estimating water fluxes. Conventional hydrogeologic methods such as aquifer tests and potentiometric mapping, though useful, are not completely effective in identifying the processes involved in the transfer of water fluxes in karst, or in characterizing the hydrogeologic framework in which they occur, and may provide erroneous results if data are not collected and interpreted in the context of a karst conceptual model. In karst terranes, a greater emphasis must generally be placed on the identification of hydrologic boundaries and subsurface flow paths, contributions of water from various concentrated and diffuse recharge sources, the hydraulic properties of conduits, and the springs that drain conduit networks. Typically, this emphasis requires the use of a multidisciplinary study approach that includes water-tracer tests conducted with fluorescent dyes and the analysis of spring-discharge and water-chemistry data.

The concepts and methods discussed in this chapter are intended to assist the water-resources investigator in determining what types of data-collection activities may be required for particular karst water-resources management and protection issues, and may aid the planning and implementation of karst hydrogeologic studies. The conceptual model of a karst drainage basin, described herein as a fundamental karst mapping unit defined by the total area of surface and subsurface drainage that contributes water to a conduit network and its outlet spring or springs, may be useful in this regard.

# References

Alexander, E.C., Jr., 2002, Scanning spectrofluorophotometer methods: Minneapolis, University of Minnesota Department of Geology and Geophysics (unpublished document dated 5/12/02), 10 p.

Alexander, S.C., 2005, Spectral deconvolution and quantification of natural organic material and fluorescent tracer dyes, *in* Beck, B.F., ed., Sinkholes and the engineering and environmental impacts of karst, Reston, Virginia: American Society of Civil Engineers, Geotechnical Special Publication No. 144, p. 441–448.

Aley, T., 1977, A model for relating land use and groundwater quality in southern Missouri, *in* Dilamarter, R.R., and Csallany, S.C., eds., Hydrologic problems in karst regions: Bowling Green, Western Kentucky University, p. 232–332.

Aley, T., 1997, Groundwater tracing in the epikarst, *in* Beck, B.F., and Stephenson, J.B., eds., The engineering geology and hydrogeology of karst terranes: Rotterdam, Netherlands, A.A. Balkema, p. 207–211.

Arikan, A., 1988, MODALP—A deterministic rainfall-runoff model for large karstic areas: Hydrological Sciences Journal, v. 33, no. 4, p. 401–414.

ASTM Subcommittee D–18–21, 2002, Standard guide for design of ground-water monitoring systems in karst and fractured-rock aquifers: West Conshohocken, Pa., American Society of Testing and Materials, Annual Book of ASTM Standards, v. 04.08, ASTM D5717–95, p. 1421–1438.

Atkinson, T.C., 1977, Diffuse flow and conduit flow in limestone terrain in Mendip Hills, Somerset, England: Journal of Hydrology, v. 35, p. 93–100.

Atkinson, T.C., Smith, D.I., Lavis, J.J., and Whitaker, R.J., 1973, Experiments in tracing underground waters in limestones: Journal of Hydrology, v. 19, p. 323–349.

Baedke, S.J., and Krothe, N.C., 2001, Derivation of effective hydraulic parameters of a karst aquifer from discharge hydrograph analysis: Water Resources Research, v. 37, no. 1, p. 13–19.

Bassett, J., 1976, Hydrology and geochemistry of the upper Lost River drainage basin, Indiana: National Speleological Society Bulletin, v. 38, no. 4, p. 79–87.

Bayless, E.R., Taylor, C.J., and Hopkins, M.S., 1994, Directions of ground-water flow and locations of ground-water divides in the Lost River watershed near Orleans, Indiana: U.S. Geological Survey Water-Resources Investigations Report 94–4195, 25 p., 2 pls.

Beck, B.F., ed., 1995, Karst geohazards: Proceedings of the Fifth Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, Gatlinburg, Tenn., April 2–5, 1995: Rotterdam, Netherlands, A.A Balkema, 581 p.

Beck, B.F., 2002, Overview of background resources on karst research emphasizing obscure and “grey area” documents, *in* Kuniansky, E.L., ed., U.S. Geological Survey Karst Interest Group Proceedings, Shepherdstown, West Virginia, August 20–22, 2002: U.S. Geological Survey Water-Resources Investigations Report 2002–4174, p. 7.

Beck, B.F., ed., 2003, Sinkholes and the engineering and environmental impacts of karst, Proceedings of the Ninth Multidisciplinary Conference, September 6–10, 2003, Huntsville, Alabama: American Society of Civil Engineers, Geotechnical Special Publication no. 122, 737 p.

Beck, B.F., and Stephenson, J.B., eds., 1997, The engineering geology and hydrogeology of karst terranes, Proceedings of the Sixth Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, Springfield, Missouri, April 6–9, 1997: Rotterdam, Netherlands, A.A Balkema, 516 p.

Beck, B.F., Pettit, A.J., and Herring, J.G., eds., 1999, Hydrogeology and engineering geology of sinkholes and karst—1999, Proceedings of the Seventh Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, Harrisburg-Hershey, Pennsylvania, April 10–14, 1999: Rotterdam, A.A Balkema, 478 p.

Beven, K.J., 2001, Rainfall-runoff modelling—The primer: New York, John Wiley and Sons, Ltd., 360 p.

Beven, K.J., and Kirkby, M.J., 1979, A physically based, variable contributing area model of basin hydrology: Hydrological Science Bulletin, v. 24, no. 1, p. 43–69.

Bonacci, O., 1993, Karst spring hydrographs as indicators of karst aquifers: Hydrological Sciences-Journal-des Sciences Hydrologiques, v. 38, no. l, p. 51–62.

Bogli, A., 1980, Karst hydrology and physical speleology: New York, Springer-Verlag, 284 p.

Brown, M.C., and Ford, D.C., 1971, Quantitative tracer methods for investigation of karst hydrologic systems: Transactions of the Cave Research Group of Great Britain, v. 13, no. 1, p. 37–51.

Brahana, J.V., and Hollyday, E.F., 1988, Dry stream reaches in carbonate terranes—Surface indicators of ground-water reservoirs: American Water Resources Association Bulletin, v. 24, no. 3, p. 577–580.

Campbell, W.C., Lumsdon-West, M., and Davies, S., 2003, Geographic information system (GIS) support for karst hydrology, *in* Beck, B.F., ed., Sinkholes and the engineering and environmental impacts of karst, Proceedings of the Ninth Multidisciplinary Conference, September 6–10, 2003, Huntsville, Alabama: American Society of Civil Engineers, Geotechnical Special Publication no. 122, p. 429–438.

Cherkauer, D.S., 2004, Quantifying ground water recharge at multiple scales using PRMS and GIS: Ground Water, v. 42, no. 1, p. 97–110.

Clark, I., and Fritz, P., 1997, Environmental isotopes in hydrogeology: New York, Lewis Publishers, CRC Press LLC, 328 p.

Crawford, N.C., 1987, The karst hydrogeology of the Cumberland Plateau escarpment of Tennessee: Nashville, Tennessee Department of Conservation, Division of Geology, Report of Investigations no. 44, pt. 1, 43 p.

Crawford, N.C., Groves, C.G., Feeney, T.P., and Keller, B.J., 1987, Agriculture and urban nonpoint source pollution impacts on karst aquifers in the Pennyroyal karst region of Kentucky: Bowling Green, Western Kentucky University Center for Cave and Karst Studies (unpublished contract research report prepared for the Division of Water, Kentucky Natural Resources and Environmental Protection Cabinet, Frankfort, Kentucky and the Barren River Area Development District, Bowling Green, Kentucky), 229 p.

Culver, D.C., and White, W.B., 2004, Encyclopedia of caves: Academic Press, Elsevier, 680 p.

Currens, J.C., 1994, Characterization and quantification of nonpoint-source pollutant loads in a conduit-flow-dominated karst aquifer underlying an intensive-use agricultural region, Kentucky: Lexington, Proceedings of the Kentucky Water Resources Symposium, University of Kentucky Water Resources Research Institute, p. 3.

Currens, J.C., and Graham, C.D.R., 1993, Flooding of the Sinking Creek karst area in Jessamine and Woodford Counties, Kentucky: Kentucky Geological Survey Report of Investigations 7, series XI, 33 p.

Currens, J.C., and Ray, J.A., 1999, Karst atlas for Kentucky, *in* Beck, B.F., Pettit, A.J., and Herring, J.G., eds., Hydrogeology and engineering geology of sinkholes and karst, Proceedings of the Seventh Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, April 10–14, 1999, Harrisburg-Hershey, Pennsylvania: Rotterdam, Netherlands, A.A. Balkema, p. 85–90.

Currens, J.C., 2001, Generalized block diagram of the Western Pennyroyal Karst: Kentucky Geological Survey, map and chart 16, series XII.

Doctor, D.H., and Alexander, E.C., Jr., 2005, Interpretation of water chemistry and stable isotope data from a karst aquifer according to flow regimes identified through hydrograph recession analysis, *in* Kuniansky, E.L., ed., U.S. Geological Survey Karst Interest Group Proceedings, Rapid City, South Dakota, September 12–15, 2005: U.S. Geological Survey Scientific Investigations Report 2005–5160, p. 82–92.

Dogwiler, T., and Wicks, C.M., 2004, Sediment entrainment and transport in fluviokarst systems: Journal of Hydrology, v. 295, p. 163–172.

Dooge, J.C.I., 1973, Linear theory of hydrologic systems: U.S. Department of Agriculture Technical Bulletin 1468, 327 p.

Dreiss, S.J, 1982, Linear kernels for karst aquifers: Water Resources Research, v. 18, no. 4, p. 865–876.

Dreiss, S.J, 1983, Linear unit-response functions as indicators of recharge areas for karst springs: Journal of Hydrology, v. 61, p. 31–44.

Dreiss, S.J, 1989, Regional scale transport in a karst aquifer, 2—Linear systems and time moment analysis: Water Resources Research, v. 25, no. 1, p. 126–134.

Duley, J.W., 1986, Water tracing using a scanning spectrofluorometer for detection of fluorescent dyes, *in* Environmental Problems in Karst Terranes and their Solutions Conference, Bowling Green, Kentucky: Dublin, Ohio, Proceedings of the National Water Well Association, October 28–30, 1986, p. 389–406.

Felton, G.K., and Currens, J.C., 1994, Peak flow rate and recession-curve characteristics of a karst spring in the Inner Bluegrass, Central Kentucky: Journal of Hydrology, v. 162, p. 99–118.

Field, M.S., 1999a, Quantitative analysis of tracer breakthrough curves from tracing tests in karst aquifers, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 163–171.

Field, M.S., 1999b, On the importance of stock dye concentrations for accurate preparation of calibration standards, *in* Palmer, A.N., Palmer, M.V. and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute, Special Publication 5, p. 229.

Field, M.S., 2002a, A lexicon of cave and karst terminology with special reference to environmental karst hydrology: United States Environmental Protection Agency Publication EPA/600/R–02/003, 214 p.

Field, M.S., 2002b, The QTRACER2 program for tracer-breakthrough curve analysis for tracer tests in karstic aquifers and other hydrologic systems: U.S. Environmental Protection Agency Publication EPA/600/R–02/001, 179 p.

Field, M.S., 2003, Tracer-test planning using the Efficient Hydrologic Tracer-Test Design (EHTD) Program: Washington, D.C., U.S. Environmental Protection Agency, Office of Research and Development, National Center for Environmental Assessment, EPA/600/R–03/034, 103 p.

Field, M.S., 2006, Tracer-test design for losing stream-aquifer systems: International Journal of Speleology, v. 35, no. 1, p. 25–36.

Field, M.S., and Nash, S.G., 1997, Risk assessment methodology for karst aquifers—(1) Estimating karst conduit-flow parameters: Environmental Monitoring and Assessment, v. 47, no. 1, p. 1–27.

Field, M.S., Wilhelm, R.G., Quinlan, J.F., and Aley, T.J., 1995, An assessment of the potential adverse properties of fluorescent tracer dyes used for groundwater tracing: Environmental Monitoring and Assessment, v. 38, no. 1, p. 75–96.

Ford, D.C., 1999, Perspectives in karst hydrogeology and cavern genesis, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 17–29.

Ford, Derek, and Williams, P.W., 1989, Karst geomorphology and hydrology: London, Unwin Hyman, 601 p.

George, A.I., 1989, Caves and drainage north of the Green River *in* White, W.B., and White, E.L., eds., Karst hydrology concepts from the Mammoth Cave area: New York, Van Nostrand Reinhold, p. 189–222.

Ginsberg, M., and Palmer, A., 2002, Delineation of source-water protection areas in karst aquifers of the Ridge and Valley and Appalachian Plateaus Physiographic Provinces—Rules of thumb for estimating the capture zones of springs and wells: U.S. Environmental Protection Agency EPA 816–R–02–015, 41 p.

Greene, E.A., 1997, Tracing recharge from sinking streams over spatial dimensions of kilometers in a karst aquifer: Ground Water, v. 35, no. 5, p. 898–904.

Greene, E.A., Shapiro, A.M., and Carter, J.M., 1999, Hydrogeologic characterization of the Minnelusa and Madison aquifers near Spearfish, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 98–4156, 64 p.

Gunn, J., 1983, Point-recharge of limestone aquifers—A model from New Zealand karst: Journal of Hydrology, v. 61, p. 19–29.

Gunn, J., 1986, A conceptual model for conduit flow dominated karst aquifers, *in* Gunay, G., and Johnson, A.I., eds., Karst water resources, Proceedings of the International Symposium, Ankara, Turkey, July 1985: International Association of Hydrological Sciences Publication 161, p. 587–596.

Hauns, M., Jeannin, P.-Y., and Atteia, O., 2001, Dispersion, retardation and scale effect in tracer breakthrough curves in karst conduits: Journal of Hydrology, v. 241, p. 177–193.

Hess, J.W., and White, W.B., 1989, Water budget and physical hydrology, *in* White, W.B., and White, E.L., eds., Karst hydrology concepts from the Mammoth Cave area: New York, Van Nostrand Reinhold, p. 105–126.

Howcroft, W.D., 1992, Ground water flow and water resources investigation of the Auburn, Summers, and Shakertown Springs karst groundwater basins, Logan and Simpson Counties, Kentucky: Bowling Green, Western Kentucky University, Department of Geography, unpublished thesis, 128 p.

Imes, J.L., Schumacher, J.G., and Kleeschulte, M.J., 1996, Geohydrologic and water-quality assessment of the Fort Leonard Wood Military Reservation, Missouri, 1994–95: U.S. Geological Survey Water-Resources Investigations Report 96–4270, 134 p.

Jones, W.K., 1984a, Dye tracer tests in karst areas: National Speleological Society Bulletin, v. 46, p. 3–9.

Jones, W.K., 1984b, Analysis and interpretation of data from tracer tests in karst areas: National Speleological Society Bulletin, v. 46, p. 41–47.

Jones, W.K., 1997, Karst hydrology atlas of West Virginia: Leesburg, Va., Karst Waters Institute Special Publication 4, 111 p.

Jones, W.K., Culver, D.C., and Herman, J.S., eds., 2004, Epikarst, Proceedings of the 2003 Epikarst symposium, October 1–4, 2003, Shepherdstown, West Virginia, USA: Leesburg, Va., Karst Waters Institute Special Publication 9, 160 p.

Kasnavia, T., Vu, D., and Sabatini, D.A., 1999, Fluorescent dye and media properties affecting sorption and tracer selection: Ground Water, v. 37, no. 3, p. 376–381.

Käss, W., 1998, Tracing technique in geohydrology: Rotterdam, Netherlands, A.A. Balkema; Brookfield, Vermont, A.A. Balkema Publishers, 581 p.

Katz, B.G., 2005, Demystification of ground-water flow and contaminant movement in karst systems using chemical and isotopic tracers, *in* Kuniansky, E.L., ed., U.S. Geological Survey Karst Interest Group Proceedings, Shepherdstown, West Virginia August 20–22, 2002: U.S. Geological Survey Water-Resources Investigations Report 2002–4174, p. 13–19.

Katz, B.G., Coplen, T.B., Bullen, T.D., and Davis, J.H., 1997, Use of chemical and isotopic tracers to characterize the interactions between ground water and surface water in mantled karst: Ground Water, v. 35, no. 6, p. 1014–1028.

Kilpatrick, F.A., and Cobb, E.D., 1984, Measurement of discharge using tracers: U.S. Geological Survey Open-File Report 84–136, 73 p.

Kincaid, T.R., Hazlett, T.J., and Davies, G.J., 2005, Quantitative groundwater tracing and effective numerical modeling in karst—An example from the Woodville Karst Plain of North Florida, *in* Beck, B.F., ed., Sinkholes and the engineering and environmental impacts of karst, Proceedings of the 10th Multidisciplinary Conference, San Antonio, Texas, September 24–28, 2005: Geological Institute of the American Society of Civil Engineers, Geotechnical Special Publication no. 144, p. 114–121.

Klimchouk, A.B., 2004, Towards defining, delimiting and classifying epikarst—Its origin, processes and variants of geomorphic evolution, *in* Jones, W.K., Culver, D.C., and Herman, J., eds., Epikarst, Proceedings of the 2003 Epikarst symposium, October 1–4, 2003, Shepherdstown, West Virginia, USA: Karst Waters Institute Special Publication 9, p. 23–35.

Klimchouk, A.B., Ford, D.C., Palmer, A.N., and Dreybrodt, W., 2000, Speleogenesis evolution of karst aquifers: National Speleological Society, 527 p.

Kresic, N., 1997, Quantitative solutions in hydrogeology and groundwater modeling: Boca Raton, CRC Lewis Publishers, 461 p.

Kuniansky, E.L., ed., 2001, U.S.Geological Survey Karst Interest Group Proceedings, St. Petersburg, Florida, February 13–16, 2001: U.S. Geological Survey Water-Resources Investigations Report 2001–4011, 211 p.

Kuniansky, E.L., ed., 2002, U.S.Geological Survey Karst Interest Group Proceedings, Shepherdstown, West Virginia, August 20–22, 2002: U.S. Geological Survey Water-Resources Investigations Report 2002–4174, 89 p.

Kuniansky, E.L. ed., 2005, U.S.Geological Survey Karst Interest Group Proceedings, Rapid City, South Dakota, September 12–15, 2005: U.S. Geological Survey Scientific Investigations Report 2005–5160, 296 p.

Kuniansky, E.L., Fahlquist, L., and Ardis, A.F., 2001, Travel times along selected flow paths of the Edwards aquifer, Central Texas, *in* Kuniansky, E.L., ed., U.S. Geological Survey Karst Interest Group Proceedings, St. Petersburg, Florida, February 13–16, 2001: U.S. Geological Survey Water-Resources Investigations Report 2001–4011, p. 69–77.

Lakey, B.L., and Krothe, N.C., 1996, Stable isotope variation of storm discharge from a perennial karst spring, Indiana: Water Resources Research, v. 32, no. 2, p. 721–731.

Lane, S.R., and Smart, C.C., 1999, A method for correction of variable background fluorescence in filter fluorometry, *in* Beck, B.F., Pettit, A.J., and Herring, J.G., eds., Hydrology and engineering geology of sinkholes and karst—1999, Proceedings of the Seventh Multidisciplinary Conference, Harrisburg-Hershey Pennsylvania, April 10–14, 1999: Rotterdam, Netherlands, A.A Balkema, p. 301–306.

Leavesley, G.H., Lichty, R.W., Troutman, B.M., and Saindon, L.G., 1983, Precipitation-Runoff Modeling System—User's manual: U.S. Geological Survey Water-Resources Investigations Report 83–4238, 207 p.

Lee, E.S., and Krothe, N.C., 2001, A four-component mixing model for water in a karst terrain in south-central Indiana, USA, using solute concentration and stable isotopes as tracers: Chemical Geology, v. 179, p. 129–143.

Liedl, R., Sauter, M., Hückinghaus, D., Clemens, T., and Teutsch, G., 2003, Simulation of the development of karst aquifers using a coupled continuum pipe flow model: Water Resources Research, v. 39, no. 3, p. 1057.

Long, A.J., and Derickson, R.G., 1999, Linear systems analysis in a karst aquifer: Journal of Hydrology, v. 219, p. 206–217.

Meiman, J., and Ryan, M.T., 1999, The development of basin-scale conceptual models of the active-flow conduit system, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 203–212.

Meinzer, O.E., 1927, Large springs in the United States: U.S. Geological Survey Water-Supply Paper 557, 94 p.

Milanović, P.T., 1981a, Water regime in deep karst—Case study of the Ombla Spring drainage area, *in* Yevjevich, V., ed., Karst hydrology and water resources, Proceedings of the U.S.-Yugoslavian Symposium, Dubrovnik, 1975: Littleton, Colo., Water Resources Publication, p. 165–191.

Milanović, P.T., 1981b, Karst hydrogeology: Littleton, Colo., Water Resources Publication, 434 p.

Mull, D.S., Smoot, J.L., and Liebermann, T.D., 1987, Dye tracing techniques used to determine ground-water flow in a carbonate aquifer system near Elizabethtown, Kentucky: U.S. Geological Survey Water-Resources Investigations Report 87–4174, 95 p., 2 pls.

Mull, D.S., Liebermann, T.D., Smoot, J.L., and Woosley, Jr., L.H., 1988, Application of dye-tracing techniques for determining solute-transport characteristics of ground water in karst terranes: Atlanta, Ga., U.S. Environmental Protection Agency, EPA Report 904/6–88–001, 103 p.

National Academy of Sciences, 2004, Groundwater fluxes across interfaces—Committee on hydrologic science: Washington, D.C., National Research Council, National Academies Press, 100 p.

National Ground Water Association, 1991, Proceedings of the Third Conference on Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes, December 4–6, 1991, Nashville, Tennessee: Dublin, Ohio, National Ground Water Association, 793 p.

National Water Well Association, 1986, Proceedings of the Environmental Problems in Karst Terranes and Their Solutions Conference, October 28–30, 1986, Bowling Green, Kentucky: Dublin, Ohio, National Water Well Association, 525 p.

National Water Well Association, 1988, The Proceedings of the Second Conference on Environmental Problems in Karst Terranes and Their Solutions, November 16–18, 1988, Nashville, Tennessee: Dublin, Ohio, National Water Well Association, 441 p.

Neuman, S.P., and de Marsily, G., 1976, Identification of linear systems response by parametric programming: Water Resources Research, v. 12, no. 2, p. 253–262.

Padilla, A., Pulio-Bosch, A., and Mangin, A., 1994, Relative importance of baseflow and quickflow from hydrographs of karst springs: Ground Water, v. 32, p. 267–277.

Paillet, F.L., 2001, Borehole geophysical applications in karst hydrology, *in* Kuniansky, E.L., ed., U.S. Geological Survey Karst Interest Group Proceedings, St. Petersburg, Florida, February 13–16, 2001: U.S. Geological Survey Water-Resources Investigations Report 2001–4011, p. 116–123.

Palmer, A.N., 1991, Origin and morphology of limestone caves: Geological Society of America Bulletin, v. 103, p. 1–21.

Palmer, A.N., 1999, Patterns of dissolution porosity in carbonate rocks, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 71–78.

Pinault, J.L., Plagnes, V., Aquilina, L., and Bakalowicz, M., 2001, Inverse modeling of the hydrological and hydrochemical behavior of hydrosystems—Characterization of karst system functioning: Water Resources Research, v. 37, no. 8, p. 2191–2204.

Quinlan, J.F., 1970, Central Kentucky karst: Méditerranée Études et Travaux, v. 7, p. 235–253.

Quinlan, J.F., 1989, Ground-water monitoring in karst terranes—Recommended protocols and implicit assumptions: Las Vegas, Nev., U.S. Environmental Protection Agency, Environmental Monitoring Systems Laboratory, EPA/600/X–89/050, 79 p.

Quinlan, J.F., and Ewers, R.O., 1989, Subsurface drainage in the Mammoth Cave area, *in* White, W.B., and White, E.L., eds., Karst hydrology concepts from the Mammoth Cave area: New York, Van Nostrand Reinhold, p. 65–104.

Quinlan, J.F., Ewers, R.O., Ray, J.A., Powell, R.L., and Krothe, N.C., 1983, Groundwater hydrology and geomorphology of the Mammoth Cave region, Kentucky, and of the Mitchell Plain, Indiana, *in* Shaver, R.H., and Sunderman, V.A., eds., Field trips in midwestern geology: Geological Society of America, v. 2, p. 1–85.

Quinlan, J.F., and Ray, J.A., 1989, Groundwater basins in the Mammoth Cave region, Kentucky, showing springs, major caves, flow routes, and potentiometric surface: Mammoth Cave, Ky., Friends of the Karst Occasional Publication no. 2, 1 pl.

Quinlan, J.F., and Ray J.A., 1995, Normalized base-flow discharge of ground-water basins—A useful parameter for estimating recharge area of springs and for recognizing drainage anomalies in karst terranes, *in* Beck, B., ed., Karst geohazards, Proceedings of the Fifth Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, April 2–5, 1995, Gatlinburg, Tennessee: Rotterdam, Netherlands, A.A. Balkema, p. 149–164.

Quinlan, J.F., Ray, J.A., and Schindel, G.M., 1995, Intrinsic limitations of standard criteria and methods for delineation of groundwater-source protection areas (springhead and wellhead protection areas) in carbonate terranes—Critical review, technically-sound resolution of limitations, and case study in a Kentucky karst, *in* Beck, B.F., and Pearson, F.M., eds., Karst geohazards engineering and environmental problems in karst terrane: A.A. Balkema, Rotterdam, Netherlands, p. 165–176.

Ray, J.A., 1994, Surface and subsurface trunk flow, Mammoth Cave, Kentucky: Mammoth Cave, Ky., Proceedings of the Third Mammoth Cave National Park Science Conference, Mammoth Cave National Park, p. 175–187.

Ray, J., 1999, A model of karst drainage basin evolution, interior low plateaus, USA, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 58.

Ray, J., 2001, Spatial interpretation of karst drainage basins: *in* Beck, B.F., and Herring, J.G., eds., Geotechnical and environmental applications of karst geology and hydrology, Proceedings of the Eighth Multidisciplinary Conference on Sinkholes and the Environmental and Engineering Impacts of Karst, April 1–4, 2001, Louisville, Kentucky: Lisse, Balkema Publishers, p. 235–244.

Ray, J.A., and Currens, J.C., 1998, Mapped karst ground-water basins in the Beaver Dam 30 × 60 minute quadrangle: Kentucky Geological Survey Map and Chart, ser. 11, map and chart 19, scale 1:100,000.

Rendell, D., 1987, Fluorescence and phosphorescence—Analytical chemistry by open learning series: London, Thames Polytechnic, John Wiley and Sons, 419 p.

Rorabaugh, M.I., 1964, Estimating changes in bank storage as ground-water contribution to streamflow: International Association Scientific Hydrological Publications 63, p. 432–441.

Rovey, C.W., II, and Cherkauer, D.S., 1994, Relation between hydraulic conductivity and texture in a carbonate aquifer: regional continuity: Groundwater v. 32, no. 2, p. 227–238.

Ryan, M.T., and Meiman, J., 1996, An examination of short-term variations in water quality at a karst spring in Kentucky: Ground Water, v. 34, no. 1, p. 23–30.

Sauter, M., 1991, Assessment of hydraulic conductivity in a karst aquifer at local and regional scale, *in* Proceedings of the Third Conference on Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes December 4–6, 1991, Nashville, Tennessee: National Ground Water Association, p. 39–56.

Sauter, M., 1992, Assessment of hydraulic conductivity in a karst aquifer at local and regional scale: Ground Water Management, v. 10, p. 39–57.

Scanlon, B.R., Mace, R.E., Barrett, M.E, and Smith, B., 2003, Can we simulate regional groundwater flow using equivalent porous media models? Case study, Barton Springs Edwards aquifer, USA: Journal of Hydrology, v. 276, p. 137–158.

Scanlon, B.R., and Thrailkill, 1987, Chemical similarities among physically distinct spring types in a karst terrain: Journal of Hydrology, v. 89, p. 259–279.

Schindel, G.M., Ray, J.A., and Quinlan, J.F., 1995, Delineation of the recharge area for Rio Springs, Kentucky—An EPA demonstration project in wellhead (springhead) protection for karst terranes, *in* Beck, B.F., and Pearson, F.M., eds., Karst geohazards engineering and environmental problems in karst terrane: Rotterdam, Netherlands, and Brookfield, Vermont, A.A. Balkema, p. 165–176.

Shapiro, A.M., 2001, Effective matrix diffusion in kilometer-scale transport in fractured crystalline rock: Water Resources Research, v. 37, p. 507–522.

Shevenell, L., 1996, Analysis of well hydrographs in a karst aquifer—Estimates of specific yields and continuum transmissivities: Journal of Hydrology, v. 174, p. 331–355.

Shuster, E.T., and White, W.B., 1971, Seasonal fluctuations in the chemistry of limestone springs—A possible means of characterizing carbonate aquifers: Journal of Hydrology, v. 14, p. 93–128.

Smart, C.C., 1988, A deductive model of karst evolution based on hydrological probability: Earth Surface Processes and Landforms, v. 13, p. 271–288.

Smart, C.C., 1998, Artificial tracer techniques for the determination of the structure of conduit aquifers: Ground Water, v. 26, p. 445–453.

Smart, C.C., 2005, Error and technique in fluorescent dye tracing, *in* Beck, B.F., ed., Sinkholes and the engineering and environmental impacts of karst: Reston, Va., American Society of Civil Engineers, Geotechnical Special Publication no. 144, p. 500–509.

Smart, C.C., and Ford, D.C., 1982, Quantitative dye tracing in a glacierized alpine karst: Beitraege Zur Geologie der Schweiz—Hydrologie, v. 28, no. 1, p. 191–200.

Smart, C.C., and Karunaratne, K.C., 2002, Characterization of fluorescence background in dye tracing: Environmental Geology, v. 42, p. 492–498.

Smart, C.C., and Simpson, B., 2001, An evaluation of the performance of activated charcoal in detection of fluorescent compounds in the environment, *in* Beck and Herring, eds., Geotechnical and environmental applications of karst geology and hydrology: Rotterdam, Netherlands, A.A Balkema, p. 265–270.

Smart, C.C., and Simpson, B., 2002, Detection of fluorescent compounds in the environment using granular activated charcoal detectors: Environmental Geology, v. 42, p. 538–545.

Smart, C.C., and Smart, P.L., 1991, Correction of background interference and cross-fluorescence in filter fluorometric analysis of water-tracer dyes, *in* Proceedings of the Third Conference on Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes, December 4–6, 1991, Nashville, Tennessee: National Ground Water Association, p. 475–491.

Smart, C.C., Zabo, L., Alexander, E.C., Jr., and Worthington, S.R.H., 1998, Some advances in fluorometric techniques for water tracing: Environmental Monitoring and Assessment, v. 53, p. 305–320.

Smart, P.L., 1972, A laboratory evaluation of the use of activated carbon for the detection of tracer dye Rhodamine WT: University of Alberta, Canada, (unpublished Master's thesis).

Smart, P.L., and Laidlaw, I.M.S., 1977, An evaluation of some fluorescent dyes for water tracing: Water Resources Research, v. 13, p. 15–33.

Smoot, J.L., Mull, D.S., and Liebermann, T.D., 1987, Quantitative dye tracing techniques for describing the solute transport characteristics of ground-water flow in karst terrane, *in* Proceedings of the Second Multidisciplinary Conference on Sinkholes and the Environmental Impacts of Karst, Orlando, Florida, February 9–11, 1987: National Water Well Association, p. 269–275.

Steinheimer, T.R., and Johnson, S.M., 1986, Investigation of the possible formation of diethylnitrosamine resulting from the use of rhodamine WT as a tracer in river waters: U.S. Geological Survey Water-Supply Paper 2290, p. 37–49.

Stoica, P., and Soderstrom, T., 1982, Instrumental variable methods for identification of Hammerstein system: International Journal of Control, v. 35, p. 459–476.

Streltsova, T.D., 1988, Well testing in heterogeneous formations: New York, John Wiley and Sons, 413 p.

Sukhodolov, A.N., Nikora, V.I., Rowiński, P.M., and Czernuszenko, W., 1997, A case study of longitudinal dispersion in small lowland rivers: Water Environmental Resources, v. 97, p. 1246–1253.

Taylor, C.J., and McCombs, G.K., 1998, Recharge-area delineation and hydrology, McCraken Springs, Fort Knox Military Reservation, Meade County, Kentucky: U.S. Geological Survey Water-Resources Investigations Report 98–4196, 12 p., 1 pl.

Taylor, C.J., Nelson, H.L., Hileman, G., and Kaiser, W.P., 2005, Hydrogeologic framework mapping of shallow, conduit-dominated karst—Components of a regional GIS-based approach, *in* Kuniansky, E.L., ed., U.S. Geological Survey Karst Interest Group Proceedings, Rapid City, South Dakota, September 12–15, 2005: U.S. Geological Survey Scientific Investigations Report 2005–5160, p. 103–113.

Teutsch, G., 1992, Groundwater modeling in karst terranes—Scale effects, data acquisition, and field validation: Ground Water Management, v. 10, p. 17–35.

Teutsch G., and Sauter, M., 1991, Groundwater modeling in karst terranes—Scale effects, data acquisition, and field verification, Proceedings of the Third Conference on Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes, Nashville, Tennessee, December 4–6, 1991: National Ground Water Association, p. 17–54.

Thrailkill, J., 1985, The Inner Bluegrass karst region, *in* Dougherty, P.H., ed., Caves and karst of Kentucky: Kentucky Geological Survey Special Publication 12, Series IX, p. 28–62.

Toth, J., 1963, A theoretical analysis of groundwater flow in small drainage basins: Journal of Geophysical Research, v. 68, p. 211–222.

Trćek, B., and Krothe, N.C., 2002, The importance of three and four-component storm hydrograph separation techniques for karst aquifers, *in* Gabrovšek, Franci, ed., Evolution of karst—From prekarst to cessation: Postojna-Ljubljana Založba ZRC, p. 395–401.

Tucker, R.B., and Crawford, N.C., 1999, Non-linear curve-fitting analysis as a tool for identifying and quantifying multiple fluorescent tracer dyes present in samples analyzed with a spectrofluorophotometer and collected as part of a dye tracer study of groundwater flow, *in* Beck, B.F., Pettit, A.J., and Herring, J.G., eds., Hydrology and engineering geology of sinkholes and karst—1999: Harrisburg-Hershey, Pennsylvania, Proceedings of the Seventh Multidisciplinary Conference, April 10–14, 1999, p. 307–312.

U.S. Geological Survey, 2005, User's Manual for the National Water Information System of the U.S. Geological Survey, Ground-Water Site-Inventory System: U.S. Geological Survey Open-File Report, version 4.5, available at <*http://wwwnwis.er.usgs.gov/nwisdocs4_5/gw/gw_complete.pdf*>

Vandike, J.E., 1992, The hydrogeology of the Bennett Spring area, Laclede, Dallas, Webster, and Wright Counties, Missouri: Missouri Department of Natural Resources Division of Geology and Land Survey Water Resources Report no. 38, 112 p.

White, W.B. 1988, Geomorphology and hydrology of karst terrains: New York, Oxford University Press, 464 p.

White, W.B., 1993, Analysis of karst aquifers, *in* Alley, W.M., ed., Regional ground-water quality: New York, Van Nostrand Reinhold, p. 471–489.

White, W.B., 1999, Conceptual models for karstic aquifers, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 11–16.

White, W.B., and Schmidt, V.A., 1966, Hydrology of a karst area in east-central West Virginia: Water Resources Research, v. 2, no. 3, p. 549–560.

White, W.B., and White, E.L., eds., 1989, Karst hydrology—Concepts from the Mammoth Cave area: New York, Van Nostrand Reinhold, 346 p.

Wicks, C.M., and Hoke, J.A., 2000, Prediction of the quantity and quality of Maramec Spring water: Ground Water, v. 38, no. 2, p. 218–225.

Williams, P.W, 1983, The role of the subcutaneous zone in karst hydrology: Journal of Hydrology, v. 61, p. 45–67.

Wilson, J.F., Jr., Cobb, E.D., and Kilpatrick, F.A., 1986, Fluorometeric procedures for dye tracing: U.S. Geological Survey Techniques for Water Resources Investigations, book 3, chap. A12, 34 p.

Worthington, S.R.H., 1991, Karst hydrology of the Canadian Rocky Mountains: Hamilton, Ontario, Canada, McMaster University, unpublished Ph. D. dissertation, 380 p.

Worthington, S.R.H., 1999, A comprehensive strategy for understanding flow in carbonate aquifers, *in* Palmer, A.N., Palmer, M.V., and Sasowsky, I.D., eds., Karst modeling: Leesburg, Va., Karst Waters Institute Special Publication 5, p. 30–37.

Worthington, S.R.H., 2003, Numerical simulation of the aquifer at Mammoth Cave, *in* Kambesis, P., Meiman, J., and Groves, C., eds., Proceedings of the International Conference on Karst Hydrogeology and Ecosystems, Bowling Green, Kentucky, USA, June 3–6, 2003: Western Kentucky University, Hoffman Environmental Research Institute, p. 10.

Worthington, S.R.H., Davies, G.J., and Ford, D.C., 2000, Matrix, fracture and channel components of storage and flow in a Paleozoic limestone aquifer, *in* Sasowsky, I.D., and Wicks, C.M., eds., Groundwater flow and contaminant transport in carbonate aquifers: Rotterdam, Netherlands, A.A. Balkema, p. 113–128.

Worthington, S.R.H., and Smart, C.C., 2003, Empirical determination of tracer mass for sink to spring tests in karst, *in* Beck, B., ed., Sinkholes and the engineering and environmental impacts of karst: Huntsville, Ala., Proceedings of the Ninth Multidisciplinary Conference, September 6–10, 2003, p. 287–295.

Zhang, Y.K., and Bai, E.W., 1996, Simulation of spring discharge from a limestone aquifer in northeastern Iowa with an identification scheme, *in* Kovar, Karl, and van der Heijde, Paul, eds., Calibration and reliability in groundwater modeling: Golden, Colo., Proceedings of ModelCARE96 Conference, September 1996, IAHS Publication no. 237, p. 31–40.

Zhang, Y.K., Bai, E.W., Rowden, R., and Liu, H., 1995, Simulation of spring discharge from a limestone aquifer in Iowa, USA: Hydrogeology Journal, v. 4, no. 4, p. 41–54.