Introduction and Characteristics of Flow

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Chapter 1 of
Field Techniques for Estimating Water Fluxes Between Surface Water and Ground Water

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Introduction

Interest in the use and development of our Nation's surface- and ground-water resources has increased significantly during the past 50 years (Alley and others, 1999; Hutson and others, 2004). At the same time, a variety of techniques and methods have been developed to examine and monitor these water resources. Quantifying the connection between surface water and ground water also has become more important because the use of one of these resources can have unintended consequences on the other (Committee on Hydrologic Science, National Research Council, 2004). In an attempt to convey the importance of the linkages and interfaces between surface water and ground water, the two have been described as a "single resource" (Winter and others, 1998). An improved understanding of the connection between surface and ground waters increasingly is viewed as a prerequisite to effectively managing these resources (Sophocleous, 2002). Thus, water-resource managers have begun to incorporate management strategies that require quantifying flow between surface water and ground water (Danskin, 1998; Bouwer and Maddock, 1997; Dokulil and others, 2000; Owen-Joyce and others, 2000; Barlow and Dickerman, 2001; Jacobs and Holway, 2004).

The use of surface water (or ground water) can change the location, rate, and direction of flow between surface water and ground water (Stromberg and others, 1996; Glennon, 2002; Galloway and others, 2003). Pumping wells in the vicinity of rivers commonly cause river water to flow into the underlying ground-water body, which can affect the quality of the ground water (Childress and others, 1991; McCarthy and others, 1992; Lindgren and Landon, 1999; Steele and Verstraeten, 1999; Zarriello and Reis, 2000; Sheets and others, 2002). In some cases, ground water is pumped to provide water for cooling industrial equipment and then discharged into lakes, ponds, or rivers (Andrews and Anderson, 1978; Hutson and others, 2004). Ground water also may be pumped specifically to maintain lake levels for recreation purposes, especially during droughts (Stewart and Hughes, 1974; Mcleod, 1980; Belanger and Kirkner, 1994; Metz and Sacks, 2002). Surface water can be directed into surface basins where water percolates to the underlying aquifer—a process known as artificial recharge (Galloway and others, 2003). Ground-water discharge areas, where ground water flows into surface water, can be important habitats for fish (Garrett and others, 1998; Power and others, 1999; Malcolm and others, 2003a, 2003b). Water in irrigation canals can flow or seep to an underlying aquifer, which eventually discharges water to rivers, thereby sustaining streamflow essential for the maintenance of fish populations (Konrad and others, 2003).

Interest in the interaction of surface water and ground water is not confined to inland waters. This interaction has been studied in coastal areas because fresh ground-water supplies can be affected by intrusion of saltwater (Barlow and Wild, 2002). Beyond the issue of water supply for human consumption, increased attention has been given to the ground water that discharges to oceans and estuaries, both in terms of water quantity and quality (Bokuniewicz, 1980; Moore, 1996, 1999; Linderfelt and Turner, 2001). Discharge of fresh ground water to oceans and estuaries, also referred to as submarine ground-water discharge, is important in maintaining the flora and fauna that have evolved to exploit this source of fresh water in a saline environment (Johannes, 1980; Simmons, 1992; Corbett and others, 1999). Nitrate in submarine ground-water discharge to estuaries and coastal waters can result in eutrophication of those waters (Johannes, 1980; Johannes and Hearn, 1985; Valiela and others, 1990; Taniguchi and others, 2002). Withdrawals or pumping of ground water at near-shore, inland locations can reduce the submarine discharge of ground water offshore and change the environmental conditions of these settings (Simmons, 1992). Some coral reefs may be endangered by diminished submarine ground-water discharge (Bacchus, 2001, 2002).

The variety of settings of interest for the examination of the interaction between surface water and ground water makes evident the need for methods to describe and quantify that flow. The exact method chosen for each setting will vary depending on the physical and hydrological conditions present in those settings, as well as the scale of the interaction. Some degree of measurement uncertainty accompanies each method or technique. Thus, it is prudent to consider using more than one method to examine the interaction between surface water and ground water. Because numerous techniques and methods are available to describe and quantify the flow between surface water and ground water, it is useful to provide water-resource investigators an overview of available techniques and methods, as well as their application.

Purpose and Scope

Several methods have been developed and applied to the study of the exchange between surface water and ground water (fig. 1). Different methods are better suited for characterizing or measuring flow over large or small areas. If an initial view of a considerable area or distance is needed to determine where measurable ground-water discharge is occurring, aerial infrared photography or imagery can be effective reconnaissance tools. On a smaller scale, some methods may involve direct measurement of sediment temperature or specific conductance along transects within a surface-water body, or use of dyes or other tracers to indicate the direction and rate of water movement. The measurement of water levels in well networks in the watershed can be used to determine ground-water gradients relative to adjacent surface water, which in turn can indicate the direction and rate of flow between the surface-water body and the underlying aquifer. In streams and rivers, measurement of flow at the endpoints of a channel reach can reveal if the reach is gaining flow from ground water or losing flow to ground water. Addition of tracers to streams also can be used to determine surface-water interaction with ground water over a range of scales. Local interaction of surface water with ground
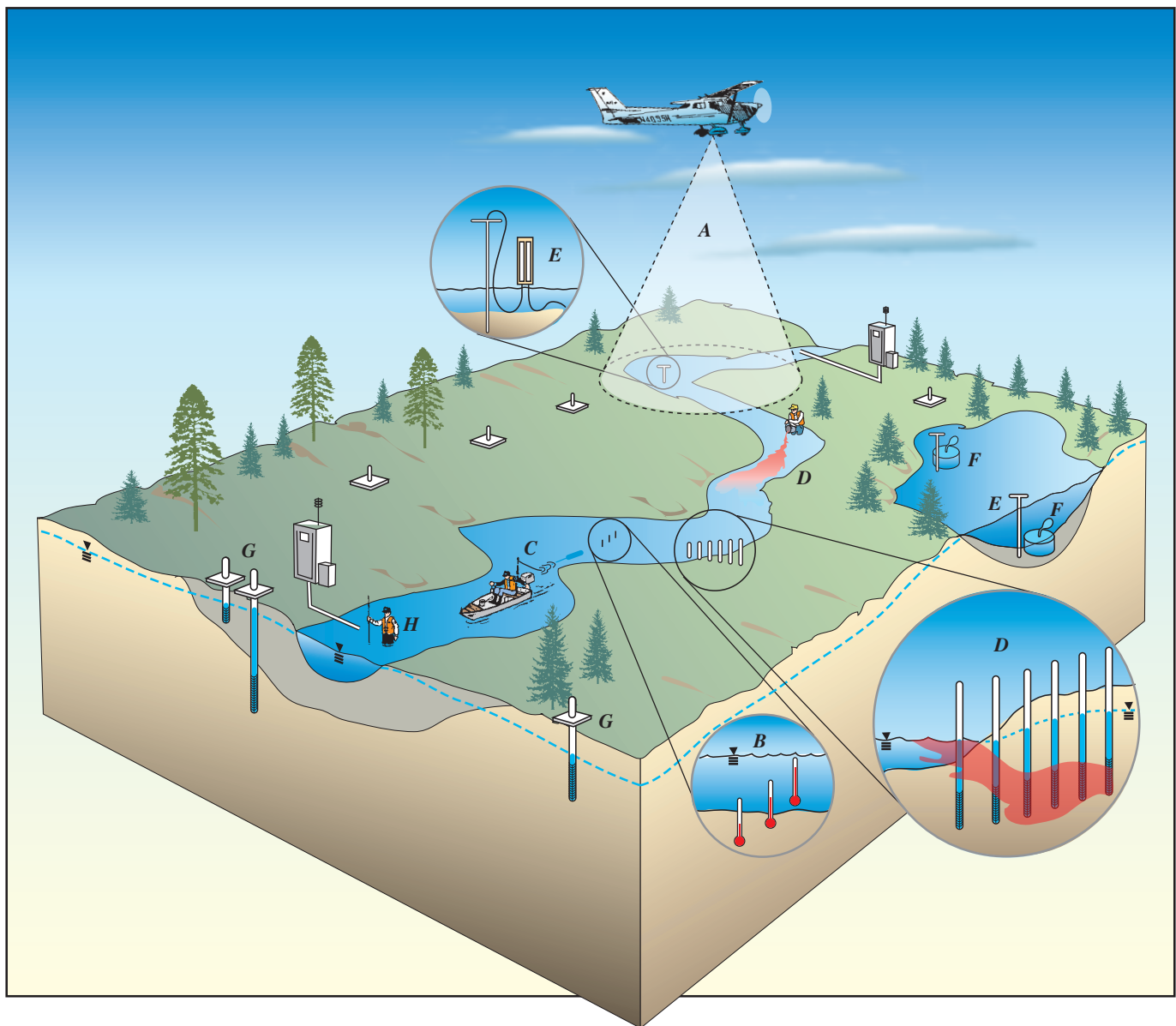
Figure 1.   Summary of techniques that have been used for the measurement or estimation of water fluxes between surface water and ground water. Techniques illustrated include: (*A*) aerial infrared photography and imagery, (*B*) thermal profiling, (*C*) the use of temperature and specific-conductance probes, (*D*) dyes and tracers, (*E*) hydraulic potentiomanometers, (*F*) seepage meters, (*G*) well networks, and (*H*) streamflow measurements. (Artwork by John M. Evans, U.S. Geological Survey, retired.)

water is measured by placing devices such as thermistors, minipiezometers, and seepage meters in the sediment, to monitor temperature gradients, hydraulic gradients, or quantity of flow. Determination of how interaction of surface water with ground water changes over time is made possible by using data-recording devices ("data loggers") in conjunction with pressure transducers, thermistors, and water-quality probes.

This report is designed to make the reader aware of the breadth of approaches (fig. 1) available for the study of the exchange between surface and ground water. To accomplish this, the report is divided into four chapters. Chapter 1 describes many well-documented approaches for defining the flow between surface and ground waters. Subsequent chapters provide an in-depth presentation of particular methods. Chapter 2 focuses on three of the most commonly used methods to either calculate or directly measure flow of water between surface-water bodies and the ground-water domain: (1) measurement of water levels in well networks in combination with measurement of water level in nearby surface water to determine water-level gradients and flow; (2) use of portable piezometers (wells) or hydraulic potentiomanometers to measure hydraulic gradients; and (3) use of seepage meters to measure flow directly. Chapter 3 focuses on describing the techniques involved in conducting water-tracer tests using fluorescent dyes, a method commonly used in the hydrogeologic investigation and characterization of karst aquifers, and in the study of water fluxes in karst terranes. Chapter 4 focuses on heat as a tracer in hydrological investigations of the near-surface environment.

This report focuses on measuring the flow of water across the interface between surface water and ground water, rather than the hydrogeological or geochemical processes that occur at or near this interface. The methods, however, that use hydrogeological and geochemical evidence to quantify water fluxes are described herein. This material is presented as a guide for those who have to examine the interaction of surface water and ground water. The intent here is that both the overview of the many available methods and the in-depth presentation of specific methods will enable the reader to choose those study approaches that will best meet the requirements of the environments and processes they are investigating, as well as to recognize the merits of using more than one approach. To that end, at this point it is useful to examine the content of each chapter in more detail.

Chapter 1 provides an overview of typical settings in the landscape where interactions between surface water and ground water occur. The chapter reviews the literature, particularly recent publications, and describes many well-documented methods for defining the flow between surface and ground waters. A brief overview of the theory behind each method is provided. Information is presented about the field settings where the method has been applied successfully, and, where possible, generalizes the requirements of the physical setting necessary to the success of the method. Strengths and weaknesses of each method are noted, as appropriate. This will aid the investigator in choosing methods to apply to their

setting. For those already familiar with some of these methods, the review of recent literature provides information about improvements in these methods.

Chapter 2 describes three of the most commonly used methods to either calculate or directly measure flow of water between surface-water bodies and the ground-water domain. The first method involves measurement of water levels in a network of wells in combination with measurement of the stage of the surface-water body to calculate gradients and then water flow. The second method involves the use of portable piezometers (wells) or hydraulic potentiomanometers to measure gradients. In the third method, seepage meters are used to directly measure flow across the sediment-water interface at the bottom of the surface-water body. Factors that affect measurement scale, accuracy, sources of error in using each of the methods, common problems and mistakes in applying the methods, and conditions under which each method is well- or ill-suited also are described.

Chapter 3 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 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.

Chapter 4 reviews early work addressing heat as a tracer in hydrological investigations of the near-surface environment, describes recent advances in the field, and presents selected new results designed to identify the broad application of heat as a tracer to investigate surface-water/ground-water exchanges. An overview of field techniques for estimating water fluxes between surface water and ground water with heat is provided.

To familiarize readers with flow conditions that may occur during their studies, the next section of Chapter 1 describes commonly observed interactions between surface water and ground water.

Characteristics of Water Exchange Between Surface Water and Ground Water

Most measurements made for the purpose of quantifying exchange between surface water and ground water are obtained at points within a short distance of the shoreline of the surface-water body. Shorelines represent the horizontal interface between ground water and surface water, an interface that is highly dynamic spatially and temporally. Because of the complex physical processes that occur in precisely the area where measurements are needed, it is important to understand those processes at shorelines and the range of potential changes in conditions at shorelines that occur over time. The following section elaborates these points.

Typically, a significant break in slope in the water table occurs where the horizontal surface of a lake, stream, or wetland intersects the sloping surface of the ground-water table (fig. 2). Because of this break in slope, ground-water flow lines diverge where they extend beneath and end at the sediment-water interface. Diverging flow lines indicate that the rate of flow per unit area is decreasing. Given homogeneous and isotropic conditions in the porous media adjacent to and beneath the sediment-water interface, seepage across the interface will decrease exponentially with distance from shore (fig. 3) (McBride and Pfannkuch, 1975; Pfannkuch and Winter, 1984). The movement of water between surface water and ground water can occur in a variety of settings or landscapes (fig. 4), each of which can be related to the break in slope of the water table defined by "an upland adjacent to a lowland separated by an intervening steeper slope" (Winter, 2001).

Ground-water flow lines bend substantially beneath the sediment-water interface just before they intersect the surface-water body. Measurements of hydraulic-head gradients typically assume that the flow lines either are horizontal (in the case of comparing heads in near-shore wells with surface-water stage) or vertical (in the case of inserting the screened intervals of wells to some depth beneath the sediment-water interface). In reality, the orientation of the flow lines are somewhere between horizontal and vertical as shown in figure 5.

Characteristics of Near-Shore Sediments

Although some investigators have found that seepage decreases exponentially with distance from shore (Lee, 1977; Fellows and Brezonik, 1980; Erickson, 1981; Attanayake and Waller, 1988; Rosenberry, 1990), other studies report that the decrease in flow across the sediment-water interface is not
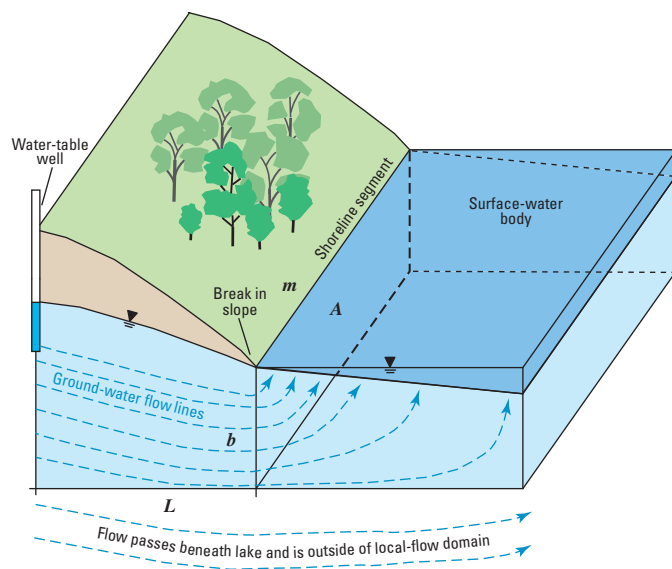
Figure 3. Decrease in seepage discharge with distance from shore (from Winter and others, 1998).
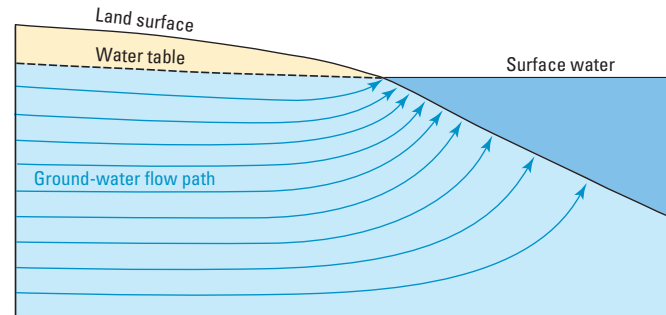
exponential because of heterogeneity of the sediment. One of the early fndings of a departure from what would be expected in a homogeneous, isotropic setting was reported by Woessner and Sullivan (1984) in their study of Lake Mead, Nevada. At many of the transects across which they collected data in Lake Mead, they found seepage did not decrease exponentially, and furthermore, that seepage sometimes decreased and then increased with distance from shore. They reported a large variability in seepage with distance from shore. This variability was attributed to heterogeneity in the sediments in the vicinity of the sediment-water interface. Krabbenhoft and Anderson (1986) also reported that seepage was focused in a gravel lens that intersected the lakebed some distance from shore at Trout Lake, Wisconsin. It now generally is recognized that aquifers adjacent to and beneath surface-water bodies rarely can be considered homogeneous, and usually are not isotropic.

Many processes act to create heterogeneity at the sediment-water interface. A few are listed below.

1. *Fluvial processes*—Depositional and erosional processes occur nearly constantly in streambeds and riverbeds, making heterogeneity a significant feature in these sediment-water interfaces. Organic deposits commonly are buried by deposition of inorganic material, resulting in interlayering of these different sediment types. Channel aggradation and flood scour can cause a shoreline to shift laterally many meters. Seasonal erosion and deposition related to spring floods also create a temporal component to the heterogeneity.

2. *Edge effects*—Shoreline erosion and deposition related to wave action in lakes, large wetlands, and rivers create heterogeneity at the sediment-water interface. Waves erode banks, which subsequently fail as new material slumps into the surface-water body. Fine-grained sediments are moved away from shore, often leaving a cobble- to boulder-sized pavement at the shoreline. Sediment deposition by overland flow commonly results in near-shore, fan-shaped deposits following heavy rainfall. Waves also rework sediments following slump events or sediment transport associated with overland flow, causing movement of fine-grained materials into voids created by movement of cobble- to boulder-sized sediments. In addition, changing surface-water stage causes the position

Figure 2. Typical hydraulic conditions in the vicinity of the shoreline of a surface-water body. (Artwork by Donald O. Rosenberry, U.S. Geological Survey.)
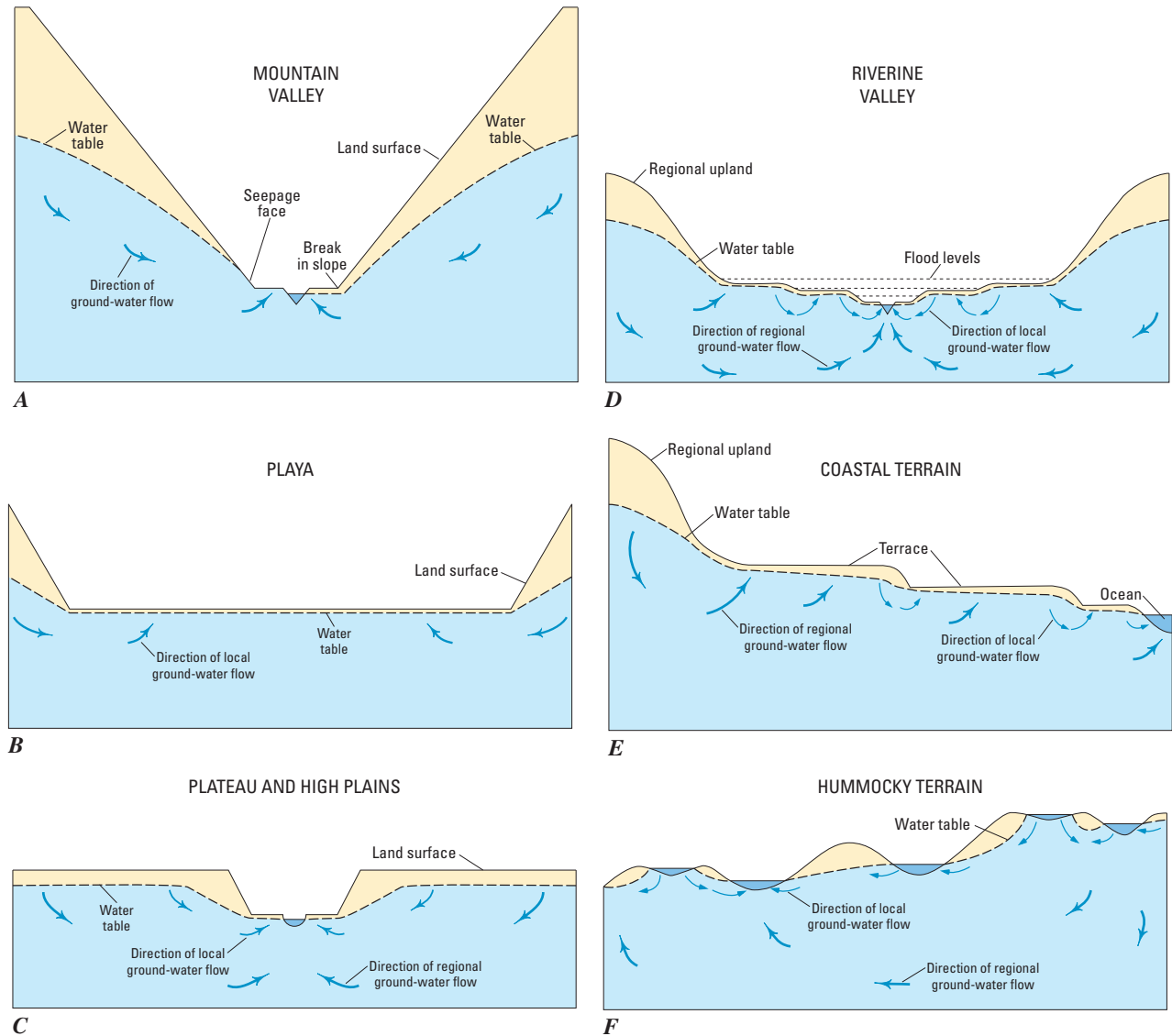
Figure 4.   Generalized hydrologic landscapes: *A*, narrow uplands and lowlands separated by a large steep valley side (mountainous terrain); *B*, large broad lowland separated from narrow uplands by steeper valley sides (playas and basins of interior drainage); *C*, small narrow lowlands separated from large broad uplands by steeper valley side (plateaus and high plains); *D*, small fundamental hydrologic landscape units nested within a large fundamental hydrologic landscape unit (large riverine valley with terraces); *E*, small fundamental hydrologic landscape units superimposed on a larger fundamental hydrologic landscape unit (coastal plain with terraces and scarps); *F*, small fundamental hydrologic landscape units superimposed at random on large fundamental hydrologic landscape units (hummocky glacial and dune terrain) (from Winter, 2001, copyright the American Water Resources Association, used with permission).

of the shoreline to change over time, resulting in lateral movement of all of the previously mentioned depositional and erosional processes that occur at the shoreline. Accumulation of organic debris, including buried logs and decayed plant matter, also contributes to heterogeneity as it is incorporated with the inorganic sediments, particularly on the downwind shores of surface-water bodies. In surface-water bodies that are ice covered during winter, ice rafting during fall and spring, when ice is forming or when the ice cover is melting, can substantially rework sediments at the downwind shoreline.

3. *Biological processes*—Benthic invertebrates constantly rework sediments, particularly organic sediments, as they carry out their life cycles. Bioturbation and bioirrigation are important processes for organic sediments in deeper water environments, but it can be significant in some near-shore settings also. Aquatic birds disturb the sediment as they search for benthic invertebrates, and fish rework sediments as they create spawning redds. Beavers and muskrats can make large-scale disturbances by removing considerable amounts of sediments for lodges and passageways, and the construction of dams.
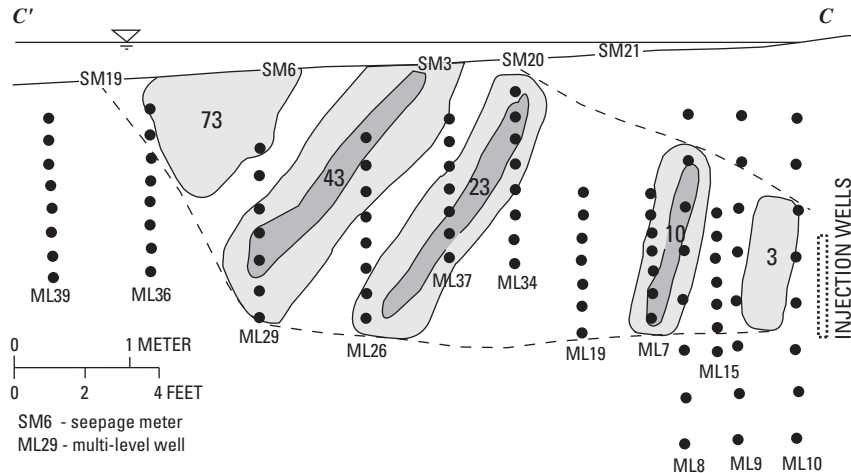
Figure 5.   Flow of salt tracer into a sandy lakebed, Perch Lake, Ontario. Numbers in shaded areas (representing center of mass of salt plume) are the days following salt injection (modified from Lee and others, 1980, copyright 1980 by the American Society of Limnology and Oceanography, Inc., used with permission).

Temporal and Spatial Variability of Flow

Flow across the sediment-water interface commonly changes in direction and velocity temporally and spatially. Many occurrences of spatial and temporal variability in the exchange between surface water and ground water are described in the literature; a few examples are provided herein. Some of this variability is summarized in figure 6. In this illustration, water flows from the surface-water body to ground water through the bottom sediments located beyond a low-permeability layer some distance from shore. Yet closer to shore, ground water flows into the surface-water body. Finally, near the shore a depression in the water table created by evapotranspiration causes flow out of the lake. At the south shoreline of Mirror Lake, New Hampshire, water flows from the lake to ground water between the shoreline and approximately 8 meters from shore, and beyond that point, flow from ground water to the lake occurs (fig. 7) (Asbury, 1990; Rosenberry, 2005).

In many settings, evapotranspiration during the summer months can depress the water table adjacent to the shoreline of wetlands, streams, and lakes below the level of the surface-water body (fig. 8) (Meyboom, 1966, 1967; Doss, 1993; Winter and Rosenberry, 1995; Rosenberry and others, 1999; Fraser and others, 2001). As a result, seasonal, and sometimes diurnal, reversals in flow between surface water and ground water may occur at the shoreline. The changes in direction of flow between surface water and ground water result from fluctuations in the amount of water removed from the water table because of evapotranspiration by plants along the margins of the surface-water body. On a seasonal basis, once evapotranspiration ceases to remove water from the near-shore regions, the near-shore depression in the water table dissipates, which then allows ground water to flow into the surface-water body.

On a diurnal basis, more evaportranspiration in the day and less at night can cause the water table to fluctuate between levels below and above the adjacent surface-water level.

In many locations, water-table mounds can develop at the edge of surface-water bodies. Many studies have shown transient water-table mounds that form in response to precipitation or snowmelt (fig. 9) (see, for example, Winter, 1986; Rosenberry and Winter, 1997; Lee and Swancar, 1997). Most of these water-table mounds were of short duration and formed in response to large rainfall events. Reversals of flow of longer duration also occur at some settings. Jaquet (1976) reported a reversal of flow along part of the shoreline at Snake Lake, Wisconsin, following spring thaw and considerable rainfall (19 centimeters) over a 5-week period that persisted for several months.
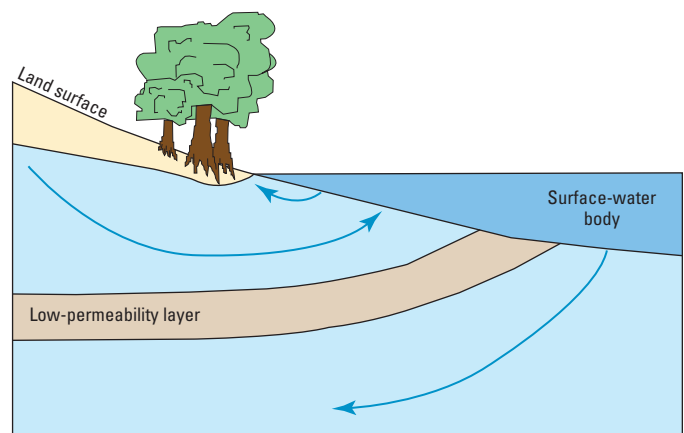
Figure 6.   The length of the flow path and the direction of flow can vary seasonally and with distance from shore. (Artwork by Donald O. Rosenberry, U.S. Geological Survey.)
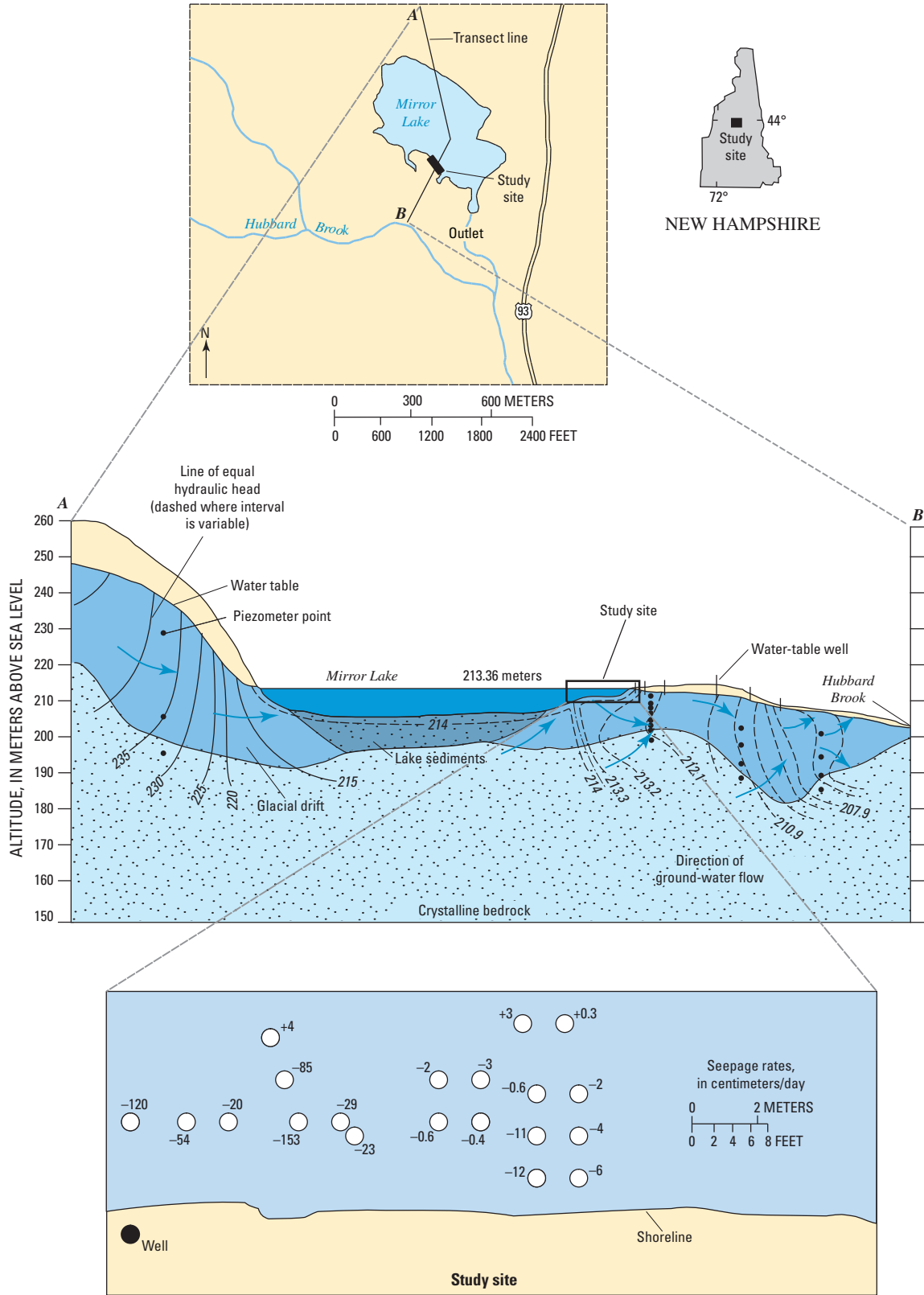
Figure 7.   Example of ground-water exchange with Mirror Lake wherein lake water flows into ground water near shore (shown by negative numbers), and ground water flows into the lake farther from the shoreline (shown by positive numbers). (Modified from Rosenberry, 2005, copyright 2005 by the American Society of Limnology and Oceanography, Inc., used with permission.)
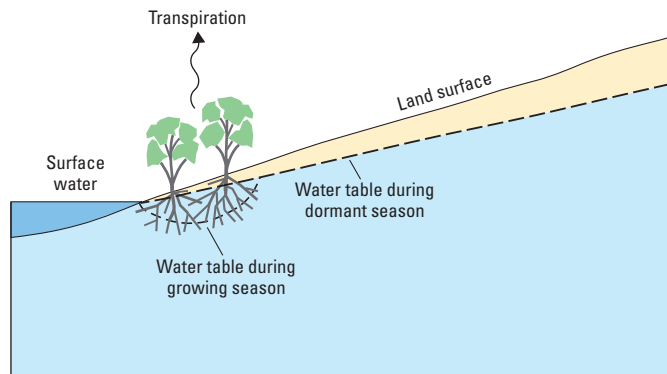
Ground-water levels adjacent to streams also fluctuate in response to the rise and fall of water in the stream (Winter, 1999). An example of the resulting changes in flow direction between a stream and ground water is illustrated by data from the Cedar River in Iowa (fig. 10). In flowing waters, movement of surface water into the subsurface and out again occurs both at the bottom of the stream channel and beneath upland areas between bends in the open channel (fig. 11). This transient flow of surface water into and out of the subsurface is also known as hyporheic flow (Orghidan, 1959). Ground water flowing toward a surface-water body may discharge directly into that body or mix with hyporheic flow prior to emerging into open-water flow. The various interactions with ground water include situations in which flow is parallel to the stream (fig. 12) and does not intersect the surface water (Woessner, 1998, 2000).

Defining the Purpose for Measuring the Exchange of Water Between Surface Water and Ground Water

Water-resource investigators and water-resource managers have many reasons to quantify the flow between surface water and ground water. Perhaps the most common reasons include: calculating hydrological and chemical budgets of surface-water bodies, collecting calibration data for watershed or ground-water models, locating contaminant plumes, locating areas of surface-water discharge to ground water, improving their understanding of processes at the interface between surface water and ground water, and determining the relation of water exchange between surface water and ground water to aquatic habitat. For many investigations, it is sufficient to make a qualitative determination regarding the direction and relative magnitude of flow, either into or out of the surface-water body.

Methods for quantifying flows should be selected to be appropriate for the scale of the study. For a watershed-scale study in which multiple basins may be involved, small-scale flow phenomena, such as near-shore depressions in the water table or spatial variability of flux related to geologic

variability, likely are of little importance to the overall study goal. In such watershed-scale studies, the net flux integrated over an entire stream reach, or lake, or wetland often is the desired result. Watershed-scale flow modeling, ground-water flow modeling, flow-net analysis, or dye- and geochemical-tracer tests, often are used in such large-scale studies, studies on the order of hundreds of meters or a kilometer or more in length or breadth.

If the goal of a study is to identify and (or) delineate zones or areas of flow of surface water to ground water, or flow of ground water to surface water, smaller scale spatial and temporal variations in flow become important, and measurement tools that provide results over an intermediate scale, many tens to hundreds of meters should be selected. In many instances, measurement of surface-water flow at two places some distance apart in a segment of stream, which enables calculation of gains or losses in flow in the segment, is appropriate for these types of studies. For local, small-scale studies in which flow to or from surface water may be focused, small-scale tools such as seepage meters, small portable wells ("minipiezometers" or hydraulic potentiomanometers), and buried temperature probes may be most appropriate. Devices designed to measure flow in a small area are known as seepage meters because the term seep refers to "a small area where water moves slowly to the land surface" (USGS Water Basics Glossary *http://capp.water.usgs.gov/GIP/h2o_gloss/*). Seepage is defined as "the slow movement of water through small cracks, pores, interstices, and so forth, of a material into or out of a body of surface or subsurface water" (USGS Water Science Glossary of Terms *http://ga.water.usgs.gov/edu/dictionary.html#S*).

Once the water-resource investigator has decided on the purpose of the study and the scale of the investigation, methods of investigation can be chosen to most effectively determine where an exchange between surface water and ground water is taking place, the direction of flow, the rate or quantity of that flow, and whether the rate and direction of flow changes over time.

Determining Locations of Water Exchange

The investigator who wishes to determine where water exchange is taking place between surface water and ground water has many options, particularly in the case of ground-water discharge to surface water. Reconnaissance tools useful over larger areas, such as dye-tracer tests, aerial photography and imagery, temperature and specific-conductance probes, and surface-water discharge measurements, can be supplemented by reconnaissance tools useful in smaller areas of interest, such as seepage meters, minipiezometers, and biological indicators.
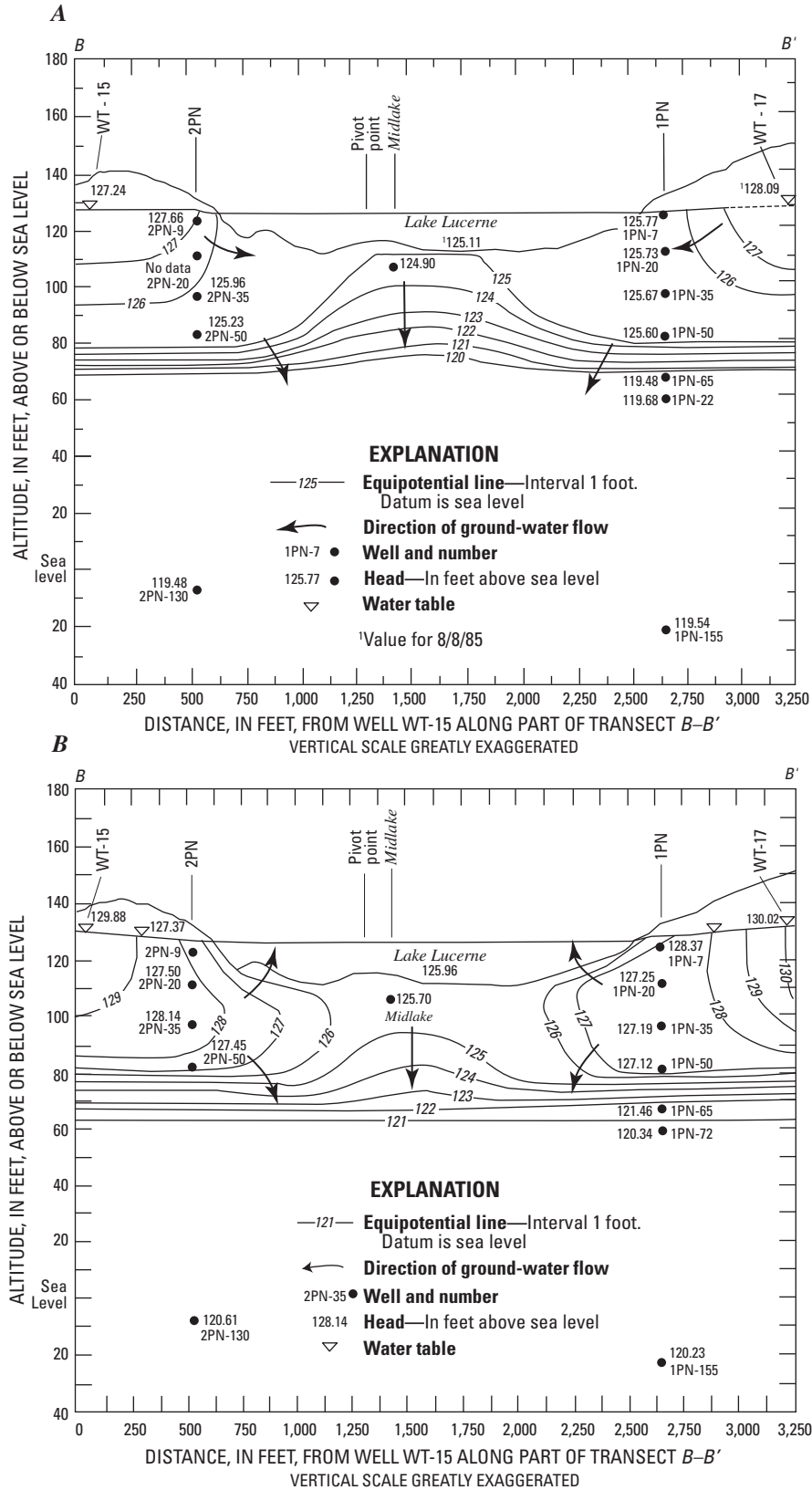
Figure 9. Example of rise and fall of water-table mounds at the edge of surface-water bodies and changes in flow direction (from Lee and Swancar, 1997). *A*, Vertical distribution of head showing downward head gradient conditions, August 6, 1985. *B*, Vertical distribution of head during high water-level conditions, October 17, 1985.
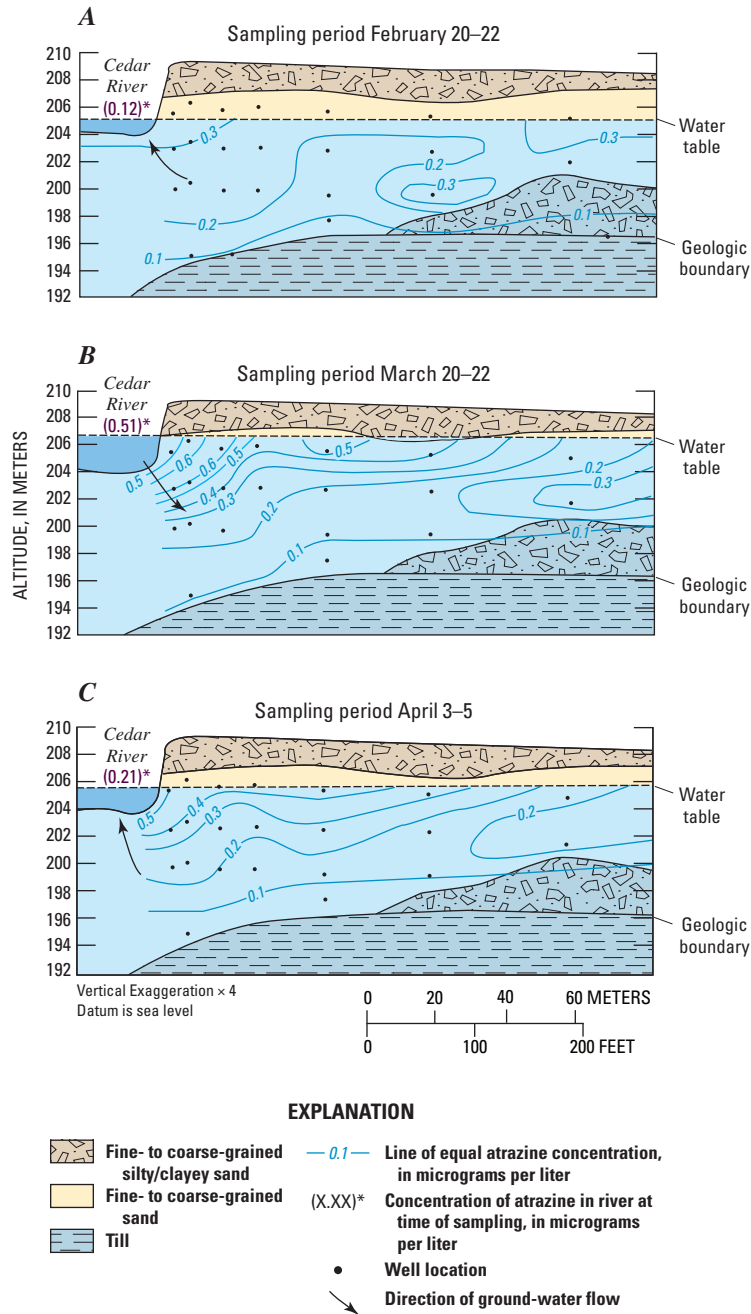
Figure 10. Example of changes in flow direction related to onset and dissipation of a water-table mound adjacent to a river (from Squillace and others, 1993, used in accordance with usage permissions of the American Geophysical Union wherein all authors are U.S. Government employees). Hydrogeologic sections for part of the Cedar River, Iowa, for three periods in 1990. *A*, Movement of ground water into the river prior to a period of high river stage. *B*, Movement of river water into the contiguous aquifer during high river stage. *C*, Return of some of the water from the aquifer during declining river stage.

A. View from Above

B. Sectional View

Figure 11. Example of flow interaction between surface water and ground water (from Dumouchelle, 2001). Schematic of flow in Chapman Creek, west-central Ohio. (Arrows indicate direction of flow. Diagrams not to scale.)

commonly known as a "seepage run" (Harvey and Wagner, 2000). If the amount of flow in the stream has increased over the reach, the increase may be attributed to ground-water discharge to the stream. If flow in the selected reach of stream has decreased, the decrease may be attributed to surface water flowing into ground water. It is important to recognize, however, that the direction of flow indicated by any change in streamflow is a "net direction" over the selected reach, and that within the reach, water may be moving into and out of the stream (and conversely, into and out of the underlying aquifer). It is important to account for any inflows or outflows within the stream reach, such as diversions for irrigation or channelized return flows from fields.

Measuring the Quantity of Flow

The volume of water flowing between surface water and ground water, either as surface water into ground water or ground water into surface water, can be measured directly with seepage meters. Measurement of changes in water temperatures over time at a specific site above the sediment, at the sediment-water interface, and within the sediment makes possible the determination of the amount of water exchange occurring between surface water and ground water. The exchange of water between surface water and ground water also can be examined and estimated by using dye tracer tests or by using other tracers. Such dyes or tracers are added directly to a stream and then their concentrations are measured at some point or points downstream. Changes in the concentration of the dye or other tracer over time downstream from where they are injected enables calculation of ground-water inputs.

Measuring Temporal Variations in Flow

In many instances, the rate of exchange between surface water and ground water varies over time scales of hours, days, or months. The direction of flow also may reverse on a seasonal basis or temporarily during a flood, for example. Measuring temporal variation in the rate of water exchange requires multiple measurements over these time periods. Measuring devices equipped with data recorders ("data loggers") enable the investigator to record repeated measurements at specified time intervals to document temporal changes.

Measuring Direction of Flow

Comparison of surface-water levels and adjacent ground-water levels indicates direction of flow. If the surface-water level is higher than adjacent ground-water levels, the direction of flow is from the surface water to ground water. If the opposite is the case—ground-water levels are higher than nearby surface-water levels—then the direction of flow is from ground water to surface water. In addition to indicating the direction of flow, water-level measurements provide information about the magnitude of the hydraulic gradients between surface water and ground water. In some instances, however, these gradients can be altered locally. For example, vegetation between the wells and the edge of the surface-water body can transpire sufficient water to cause a local depression in the water table close to the edge of the surface water (Meyboom, 1966, 1967; Doss, 1993; Rosenberry and Winter, 1997; Fraser and others, 2001). Thus, it can be important to measure the direction of flow at a local scale using portable wells, minipiezometers, or hydraulic potentiomanometers.

Another way to determine if a section of stream or river is receiving ground-water discharge or is losing water to the underlying aquifer is by measurement of surface-water flow at two places some distance apart in a reach of stream, a practice

Figure 12.   Fluvial-plain ground-water and stream-channel interactions showing channel cross sections classified as: *A*, gaining; *B* and *C*, losing; *D*, zero exchange; and *E*, flow-through. The stream is dark blue. The water table and stream stage (thicker lines), ground-water flow (arrows), and equipotential lines (dashed) are shown (from Woessner, 1998, copyright American Institute of Hydrology, used with permission).

Methods of Investigation

Common methods to examine exchange of water between surface-water bodies and ground-water bodies are described below. Some of these methods make use of already installed hydrological instruments and existing data, rather than requiring the investigator to make measurements of hydrologic characteristics. When using such methods, however, the investigator may install wells, stream-gaging equipment, or rain gages, as needed, to obtain sufficient data to make the application of methods possible and the results less uncertain. Other methods require that the investigator make additional, specific measurements or observations of hydrological, physical, chemical, or biological characteristics.

Watershed-Scale Rainfall-Runoff Models

Many analytical and numerical models that relate precipitation, ground-water recharge, and ground-water discharge to temporal variability of flow in a stream have been developed. A fundamental assumption in these models is that streamflow is an integrated response to these processes over the stream's watershed, and that ground-water discharge to the stream provides the steady flow in the stream between rainfall events, commonly referred to as baseflow. Analytical models generally determine baseflow through hydrograph separation techniques. Several automated routines have been developed to assist in this determination (Rutledge, 1992; Rutledge, 1998) (fig. 13). Other analytical methods also have been used to quantify the interaction between ground water and surface water, including an analytic-element method (Mitchell-Bruker and Haitjema, 1996) and a nonparametric regression model (Adamowski and Feluch, 1991).

Several numerical models commonly referred to as rainfall-runoff models have been developed; these models are ally divide watersheds and subwatersheds and calculate hydrologic parameters for each smaller area (for example, Federer and Lash, 1978; Leavesley and others, 1983, 1996, 2002; Beven and others, 1984; Beven, 1997; Buchtele and others, 1998). Rainfall-runoff models generally are calibrated to match river flow at the outlet of a watershed or subwatershed. Some models include the ground-water component of flow in each area. The current trend is to couple distributed-area watershed-scale models with ground-water flow models in order to better determine the temporal and spatial variability of the interaction between ground water and surface water (for example, Leavesley and Hay, 1998; Beven and Feyen, 2002).

A

B
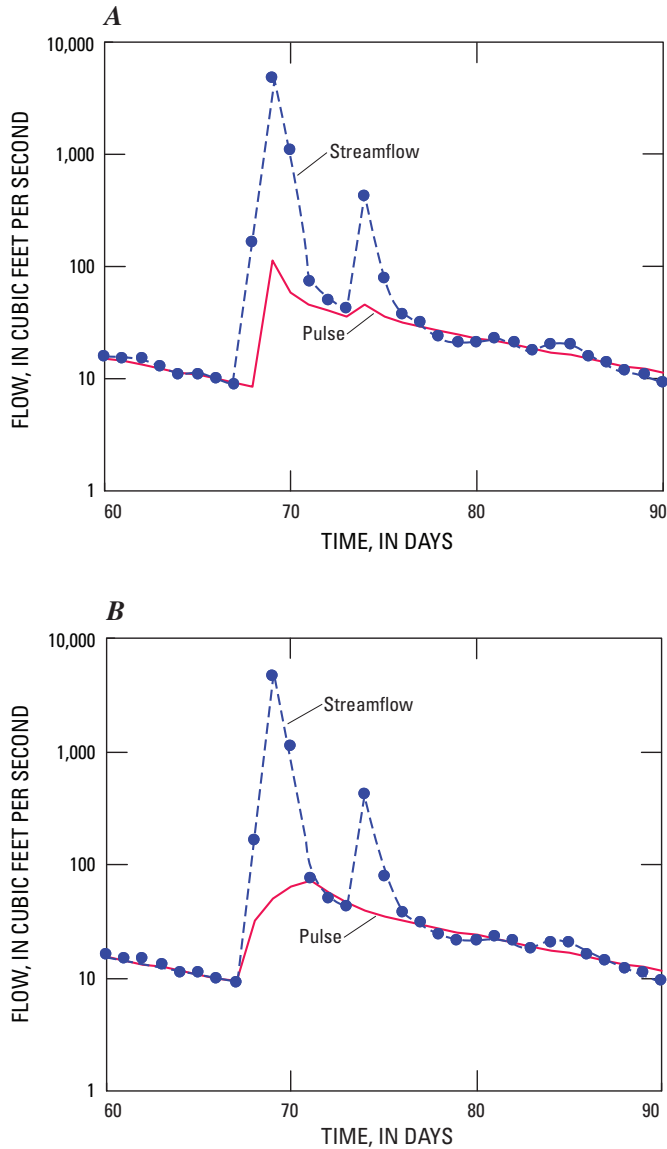
Figure 13. Example of the use of hydrographs to determine the amount of ground-water discharge to a stream (from Rutledge, 2000). Hydrographs of streamflow for Big Hill Creek near Cherryvale, Kansas, for March 1974 (blue circles and dashed line), and hydrograph of estimated ground-water discharge using the PULSE model (red line). (Note: In each example, the total recharge modeled is 0.73 inch, which is the same as the total recharge estimated from RORA [a recession-curve-displacement method for estimating recharge] for this period. In example *A*, recharge is modeled as 0.65 inch on day 69 and 0.08 inch on day 74. In example *B*, recharge is modeled as a gradual process that is constant from day 68 to day 72).

Stream Discharge Measurements

Measurements of stream discharge (Rantz and others, 1982a, b; Oberg and others, 2005) made as part of seepage runs (described earlier) can be used to determine the occurrence and rate of exchange of water between surface water and ground water in streams and rivers (fig. 14). The results of seepage runs have been used to provide an integrated value for flow between a stream and ground water along a specific stream reach. This method works well in small streams, but for larger streams and rivers, the errors associated with the measurement of flow in the channel often are greater than the net exchange of water to or from the stream or river. This method also requires that any tributaries that discharge to a stream along the reach of interest be measured and subtracted from the downstream discharge measurement. Likewise, withdrawals from the stream, such as that for irrigation, must be measured and added to the downstream discharge measurement.

The application of seepage-run data, however, is limited by the ratio of the net flow of water to or from the stream along a stream reach to the flow of water in the stream. The net exchange of water across the streambed must be greater than the cumulative errors in streamflow measurements. For example, if the errors in the stream discharge measurements are 5 percent of the true, actual flow, then according to the rules of error propagation, in order to be able to detect the net flow of water to or from the stream along the reach of interest, the value of net flow must be greater than 7 percent of the streamflow. Despite these limitations, many hydrologic studies have made use of this method with good results [for example, Ramapo River, New Jersey–Hill and others (1992); Bear River, Idaho and Utah–Herbert and Thomas (1992); Souhegan River, New Hampshire–Harte and others (1997); Lemhi River, Idaho–Donato (1998); constructed stream channel Baden-Württemberg, Germany–Kaleris (1998)]. The information gained from seepage runs can be enhanced with data obtained by using other techniques such as minipiezometers, seepage meters, temperature and specific-conductance measurements to better define surface-water/ground-water fluxes [for example, creeks and rivers in the Puget Sound area of Washington–Simonds and others (2004); and Chapman Creek, Ohio–Dumouchelle (2001)]. Seepage-run results also can provide estimates of hydraulic conductivity of the streambed on a scale appropriate for ground-water flow modeling (Hill and others, 1992).

Ground-Water Flow Modeling

Since 1983, most investigators who have used the numerical modeling approach in the quantification of flows between surface water and ground water have used the U.S. Geological Survey MODFLOW modular modeling code (Harbaugh and others, 2000). This finite-difference model contains an original "river package" that can simulate flows to or from a river, assuming the river stage does not change during a specified time period (referred to as a stress period in MODFLOW), but can change from one time period to the next. Several other MODFLOW modules or packages also have been developed to simulate fluxes between surface water and ground water. These include streamflow routing packages (Prudic, 1989;
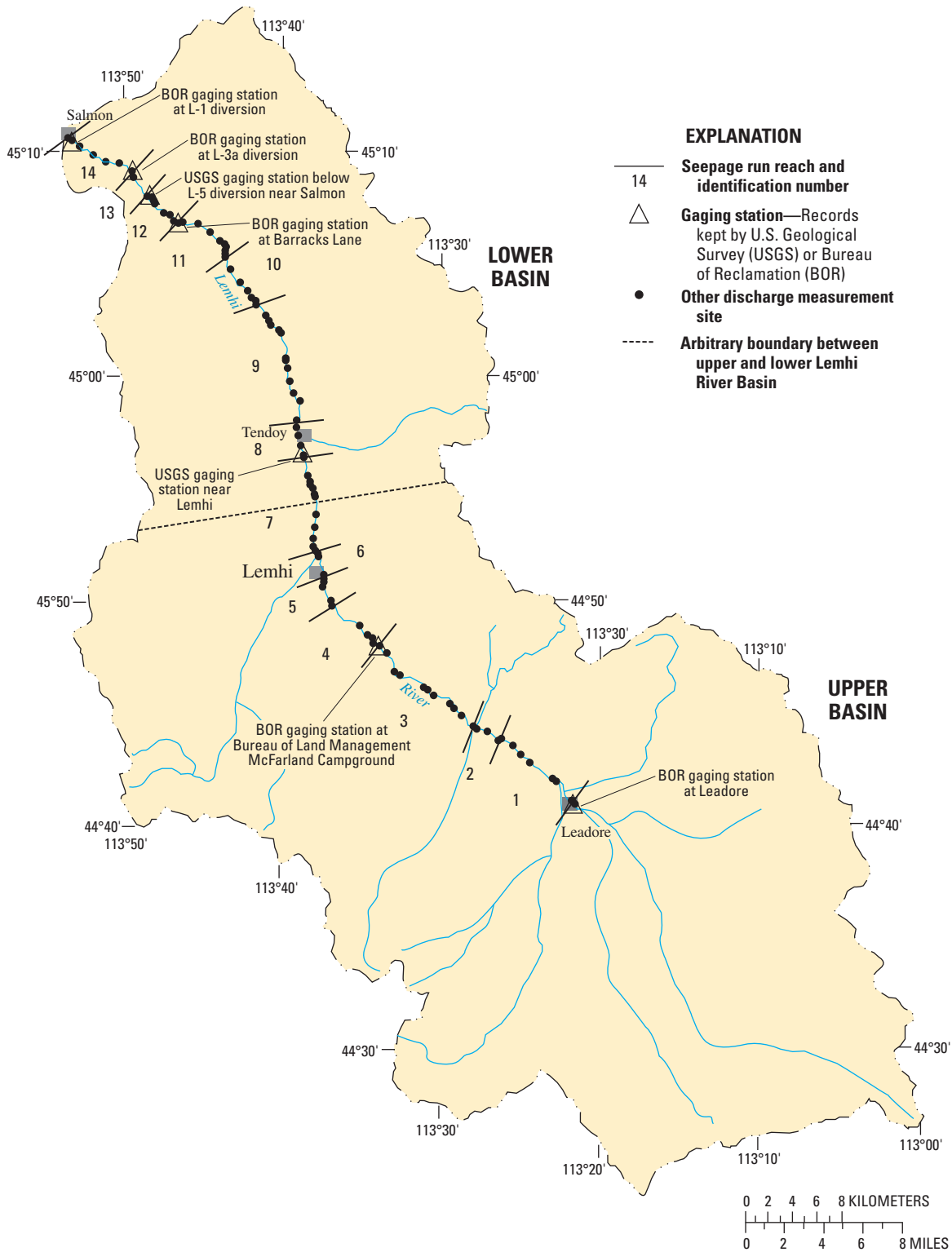
Figure 14.   Example of the use of discharge measurements to determine surface-water/ground-water interaction (from Donato, 1998). Seepage run reaches, gaging stations, and discharge measurement sites in the Lemhi River Basin, east-central Idaho, August and October 1997.
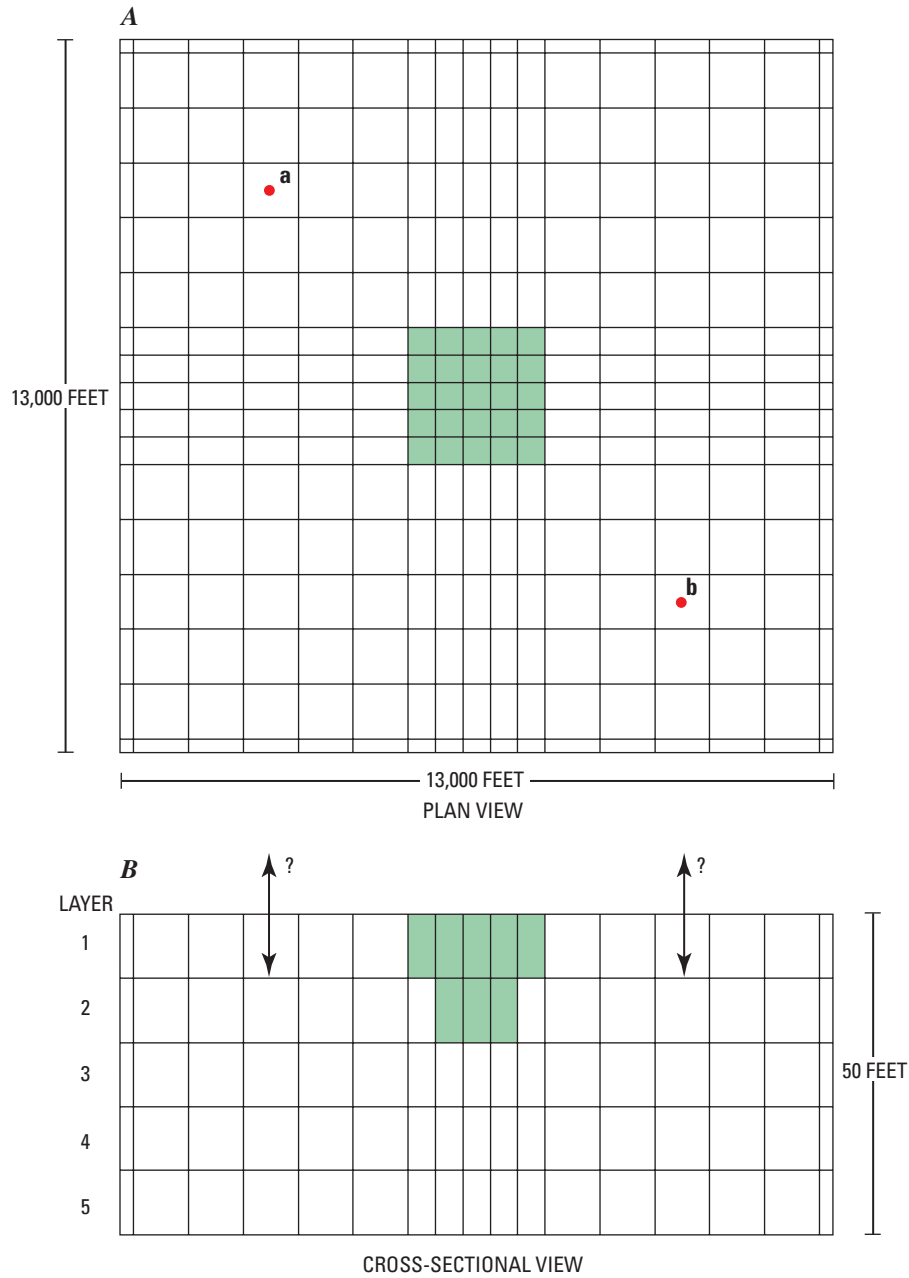
Figure 15.   Example of model grid used for simulation of surface-water/ground-water interaction (from Merritt and Konikow, 2000). The lateral and vertical grid discretization for test simulation 1: *A*, Plan view—Shaded area is the surface extent of lake cells in layer 1. Interior grid dimensions are 500 and 1,000 meters. Border row/column cells are 250 feet thick. The locations of hypothetical observations wells are denoted by a and b. *B*, Cross-sectional view—Shaded area is the cross section of the lake. Although a nominal 10-foot thickness is shown for layer 1, the upper surface of layer 1 is not actually specified, and lake stages and aquifer water-table altitudes may rise higher than the nominal surface shown above.

Prudic and others, 2004), a reservoir package (Fenske and others, 1996), a lake package (Cheng and Anderson, 1993), and more recently, a more elaborate lake package (Merritt and Konikow, 2000) (fig. 15). More advanced MODFLOW-based programs have been developed to couple one-dimensional, unsteady streamflow routing with MODFLOW (Jobson and Harbaugh, 1999; Swain and Wexler, 1996). Advances also are being made in coupling MODFLOW with watershed models that simulate many of the surface-water processes within a basin (Sophocleous and others, 1999; Sophocleous and Perkins, 2000; Niswonger and others, 2006). One of the most challenging aspects of coupling ground-water and surface-water models has been representation of flow through the unsaturated zone beneath a stream. Two new programs have recently been developed for MODFLOW to simulate one-dimensional (Niswonger and Prudic, 2005) and three-dimensional (Thoms and others, 2006) flow in the unsaturated zone. Many of these packages require a determination of the transmissivity of the sediments at the interface between the aquifer and the surface-water body. Transmissivity is determined by multiplying hydraulic conductivity by the thickness of the lakebed or riverbed sediments.

Direct Measurement of Hydraulic Properties

The relation between the stage of a surface-water body and the hydraulic head measured in one or more nearby water-table wells can be used to calculate flows of water between surface water and ground water [Williams Lake, Minnesota–LaBaugh and others (1995); Vandercook Lake, Wisconsin–Wentz and others (1995); large saline lakes in central Asia–Zekster (1996); Lake Lucerne, Florida–Lee and Swancar (1997); Waquoit Bay, Cape Cod, Massachusetts–Cambareri and Eichner (1998); Otter Tail River, Minnesota–Puckett and others (2002)]. The Darcy equation (eq. 1) is used to calculate flow between ground water and surface water along specific segments of shoreline.

$$Q = KA\frac{h_1 - h_2}{L}, \qquad (1)$$

where

Q is flow through a vertical plane that extends beneath the shoreline of a surface-water body (L^3/T),

A is the area of the plane through which all water must pass to either originate from the surface-water body or end up in the surface-water body, depending on the direction of flow (L^2),

K is horizontal hydraulic conductivity (L/T),

h_1 is hydraulic head at the upgradient well (L),

h_2 is hydraulic head at the shoreline of the surface-water body (L),

and

L is distance from the well to the shoreline (L).

Shoreline segments are delineated/selected on the assumption that the gradient between a nearby well and the surface-water body, the hydraulic conductivity of the sediments, and the cross-sectional area through which water flows to enter or leave the lake, are uniform along the entire segment (fig. 16). Flows through each segment are summed for the entire surface-water body to compute net flow. The scale of the shoreline segments, and the scale of the study, depend on the scale of the physical setting of interest and the density of monitoring wells. Further detail regarding this method is provided in Chapter 2, in the section "Wells and Flow-Net Analysis."

Examination and Analysis of Aerial Infrared Photography and Imagery

Aerial infrared photography and imagery have been used to locate areas of ground-water discharge to surface waters (Robinove, 1965; Fischer and others, 1966; Robinove and Anderson, 1969; Taylor and Stingelin, 1969). This technique is effective only if the temperatures of surface water and ground water are appreciably different. Information obtained from infrared scanners can be captured electronically or transferred to film, on which tonal differences correspond to differences in temperature (Robinove and Anderson, 1969; Banks and others, 1996) (fig. 17). Published studies indicate tonal differences corresponding to a difference in temperature of approximately 2 degrees Celsius are distinguishable (Pluhowski, 1972; Rundquist and others, 1985; Banks and others, 1996).

Within the limits of the ability of infrared imagery to distinguish temperature differences between surface water and ground water, the inspection of such imagery enables more rapid identification of gaining reaches in streams over large areas than can be accomplished by stream surveys that measure temperature directly (Pluhowski, 1972). Another advantage of this method in identifying areas of ground-water discharge to surface-water bodies is its application where using other techniques such as dye tracing or direct temperature measurements are impractical, or access on the ground is difficult (Campbell and Keith, 2001) or dangerous (Banks and others, 1996).

Using thermal-infrared imagery to distinguish zones of ground-water discharge is practical for locating diffuse and focused ground-water discharge (Banks and others, 1996). This capability has been demonstrated in a variety of environments. Examples for lakes are Crescent Lake, Nebraska (Rundquist and others, 1985), where the flow is diffuse and occurs over a large area, and Great Salt Lake and Utah Lake in Utah (Baskin, 1998), where the ground-water flow into the lake is focused at springs. Campbell and Keith (2001) found the technique useful in locating many springs flowing into streams and reservoirs in northern Alabama. Examples for estuaries are creeks flowing into Chesapeake Bay, Maryland, and the shorelines of the Gunpowder River and the Chesapeake Bay into which the river flows (Banks and others,
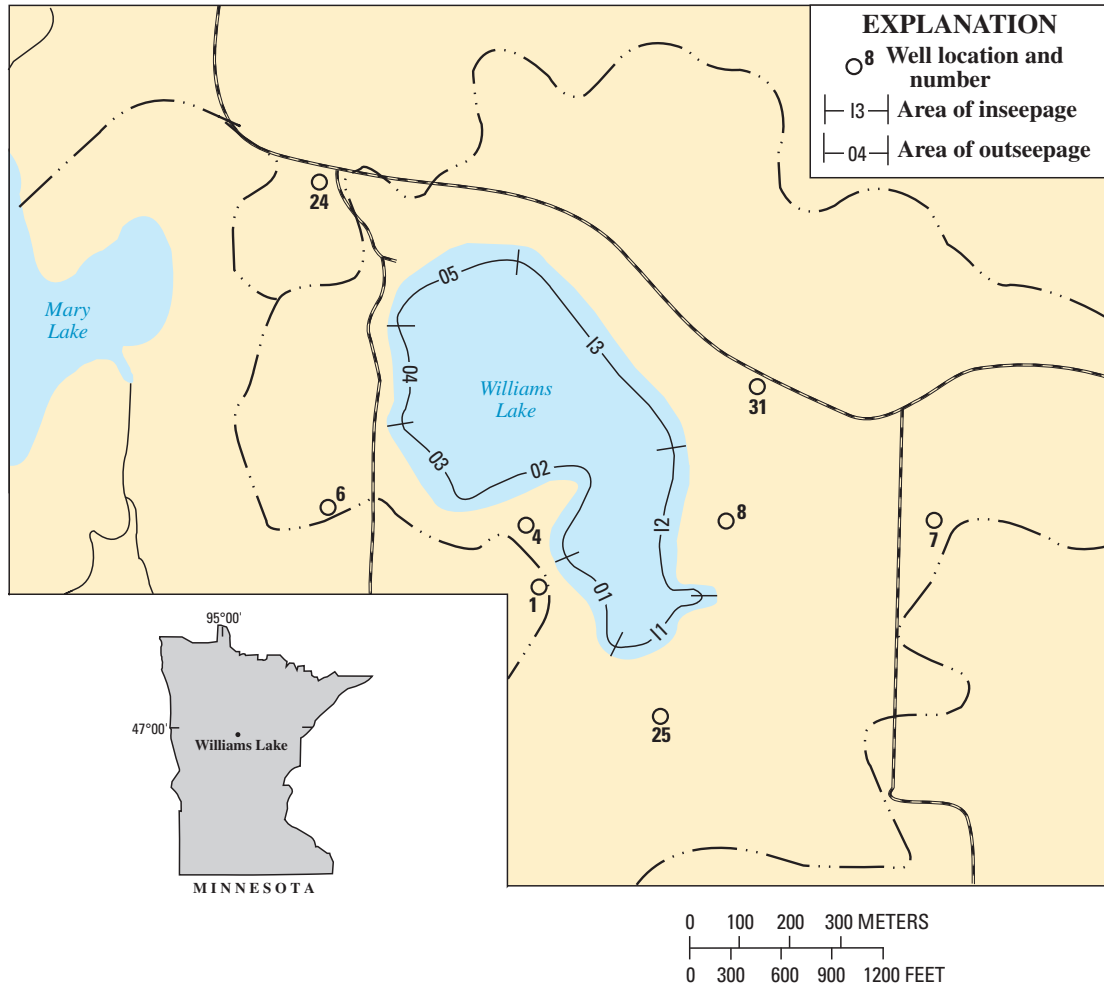
Figure 16.   Example of shoreline segment definition for the calculation of water fluxes between surface water and ground water at a lake (modified from LaBaugh and others, 1995, used in accordance with author rights of the National Research Council of Canada Press). Location of wells and shoreline segments used to calculate flow between surface water and ground water at Williams Lake, Minnesota.

1996), as well as creeks and rivers flowing into Long Island Sound, New York (Pluhowski, 1972). Examples of the use of infrared imagery to detect areas of ground-water discharge to marine waters include the delineation of areas of diffuse ground-water flow into Long Island Sound (Pluhowski, 1972) and focused ground-water flow as springs to the ocean, such as around the perimeter of the island of Hawaii (Fischer and others, 1966).

Dye and Tracer Tests

Dyes and other soluble tracers can be added to water and then "tracked" to provide direct, qualitative information about ground-water movement to streams. Fluorescent dyes that are readily detected at small concentrations and pose little environmental risk make a useful tool for tracing ground-water flow paths, particularly in karst terrane (Aley and Fletcher, 1976; Smart and Laidlaw, 1977; Jones, 1984; Mull and others,

1988). Thus, dye-tracer studies can be used to determine the time-of-travel for ground water to move to and into surface water, as well as hydraulic properties of aquifer systems (Mull and others, 1988). The use of dyes as tracers is described in more detail in Chapter 3. Commonly, a reconnaissance of the ground-water basin is made to identify likely areas of potential surface-water flow into ground water or ground-water flow to the surface. An inventory is made of springs, sinkholes, boreholes or screened wells, and sinking streams. Appropriate sites then are picked for dye injection, and the potential discharge areas, springs, and stream reaches are monitored over an appropriate period of time, hours or days, for appearance of the dye (fig. 18).

Solute tracers have been used to aid in the determination of water gains or losses within the channel of a stream or river (Kilpatrick and Cobb, 1985). This technique is known as dilution-gaging. A variety of tracers have been used in such studies, either alone or in combination, usually including a

Figure 17.   Example of use of thermal infrared imagery to delineate areas of discrete and diffuse ground-water discharge to surface water (from Banks and others, 1996, reprinted from Ground Water with permission from the National Ground Water Association, copyright 1996, thermal imagery of O-Field study area, Aberdeen Proving Ground, Maryland).

Figure 18.   Example of the use of dye to examine water fluxes between surface water and ground water (from Carter and others, 2002). Dye testing has been done in Boxelder Creek, South Dakota, which can lose as much as 50 cubic feet per second of flow to the bedrock aquifers. In the upper left photograph, nontoxic, red dye is poured into Boxelder Creek upstream from a major loss zone. In the upper right photograph, dye in the stream can be seen disappearing into a sinkhole in the Madison Limestone. In the bottom photograph, dye in the stream emerges downstream at Gravel Spring, which is about 671 meters (2,200 feet) (linear distance) from the major loss zone. The length of time for the first arrival of dye to travel this distance is variable depending on flow conditions but generally is about 1 to 2 hours (Strobel and others, 2000). Thus, the ground-water velocity is about 0.3 to 0.6 kilometer per hour (0.2 to 0.4 mile per hour), which is a very fast rate for ground water. Dye also has been recovered at City Springs, which is in the Rapid Creek Basin, about 30 days after injection. This demonstrates that ground-water flow paths are not necessarily restricted by surface-water drainage basins. (Photographs by Derric L. Isles, South Dakota Department of Environment and Natural Resources.)

tracer expected to be nonreactive in the waters of the stream to which it is added, such as lithium (for example, in the Snake River, Colorado–Bencala and others, 1990) or chloride (for example, Chalk Creek, Colorado–Kimball, 1997). In this technique, a known quantity of solute is added at a specified rate for a short interval of time at the upstream cross section of the stream segment of interest, and concentrations of the solute are measured at one or more points downstream over time. Discharge is calculated from the amount of dilution that occurs at the downstream point or points. Kimball (1997) indicated that Q_s (the discharge in the stream) is calculated as follows:

$$Q_s = (C_i Q_i) / (C_B - C_A) \qquad (2)$$

where

C_i is the tracer concentration in the injection solution,

Q_i is the rate of injection into the stream,

C_B is the tracer concentration downstream,

and

C_A is the tracer concentration upstream from the injection point.

When coupled with stream-segment discharge measurements, use of solute tracers also enables calculation of the rates of ground-water inflow and outflow within a stream segment (Harvey and Wagner, 2000). At the same time the solute is injected and monitored within the stream segment, physical velocity measurements (streamflow) are made at the upstream and downstream sections of the stream reach. The streamflow measurements provide information on whether or not there was a net loss or gain of flow within the reach due to interaction with ground water. Harvey and Wagner (2000) indicate the solute tracer, or dilution-gaging, values determine ground-water inflow. Thus, ground-water outflow can be calculated by subtracting the net loss or gain from the solute tracer-derived ground-water inflow value.

Calculation of chemical budgets for a stream, lake, or wetland is another way in which solutes can be used to make quantitative estimates of surface-water exchange with ground water. Conservative chemicals in a watershed are those that are not altered by the porous media through which they flow, and occur at concentrations for which changes in concentration because of chemical precipitation are not likely to occur. Conservative chemicals can be used to determine the volume of ground water that flows into or out of a surface-water body, provided that all other fluxes are known. A common form of the chemical-budgeting equation for a lake or wetland is

$$P(C_P) + GWI(C_{GWI}) + SI(C_{SI}) - GWO(C_{GWO})$$
$$- SO(C_{SO}) = \Delta V_L(C_L) \pm R, \qquad (3)$$

where

P is precipitation,

GWI is ground-water flux into lake or wetland,

SI is streamflow into lake or wetland,

E is evaporation,

GWO is flux of lake or wetland water to ground water,

SO is streamflow out of lake or wetland,

ΔV_L is change in lake or wetland volume,

C_x is chemical concentration of hydrologic component,

and

R is residual.

Chemical budgets have been calculated in lake and wetland studies where water exchange between surface-water bodies and underlying ground water was of interest [see, for example, Rawson Lake, Ontario–Schindler and others (1976); Thoreau's Bog, Massachusetts–Hemond (1983); Williams Lake, Minnesota–LaBaugh and others (1995); LaBaugh and others (1997); multiple lakes in Polk and Highlands Counties, Florida–Sacks and others (1998); Lake Kinneret, Israel–Rimmer and Gideon (2003)]. The equation can be modified to solve for any unknown flow term, provided that the remainder of the flow terms are known. The accuracy of the method depends greatly on the accuracy of the other flow and chemical-concentration measurements. The size of the residual term often is considered a general indicator of the accuracy of the method, but a small residual does not always indicate an accurate chemical balance. LaBaugh (1985) and Choi and Harvey (2000) provide examples of the use of error analysis to quantify the uncertainty associated with water-flux results obtained using this method.

The ratios of the isotopes of oxygen and hydrogen present in water have been used for decades to distinguish sources of water, including ground-water discharge to surface-water bodies (for example, Dincer, 1968). These isotopes are useful because they are part of the water and not solutes dissolved in the water. The method works well when the degree of isotopic fractionation of the water is different for different sources of water. The process of evaporation tends to remove lighter isotopes, leaving the heavier isotopes behind. Thus, the ratio of lighter to heavier isotopes will change over time in the water and the water vapor. More detailed explanation of the isotopic fractionation in catchment water is given in Kendall and others (1995). If the isotopic compositions of different sources of water are distinct, then simple mixing models can be used to identify sources of water. A brief example is presented here, but more detailed explanations and examples of the use of this method can be found in Krabbenhoft and others (1994), LaBaugh and others (1997), Sacks (2002) (applied

to lakes), Kendall and others (1995) (a brief description of the methods), and Kendall and McDonnell (1998) (detailed descriptions of numerous isotopic methods).

For determination of ground-water discharge to streams and rivers, a simple two-component mixing model often is used:

$$Q_S \delta_S = Q_{GW} \delta_{GW} + Q_P \delta_P \qquad (4)$$

where

Q	is discharge,	
δ	is the stable-isotopic composition in parts per thousand enrichment or depletion ("per mil") relative to a standard,	
S	is stream water,	
GW	is ground water,	

and

P is precipitation.

For lakes and wetlands, where sources of water are more numerous, slightly more complex mixing models can be used, such as those provided by Krabbenhoft and others (1990):

$$GWI = \frac{P(\delta_L - \delta_P) + E(\delta_E - \delta_L)}{\delta_{GWI} - \delta_L} \qquad (5)$$

or that provided by Krabbenhoft and others (1994):

$$GWO = \frac{P \cdot \delta_P - \Delta V_L \cdot \delta_L - GWI \cdot \delta_{GWI}}{\delta_L}, \qquad (6)$$

where

GWI	is ground-water flux into lake,
GWO	is flux of water from lake to ground water,
P	is precipitation,
E	is evaporation,
ΔV_L	is change in the volume of the lake,

and

δ_X is per mil value for hydrologic component.

Where equations 5 and 6 are derived from equation 3 applied to stable isotopes at steady state:

$$d(V\delta_L)/dt = GWI\, \delta_{GWI} + P\, \delta_P + Si\, \delta_{Si}$$
$$- GWO\, \delta_L - E\, \delta_E - So\, \delta_L = 0. \qquad (7)$$

Investigation of ground-water discharge into inland and marine surface water also is feasible through measurement of radon and radium isotopes (Corbett and others, 1998; Moore, 2000). In the radon isotope method, a mass balance is constructed for radon-222 (^{222}Rn), which is a chemically and biologically inert radioactive gas formed by the disintegration of the parent nuclide radium (Corbett and others, 1998, 1999). Because radon is a gas, radon in water in contact with the atmosphere will be lost from that water because of volatilization. Thus, ground water commonly contains higher activities of ^{222}Rn than does surface water, from 3 to 4 orders of magnitude greater (Burnett and others, 2001). ^{222}Rn is a radioactive daughter isotope of radium 226 (^{226}Ra) and has a half life of 3.82 days. Determination of the activity of ^{222}Rn and ^{226}Ra in surface water enables the calculation of the ^{222}Rn excess—how much more ^{222}Rn is present in surface water than would be expected based on the ^{226}Ra content of the water. Determination of the activity of ^{222}Rn and ^{226}Ra in sediment water or ground water is used to determine the ^{222}Rn flux into surface water (Cable and others, 1996; Corbett and others, 1998), which can account for the excess ^{222}Rn in the surface water. The mass balance or flux of ^{222}Rn has been used to determine ground-water discharge to several types of surface-water bodies (Kraemer and Genereaux, 1998): in streams, such as the Bickford watershed, Massachusetts (Genereaux and Hemond, 1990); rivers, such as the Rio Grande de Manati, Puerto Rico (Ellins and others, 1990); lakes, such as Lake Kinneret, Israel (Kolodny and others, 1999); estuaries, such as Chesapeake Bay (Hussain and others, 1999), Charlotte Harbor, Florida (Miller and others, 1990), and Florida Bay (Corbett and others, 1999), as well as the coastal ocean, such as in Kanaha Bay, Oahu, Hawaii (Garrison and others, 2003); the Gulf of Mexico off of Florida (Cable and others, 1996; Burnett and others, 2001); and the Atlantic Ocean off the coast of South Carolina (Corbett and others, 1998).

In the radium isotope method, the surface-water activities of the four naturally occurring radium isotopes—^{226}Ra, ^{228}Ra, ^{223}Ra, and ^{224}Ra—are compared to activities in sediment water or ground water to determine fluxes (Moore, 2000). The source of the radium isotopes is the decay of uranium and thorium in sediments or rocks. Water in contact with solid materials containing the source of the isotopes will accumulate the isotopes. Ground water will accumulate more of the isotopes because of water's presence within the matrix of the sediments or rocks. Surface waters will accumulate less of the isotopes because the sediments or rocks are less abundant relative to the water (Kraemer, 2005). Kraemer indicates the ratio of the longer lived isotopes (^{226}Ra half-life of 1,601 years, ^{228}Ra half-life of 5.8 years) can be used as an indicator of the types of sediments or rocks through which ground water has traveled, because of differences in uranium and thorium content between rock types. Kraemer (2005) also notes that the short-lived isotopes (^{223}Ra half life of 11.4 days, and ^{224}Ra half-life of 3.7 days) provide some indication of the timing of ground-water discharge. Naturally occurring radium isotopes have been useful in the identification of ground-water inflow to lakes, such as Cayuga Lake, New York (Kraemer, 2005), freshwater wetlands in the Florida Everglades (Krest and Harvey, 2003), estuarine wetlands, North Inlet salt marsh, South Carolina (Krest and others, 2000), and coastal waters in the central South Atlantic Bight (Moore, 2000).

Thermal Profiling

Measurements of temperature have been used to determine qualitatively the locations of rapid discharge of ground water to surface water; common measurement methods include towing a tethered temperature probe, in-situ measurements of temperature, and thermal imagery (Lee, 1985; Baskin, 1998; Rosenberry and others, 2000). Temperature also can be measured at different depths beneath a stream or lake to determine the rate of vertical flow through a surface-water bed either into or out of the surface-water body (Lapham, 1989). This method is effective when flow through the lakebed is sufficient to allow advective processes to be significant relative to conductive temperature signals. The method requires multiple measurements of temperature over weeks or months and the simultaneous solution of the flow of fluid and heat in one dimension. Depths of temperature measurement typically extend up to 3 to 6 meters beneath the sediment bed. The solution requires the assumption that flow is vertical through the surface-water bed and that the media are homogeneous and isotropic. Taniguchi (1993) used seasonal changes in sediment temperature beneath a surface-water body to develop type curves that can be used to estimate vertical fluxes through the surface-water bed.
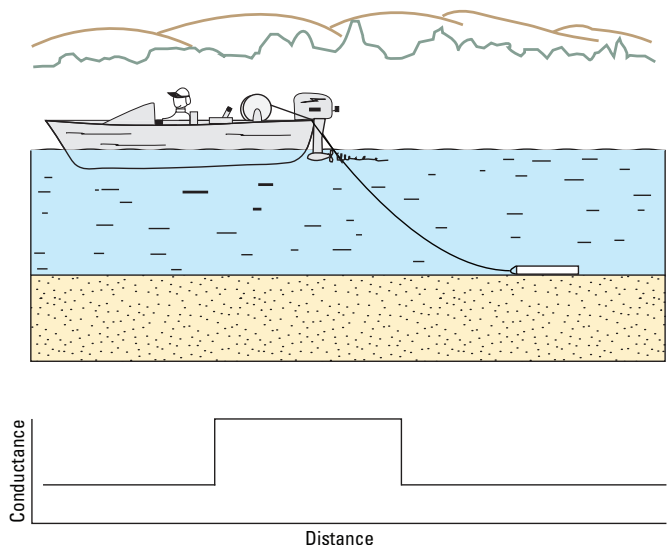
Subsediment temperature has been used over short distances beneath the surface-water bed and makes use of the temperature response in the sediments to diurnal changes in surface-water temperature (Constantz and others, 1994; Stonestrom and Constantz, 2003). This method adjusts parameters in a one-dimensional, variably saturated heat-transport model (VS2DH) [developed in the USGS (Healy and Ronan, 1996)] until the simulation results match the temperature data collected beneath the surface-water body. Sediment temperature measurements and data also have been used to determine streamflow frequency in ephemeral stream channels (Constantz and others, 2001; Stonestrom and Constantz, 2003). This method is described in greater detail in Chapter 4. Conant (2004) used measurements of hydraulic-head gradient and hydraulic conductivity in wells installed in a streambed to determine rates of exchange between ground water and surface water. Conant then developed an empirical relation between streambed flux and streambed temperature relative to stream-water temperature

and used the empirical relation to indicate rate of discharge of ground water to the stream at locations where only streambed temperature was measured.

Use of Specific-Conductance Probes

Specific-conductance probes are another tool that can be useful for locating areas of ground-water discharge to surface waters (Lee, 1985; Vanek and Lee, 1991; Harvey and others, 1997) (fig. 19). Such probes are suspended from a boat with a cable connecting the probes to a specific-conductance meter on board the boat. The housing for the electrically conductive probes is designed to maintain contact with the sediments (Lee, 1985) so that the probes are dragged through bottom sediments at a depth of 1 to 3 centimeters (Vanek and Lee, 1991). This method depends on having the existence of a difference in the electrical conductance of surface and ground water great enough to be detected by the sensors (probes) and thus to identify points or areas of ground-water discharge to the surface water. Thus, saline waters receiving ground-water discharge of less salinity (Vanek and Lee, 1991) or fresh surface waters receiving more mineralized ground-water discharge (Harvey and others, 1997) are environments where this technique is effective. Changes in electrical conductance, however, also may reflect changes in sediment type so that the technique should be considered only as a reconnaissance tool. The identification of places where ground-water discharge may be occurring should be verified by other, complementary techniques (Vanek and Lee, 1991). Specific-conductance and temperature probes have been used in this way along many kilometers of river reaches in combination with Global Positioning System information to determine the precise locations of ground-water discharge and their variation with season (Vaccaro and Maloy, 2006).

Figure 19. Example of the use of a towed specific-conductance probe to identify ground-water discharge to surface waters (from Harvey and others, 1997, used in accordance with usage permissions of the American Geophysical Union). The sediment probe is being towed behind a small boat. The probe is used to detect areas of more electrically conductive ground-water inflow. In saline waters, the probe is used to detect less electrically conductive fresh ground-water inflow.

Electrical Resistivity Profiling

Electrical geophysical survey methods are applicable to delineating submarine ground-water discharge (Loke and Lane, 2004). Salinity differences in water can be detected by the resistance to the passage of current between electrodes. Fresh or weakly saline waters will be more resistant to movement of an electric current than more saline waters. Two resistivity data acquisition geometries commonly are used: (1) continuous resistivity profiling (Belaval and others, 2003; Manheim and others, 2004; Day-Lewis and others, 2006), and (2) marine resistivity (Taniguchi and others, 2006). In continuous resistivity profiling, a streamer comprising a set of floating electrodes spaced at a regular interval is towed on the water surface behind a boat. As the boat travels along a transect, electrical current is applied at a fixed time interval at one or more electrode pairs, and electrical potentials are measured simultaneously between other electrode pairs. At the same time, water depths along the transect are measured with echo sounding. In marine resistivity profiling, measurements are made using electrodes placed on the water bottom. Regardless of the acquisition geometry, data are inverted to produce two-dimensional cross sections, or tomograms, of subsurface resistivity. Such data can indicate locations within a transect where submarine ground-water discharge is occurring, as well as delineating the subsurface saltwater/freshwater interface or geologic structure. Tomograms are commonly interpreted in the context of direct, discrete measurements of conductivity, temperature, and depth.

Hydraulic Potentiomanometer (Portable Wells) Measurements

Many devices have been designed to be installed through the bed of a surface-water body to measure the vertical hydraulic-head gradient beneath the surface-water body. These devices provide a direct measurement of hydraulic head relative to surface-water stage at the depth to which the probe is inserted beneath the surface-water bed. One of the most commonly used devices of this type is the hydraulic potentiomanometer (fig. 20), also sometimes called a minipiezometer

A

B
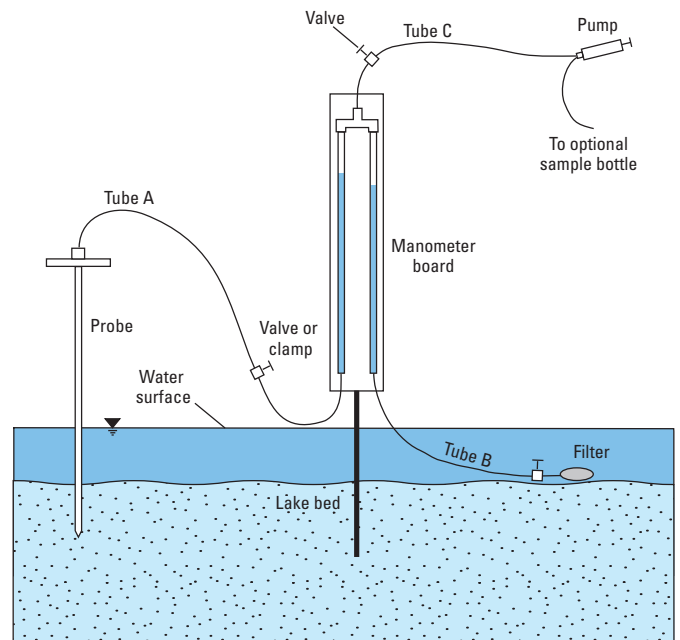
Figure 20. *A*, Hydraulic potentiomanometer with users demonstrating hydraulic-head difference between surface water and ground water. The hydraulic head of the river (right side of the board) is higher than the hydraulic head of the ground water (left side of the board), indicating a downward gradient (from Simonds and Sinclair, 2002). Photograph by Kirk A. Sinclair, Washington State Department of Ecology, 2002. *B*, Components of a hydraulic-potentiomanometer system (from Winter and others, 1988, copyright 1988 by the American Society of Limnology and Oceanography, Inc., used with permission).

(Winter and others, 1988). Although this device does not provide direct measurements of flow across the sediment-water interface, it is useful for making qualitative determinations of direction of flow across the sediment-water interface. Estimates of hydraulic conductivity also can be made on the basis of the amount of suction required to pull water through the screen at the end of the probe. These devices often are used in conjunction with a network of wells to obtain additional hydraulic-head data along shoreline segments where wells are not or cannot be situated. Details about use of these devices is provided in Chapter 2.

Seepage Meter Measurements

Seepage meters are devices that isolate a small area of the bed of a surface-water body and measure the flow of water across that area. References as early as 1944 indicate that early seepage meters were designed to measure water loss through unlined canals (Carr and Winter, 1980). Beginning in the early 1970s, seepage meters have been used in lakes, rivers, wetlands, and estuaries to measure flows between surface water and ground water in natural settings. Use in coastal environments has increased during recent years (Cable and others, 1997). One of the most common devices, called a half-barrel seepage meter, uses a cut-off end of a steel (or plastic) storage drum to isolate a small circular area of the surface-water bed, and a plastic bag is attached to the barrel to register the change in water volume over the time of bag attachment (Lee, 1977; Lee and Cherry, 1978) (fig. 21). More details about the use and interpretation of data from seepage meters are included in Chapter 2. Many device modifications have been made to the basic design of seepage meters for use in deep water, soft sediments, shallow water, and areas exposed to large waves.

Many investigations of surface-water and ground-water interaction require integrating point measurements of water flux in order to interpret the total flow between surface water and ground water. The most common method is to average point measurements and apply that average value to all or part of the surface-water body of interest. This method is not appropriate, however, where ground-water flux into a surface water declines with greater distance from shore. If data are collected along transects perpendicular to the shore, a curve can be fit through the data. Because ground-water flow to surface water commonly is distributed exponentially with distance from shore, exponential curves have been fit through transect data, and the equation for the curve has been used to calculate water flux for an area deemed to be representative of the transect point measurements (for example, Fellows and Brezonik, 1980). Where ground-water flux is not exponentially distributed with distance from shore, a plot of point measurements of seepage flux with distance from shore is made, and the area under the curve can be determined to represent total seepage for a unit width of shoreline. That value can
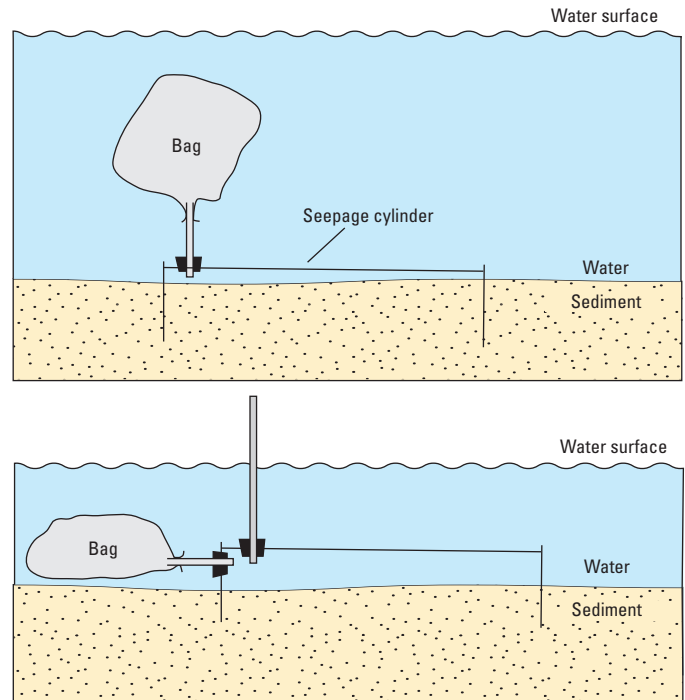
Figure 21.   Full-section view of seepage meter showing details of placement in the sediment (modified from Lee and Cherry, 1978, used with permission of the Journal of Geoscience Education).

then be multiplied by an appropriate shoreline length to determine a flux volume for a specific area of the surface-water body. If point measurements are made at a sufficient number of transects, ground-water flux for an area of a surface-water body can be determined.

Biological Indicators

The biological response to conditions of flow at the sediment-water interface can be an indicator of the direction and relative magnitude of flow. The growing field of ground-water ecology has made frequent use of distributions of specific types of plants and animals as an indicator of ground-water/surface-water interaction (Danielopol, 1984; Danielopol and others, 1997; Lodge and others, 1989; Malard and others, 1996; Goslee and others, 1997; Wetzel, 1999) (fig. 22). These methods are useful reconnaissance tools to aid in locating areas in need of more detailed investigations. Typically, these methods involve identifying species or groups of species of plants or animals that are known to thrive in places where ground water discharges to surface water, but some of the indicators also indicate areas where surface water flows into ground water. Identification of specific plant and animal species is necessary for use of these methods, but some of the species are readily identifiable without requiring intensive biological or ecological training (Rosenberry and others, 2000).

Figure 22.   Marsh marigold in Shingobee Lake, Minnesota. Presence indicates location of ground-water discharge to the lake. (Photographs by Donald O. Rosenberry, U.S. Geological Survey.)

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