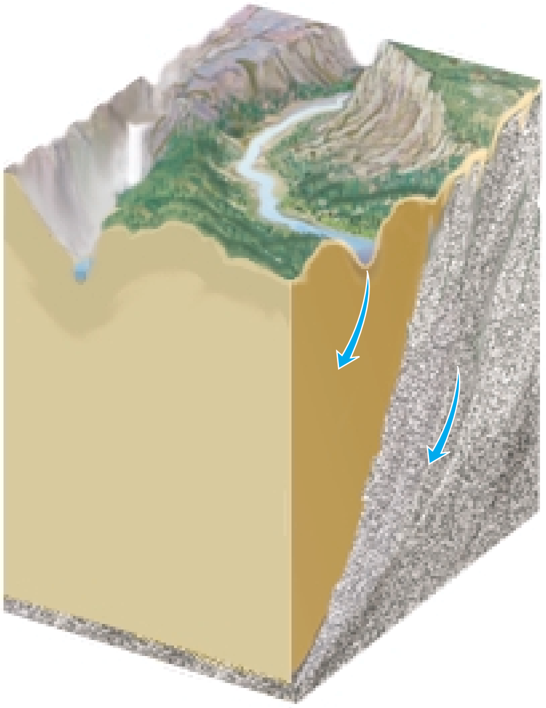
Interaction of Ground Water and Surface Water in Different Landscapes

Ground water is present in virtually all landscapes. The interaction of ground water with surface water depends on the physiographic and climatic setting of the landscape. For example, a stream in a wet climate might receive ground-water inflow, but a stream in an identical physiographic setting in an arid climate might lose water to ground water. To provide a broad and unified perspective of the interaction of ground water and surface water in different landscapes, a conceptual landscape (Figure 2) is used as a reference. Some common features of the interaction for various parts of the conceptual landscape are described below. The five general types of terrain discussed are mountainous, riverine, coastal, glacial and dune, and karst.

MOUNTAINOUS TERRAIN

The hydrology of mountainous terrain (area M of the conceptual landscape, Figure 2) is characterized by highly variable precipitation and water movement over and through steep land slopes. On mountain slopes, macropores created by burrowing organisms and by decay of plant roots have the capacity to transmit subsurface flow downslope quickly. In addition, some rock types underlying soils may be highly weathered or fractured and may transmit significant additional amounts of flow through the subsurface. In some settings this rapid flow of water results in hillside springs.



A general concept of water flow in mountainous terrain includes several pathways by which precipitation moves through the hillside to a stream (Figure 20). Between storm and snowmelt periods, streamflow is sustained by discharge from the ground-water system (Figure 20*A*). During intense storms, most water reaches streams very rapidly by partially saturating and flowing through the highly conductive soils. On the lower parts of hillslopes, the water table sometimes rises to the land surface during storms, resulting in overland flow (Figure 20*B*). When this occurs, precipitation on the saturated area adds to the quantity of overland flow. When storms or snowmelt persist in mountainous areas, near-stream saturated areas can expand outward from streams to include areas higher on the hillslope. In some settings, especially in arid regions, overland flow can be generated when the rate of rainfall exceeds the infiltration capacity of the soil (Figure 20*C*).

Near the base of some mountainsides, the water table intersects the steep valley wall some distance up from the base of the slope (Figure 21, left side of valley). This results in perennial discharge of ground water and, in many cases, the presence of wetlands. A more common hydrologic process that results in the presence of wetlands in some mountain valleys is the upward discharge of ground water caused by the change in slope of the water table from being steep on the valley side to being relatively flat in the alluvial valley (Figure 21, right side of valley). Where both of these water-table conditions exist, wetlands fed by ground water, which commonly are referred to as fens, can be present.

Another dynamic aspect of the interaction of ground water and surface water in mountain settings is caused by the marked longitudinal component of flow in mountain valleys. The high gradient of mountain streams, coupled with the coarse texture of streambed sediments, results in a strong down-valley component of flow accompanied by frequent exchange of stream water with water in the hyporheic zone (Figure 14) (see Box H). The driving force for water exchange between a stream and its hyporheic zone is created by the surface water flowing over rough streambeds, through pools and riffles, over cascades, and around boulders and logs. Typically, the stream enters the hyporheic zone at the downstream end of pools and then flows beneath steep sections of the stream (called riffles), returning to the stream at the upstream end of the next pool (Figure 14*A*). Stream water also may enter the hyporheic zone upstream from channel meanders, causing stream water to flow through a gravel bar before reentering the channel downstream (Figure 14*B*).

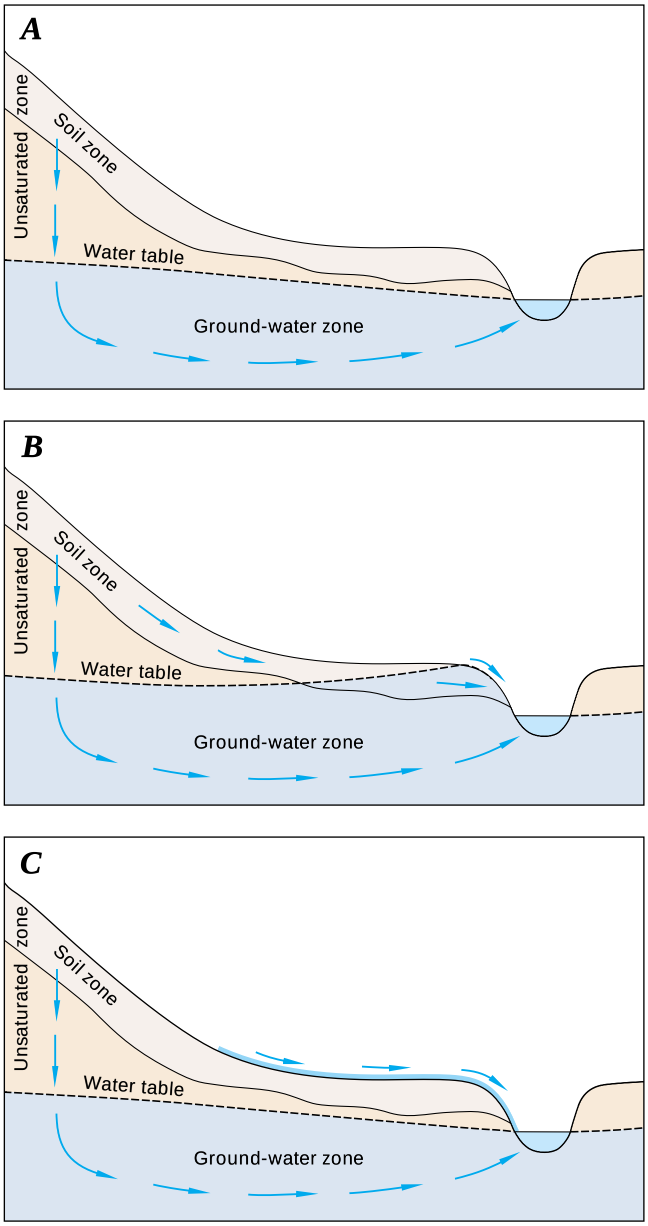


Figure ***20.*** *Water from precipitation moves to mountain streams along several pathways. Between storms and snowmelt periods, most inflow to streams commonly is from ground water (A). During storms and snowmelt periods, much of the water inflow to streams is from shallow flow in saturated macropores in the soil zone. If infiltration to the water table is large enough, the water table will rise to the land surface and flow to the stream is from ground water, soil water, and overland runoff (B). In arid areas where soils are very dry and plants are sparse, infiltration is impeded and runoff from precipitation can occur as overland flow (C). (Modified from Dunne, T., and Leopold, L.B., 1978, Water in environmental planning: San Francisco, W.H. Freeman.) (Used with permission.)*

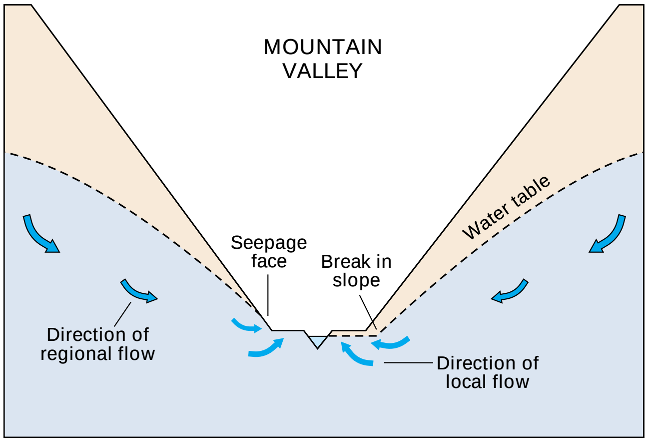


Figure ***21****. In mountainous terrain, ground water can discharge at the base of steep slopes (left side of valley), at the edges of flood plains (right side of valley), and to the stream.*

Streams flowing from mountainous terrain commonly flow across alluvial fans at the edges of the valleys. Most streams in this type of setting lose water to ground water as they traverse the highly permeable alluvial fans. This process has long been recognized in arid western regions, but it also has been documented in humid regions, such as the Appalachian Mountains. In arid and semiarid regions, seepage of water from the stream can be the principal source of aquifer recharge. Despite its importance, ground-water recharge from losing streams remains a highly uncertain part of the water balance of aquifers in these regions. Promising new methods of estimating ground-water recharge, at least locally, along mountain fronts are being developed—these methods include use of environmental tracers, measuring vertical temperature profiles in streambeds, measuring hydraulic characteristics of streambeds, and measuring the difference in hydraulic head between the stream and the underlying aquifer.

The most common natural lakes in mountainous terrain are those that are dammed by rock sills or glacial deposits high in the mountains. Termed cirque lakes, they receive much of their water from snowmelt. However, they interact with ground water much like the processes shown in Figure 21, and they can be maintained by ground water throughout the snow-free season.

The geochemical environment of mountains is quite diverse because of the effects of highly variable climate and many different rock and soil types on the evolution of water chemistry. Geologic materials can include crystalline, volcanic, and sedimentary rocks and glacial deposits. Sediments can vary from those having well-developed soil horizons to stream alluvium that has no soil development. During heavy precipitation, much water flows through shallow flow paths, where it interacts with microbes and soil gases. In the deeper flow through fractured bedrock, longer term geochemical interactions of ground water with minerals determine the chemistry of water that eventually discharges to streams. Base flow of streams in mountainous terrain is derived by drainage from saturated alluvium in valley bottoms and from drainage of bedrock fractures. Mixing of these chemically different water types results in geochemical reactions that affect the chemistry of water in streams. During downstream transport in the channel, stream water mixes with ground water in the hyporheic zone. In some mountain streams, the volume of water in the hyporheic zone is considerably larger than that in the stream channel. Chemical reactions in hyporheic zones can, in some cases, substantially alter the water chemistry of streams (Figure 19).

H

Field Studies of Mountainous Terrain

The steep slopes and rocky characteristics of mountainous terrain make it difficult to determine interactions of ground water and surface water. Consequently, few detailed hydrogeologic investigations of these interactions have been conducted in mountainous areas. Two examples are given below.

A field and modeling study of the Mirror Lake area in the White Mountains of New Hampshire indicated that the sizes of ground-water flow systems contributing to surface-water bodies were considerably larger than their topographically defined watersheds. For example, much of the ground water in the fractured bedrock that discharges to Mirror Lake passes beneath the local flow system associated with Norris Brook (Figure H–1). Furthermore, a more extensive deep ground-water flow system that discharges to the Pemigewasset River passes beneath flow systems associated with both Norris Brook and Mirror Lake.

Studies in mountainous terrain have used tracers to determine sources of ground water to streams (see Box G). In addition to revealing processes of water exchange between ground water and stream water, solute tracers have proven useful for defining the limits of the hyporheic zone surrounding mountain streams. For example, solute tracers such as chloride or bromide ions are injected into the stream to artificially raise concentrations above natural background concentrations. The locations and amounts of ground-water inflow are determined from a simple dilution model. The extent that tracers move into the hyporheic zone can be estimated by the models and commonly is verified by sampling wells placed in the study area.

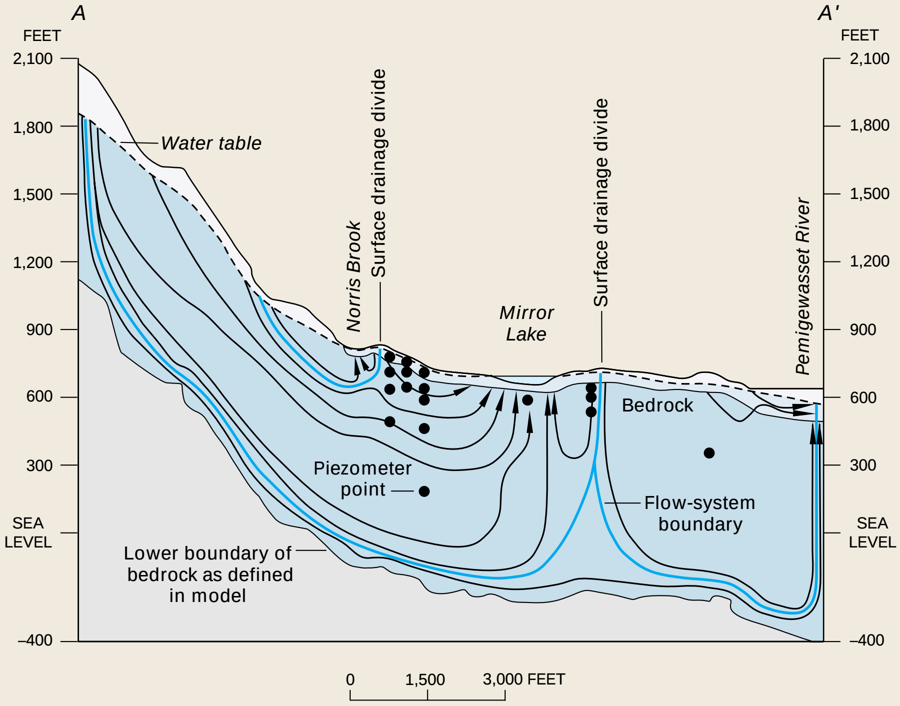


Figure H–1. Ground-water flow systems in the Mirror Lake area extend beyond the topographically defined surface-water watersheds. (Modified from Harte, P.T., and Winter, T.C., 1996, Factors affecting recharge to crystalline rock in the Mirror Lake area, Grafton County, New Hampshire: in Morganwalp, D.W., and Aronson, D.A., eds., U.S. Geological Survey Toxic Substances Hydrology Program—Proceedings of Technical Meeting, Colorado Springs, Colorado, September 20–24, 1993: U.S. Geological Survey Water-Resources Investigations Report 94–4014, p. 141–150.)

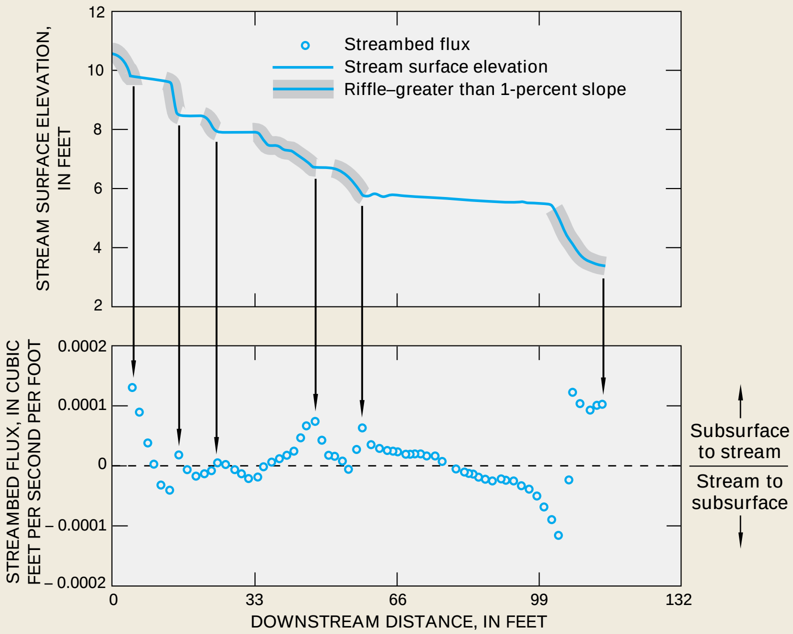


Figure H–2. In mountain streams characterized by pools and riffles, such as at Saint Kevin Gulch in Colorado, inflow of water from the hyporheic zone to the stream was greatest at the downstream end of riffles. (Modified from Harvey, J.W., and Bencala, K.E., 1993, The effect of streambed topography on surface-subsurface water exchange in mountain catchments: Water Resources Research, v. 29, p. 89–98.)

A study in Colorado indicated that hyporheic exchange in mountain streams is caused to a large extent by the irregular topography of the streambed, which creates pools and riffles characteristic of mountain streams. Ground water enters streams most readily at the upstream end of deep pools, and stream water flows into the subsurface beneath and to the side of steep sections of streams (riffles) (Figure H–2). Channel irregularity, therefore, is an important control on the location of ground-water inflow to streams and on the size of the hyporheic zone in mountain streams because changes in slope determine the length and depth of hyporheic flow paths.

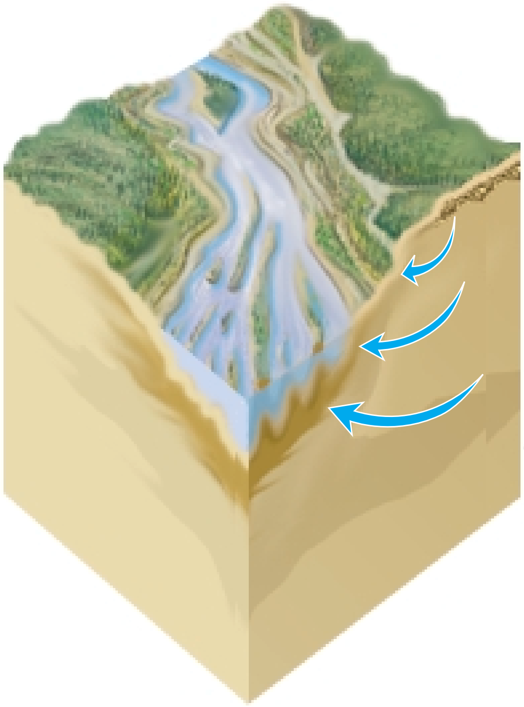
The source and fate of metal contaminants in streams receiving drainage from abandoned mines can be determined by using solute tracers. In addition to surface drainage from mines, a recent study of Chalk Creek in Colorado indicated that contaminants were being brought to the stream by ground-water inflow. The ground water had been contaminated from mining activities in the past and is now a new source of contamination to the stream. This nonpoint ground-water source of contamination will very likely be much more difficult to clean up than the point source of contamination from the mine tunnel.

RIVERINE TERRAIN

In some landscapes, stream valleys are small and they commonly do not have well-developed flood plains (area R of the conceptual landscape, Figure 2) (see Box I). However, major rivers (area V of the reference landscape, Figure 2) have valleys that usually become increasingly wider downstream. Terraces, natural levees, and abandoned river meanders are common landscape features in major river valleys, and wetlands and lakes commonly are associated with these features.

The interaction of ground water and surface water in river valleys is affected by the interchange of local and regional ground-water flow systems with the rivers and by flooding and evapotranspiration. Small streams receive ground-water inflow primarily from local flow systems, which usually have limited extent and are highly variable seasonally. Therefore, it is not unusual for small streams to have gaining or losing reaches that change seasonally.

For larger rivers that flow in alluvial valleys, the interaction of ground water and surface water usually is more spatially diverse than it is for smaller streams. Ground water from regional flow systems discharges to the river as well as at various places across the flood plain (Figure 22). If terraces are present in the alluvial valley, local ground-water flow systems may be associated with each terrace, and lakes and wetlands may be formed because of this source of ground water. At some locations, such as at the valley wall and at the river, local and regional ground-water flow systems may discharge in close proximity. Furthermore, in large alluvial valleys, significant down-valley components of flow in the streambed and in the shallow alluvium also may be present (see Box I).



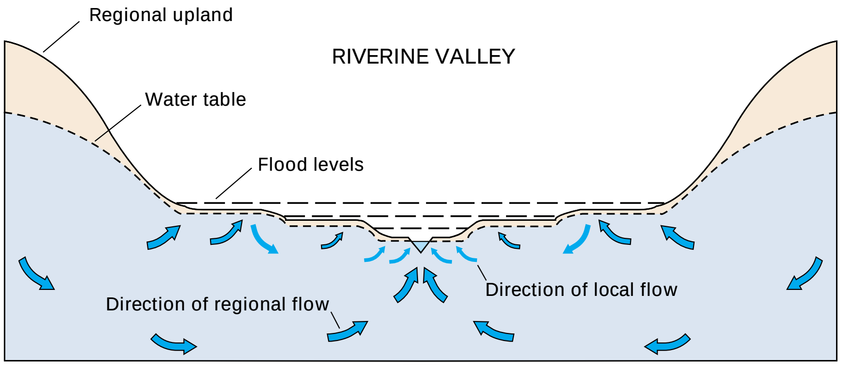


Figure 22. In broad river valleys, small local ground-water flow systems associated with terraces overlie more regional ground-water flow systems. Recharge from flood waters superimposed on these ground-water flow systems further complicates the hydrology of river

Added to this distribution of ground-water discharge from different flow systems to different parts of the valley is the effect of flooding. At times of high river flows, water moves into the ground-water system as bank storage (Figure 11). The flow paths can be as lateral flow through the riverbank (Figure 12*B*) or, during flooding, as vertical seepage over the flood plain (Figure 12*C*). As flood waters rise, they cause bank storage to move into higher and higher terraces.

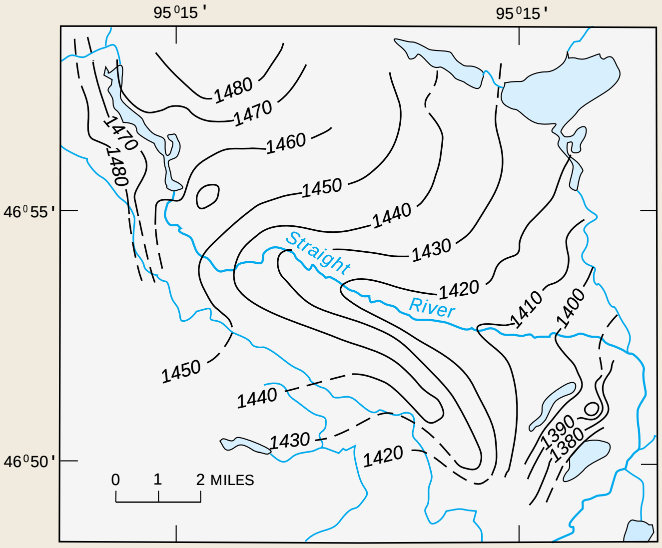
The water table generally is not far below the land surface in alluvial valleys. Therefore, vegetation on flood plains, as well as at the base of some terraces, commonly has root systems deep enough so that the plants can transpire water directly from ground water. Because of the relatively stable source of ground water, particularly in areas of ground-water discharge, the vegetation can transpire water near the maximum potential transpiration rate, resulting in the same effect as if the water were being pumped by a well (see Figure 7). This large loss of water can result in drawdown of the water table such that the plants intercept some of the water that would otherwise flow to the river, wetland, or lake. Furthermore, in some settings it is not uncommon during the growing season for the pumping effect of transpiration to be significant enough that surface water moves into the subsurface to replenish the transpired ground water.

Riverine alluvial deposits range in size from clay to boulders, but in many alluvial valleys, sand and gravel are the predominant deposits. Chemical reactions involving dissolution or precipitation of minerals (see Box D) commonly do not have a significant effect on water chemistry in sand and gravel alluvial aquifers because the rate of water movement is relatively fast compared to weathering rates. Instead, sorption and desorption reactions and oxidation/reduction reactions related to the activity of microorganisms probably have a greater effect on water chemistry in these systems. As in small streams, biogeochemical processes in the hyporheic zone may have a significant effect on the chemistry of ground water and surface water in larger riverine systems. Movement of oxygen-rich surface water into the subsurface, where chemically reactive sediment coatings are abundant, causes increased chemical reactions related to activity of microorganisms. Sharp gradients in concentration of some chemical constituents in water, which delimit this zone of increased biogeochemical activity, are common near the boundary between ground water and surface water. In addition, chemical reactions in the hyporheic zone can cause precipitation of some reactive solutes and contaminants, thereby affecting water quality.

Field Studies of Riverine Terrain

Streams are present in virtually all landscapes, and in some landscapes, they are the principal surface-water features. The interaction of ground water with streams varies in complexity because they vary in size from small streams near headwaters areas to large rivers flowing in large alluvial valleys, and also because streams intersect ground-water flow systems of greatly different scales. Examples of the interaction of ground water and surface water for small and large riverine systems are presented below.

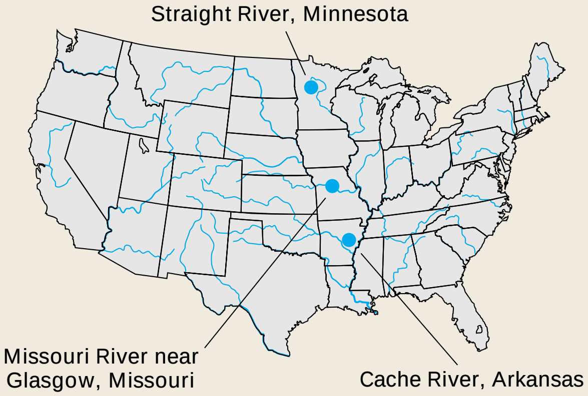
The Straight River, which runs through a sand plain in central Minnesota, is typical of a small stream that does not have a flood plain and that derives most of its water from ground-water inflow. The water-table contours near the river bend sharply upstream (Figure I–1), indicating that ground water moves directly into the river. It is estimated from base-flow studies (see Box B) that, on an annual basis, ground water accounts for more than 90 percent of the water in the river.



EXPLANATION

—*1420*— WATER-TABLE CONTOUR—Shows altitude of the water table in feet above sea level. Dashed where approximately located. Contour interval 10 feet

Figure I–1. Small streams, such as the Straight River in Minnesota, commonly do not have flood plains. The flow of ground water directly into the river is indicated by the water-table contours that bend sharply upstream. (Modified from Stark, J.R., Armstrong, D.S., and Zwilling, D.R., 1994, Stream-aquifer interactions in the Straight River area, Becker and Hubbard Counties, Minnesota: U.S. Geological Survey Water-Resources Investigations Report 94–4009, 83 p.)



In contrast, the results of a study of the lower Missouri River Valley indicate the complexity of ground-water flow and its interaction with streams in large alluvial valleys. Configuration of the water table in this area indicates that ground water flows into the river at right angles in some reaches, and it flows parallel to the river in others (Figure I–2*A*). This study also resulted in a map that showed patterns of water-table fluctuations with respect to proximity to the river (Figure I–2*B*). This example shows the wide variety of ground-water flow conditions that can be present in large alluvial valleys.

Another study of part of a large alluvial valley provides an example of the presence of smaller scale flow conditions. The Cache River is a stream within the alluvial valley of the Mississippi River Delta system in eastern Arkansas. In a study of the Black Swamp, which lies along a reach of the river, a number of wells and piezometers were installed to determine the interaction of ground water with the swamp and the river. By measuring hydraulic head at different depths in the

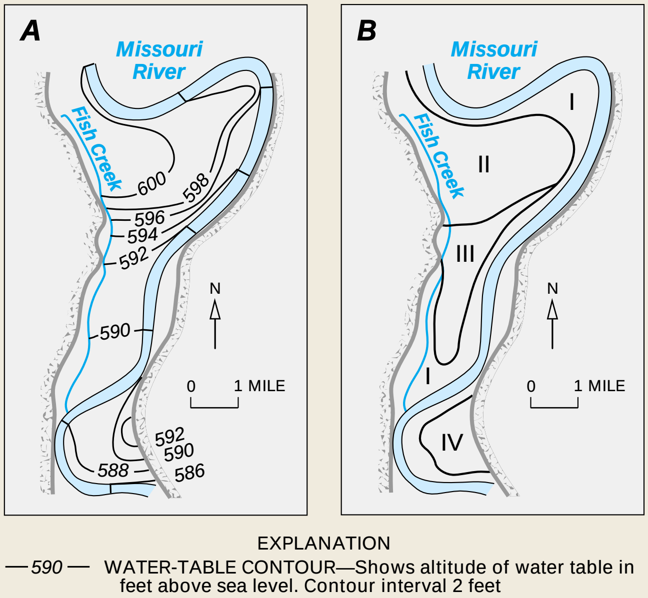


Figure I–2. In flood plains of large rivers, such as the Missouri River near Glasgow, Missouri, patterns of ground-water movement (A) and water-table fluctuations (B) can be complex. Zone I is an area of rapidly fluctuating water levels, zone II is an area of long-term stability, zone III is an area of down-valley flow, and zone IV is a persistent ground-water high. (Modified from Grannemann, N.G., and Sharp, J.M., Jr., 1979, Alluvial hydrogeology of the lower Missouri River: Journal of Hydrology, v. 40, p. 85–99.) (Reprinted with permission from Elsevier Science-NL, Amsterdam, The Netherlands.)

alluvium, it was possible to construct a hydrologic section through the alluvium (Figure I–3), showing that the river receives ground-water discharge from both local and regional ground-water flow systems. In addition, the section also shows the effect of the break in slope associated with the terrace at the edge of the swamp, which causes ground water from a local flow system to discharge into the edge of the swamp rather than to the river.

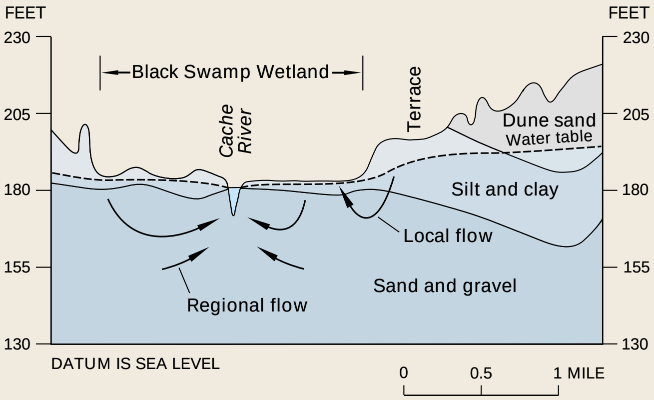


Figure I–3. The Cache River in Arkansas provides an example of contributions to a river from regional and local ground-water flow systems. In addition, a small local ground-water flow system associated with a terrace discharges to the wetland at the edge of the flood plain. (Modified from Gonthier, G.J., 1996, Ground-water flow conditions within a bottomland hardwood wetland, eastern Arkansas: Wetlands, v. 16, no. 3, p. 334–346.) (Used with permission.)

COASTAL TERRAIN

Coastal terrain, such as that along the east-central and southern coasts of the United States, extends from inland scarps and terraces to the ocean (area C of the conceptual landscape, Figure 2). This terrain is characterized by (1) low scarps and terraces that were formed when the ocean was higher than at present; (2) streams, estuaries, and lagoons that are affected by tides; (3) ponds that are commonly associated with coastal sand dunes; and (4) barrier islands. Wetlands cover extensive areas in some coastal terrains (see Figure 18).

The interaction of ground water and surface water in coastal terrain is affected by discharge of ground water from regional flow systems and from local flow systems associated with scarps and terraces (Figure 23), evapotranspiration, and tidal flooding. The local flow systems associated with scarps and terraces are caused by the configuration of the water table near these features (see Box J). Where the water table has a downward break in slope near the top of scarps and terraces, downward components of ground-water flow are present; where the water table has an upward break in slope near the base of these features, upward components of ground-water flow are present.

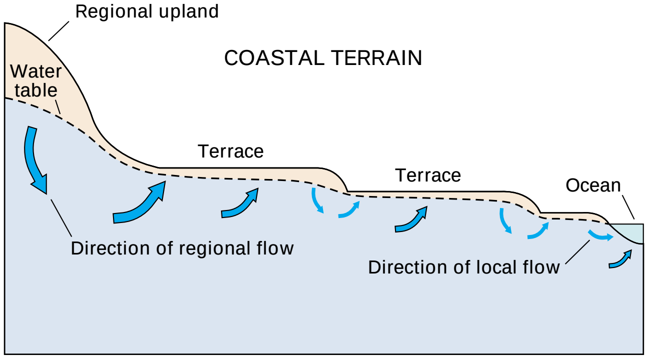
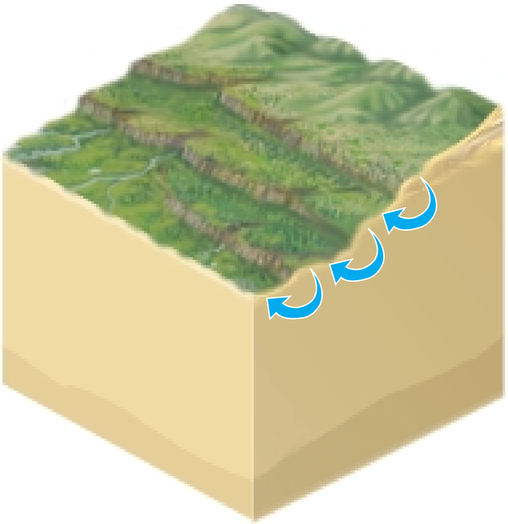


Figure 23. *In coastal terrain, small local ground-water flow cells associated with terraces overlie more regional ground-water flow systems. In the tidal zone, saline and brackish surface water mixes with fresh ground water from local and regional flow systems.*



Evapotranspiration directly from ground water is widespread in coastal terrain. The land surface is flat and the water table generally is close to land surface; therefore, many plants have root systems deep enough to transpire ground water at nearly the maximum potential rate. The result is that evapotranspiration causes a significant water loss, which affects the configuration of ground-water flow systems as well as how ground water interacts with surface water.

In the parts of coastal landscapes that are affected by tidal flooding, the interaction of ground water and surface water is similar to that in alluvial valleys affected by flooding. The principal difference between the two is that tidal flooding is more predictable in both timing and magnitude than river flooding. The other significant difference is in water chemistry. The water that moves into bank storage from rivers is generally fresh, but the water that moves into bank storage from tides generally is brackish or saline.

Estuaries are a highly dynamic interface between the continents and the ocean, where discharge of freshwater from large rivers mixes with saline water from the ocean. In addition, ground water discharges to estuaries and the ocean, delivering nutrients and contaminants directly to coastal waters. However, few estimates of the location and magnitude of ground-water discharge to coasts have been made.

In some estuaries, sulfate-rich regional ground water mixes with carbonate-rich local ground water and with chloride-rich seawater, creating sharp boundaries that separate plant and wildlife communities. Biological communities associated with these sharp boundaries are adapted to different hydrochemical conditions, and they undergo periodic stresses that result from inputs of water having different chemistry. The balance between river inflow and tides causes estuaries to retain much of the particulate and dissolved matter that is transported in surface and subsurface flows, including contaminants.

"Ground water discharges to estuaries and the ocean, delivering nutrients and contaminants directly to coastal waters"

J

Field Studies of Coastal Terrain

Along the Atlantic, Gulf of Mexico, and Arctic Coasts of the United States, broad coastal plains are transected by streams, scarps, and terraces. In some parts of these regions, local ground-water flow systems are associated with scarps and terraces, and freshwater wetlands commonly are present. Other parts of coastal regions are affected by tides, resulting in very complex flow and biogeochemical processes.

Underlying the broad coastal plain of the mid-Atlantic United States are sediments 600 or more feet thick. The sands and clays were deposited in stratigraphic layers that slope gently from west to east. Ground water moves regionally toward the east in the more permeable sand layers. These aquifers are separated by discontinuous layers of clay that restrict vertical ground-water movement. Near land surface, local ground-water flow systems are associated with changes in land slope, such as at major scarps and at streams.

Studies of the Dismal Swamp in Virginia and North Carolina provide examples of the interaction of ground water and wetlands near a coastal scarp. The Suffolk Scarp borders the west side of Great Dismal Swamp. Water-table wells and deeper piezometers placed across the scarp indicated a downward component of ground-water flow in the upland and an upward component of ground-water flow in the lowland at the edge of the swamp (Figure J–1*A*). However, at the edge of the swamp the direction of flow changed several times between May and October in 1982 because transpiration of ground water lowered the water table below the water level of the deep piezometer (Figure J–1*B*).

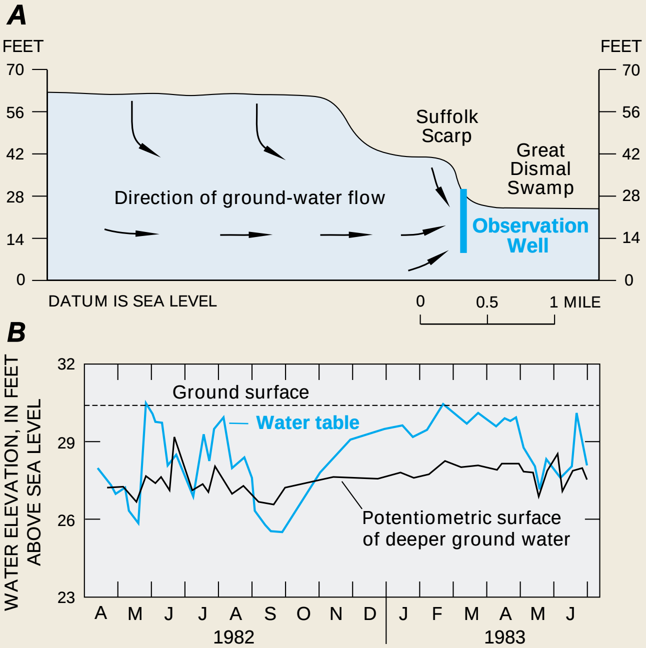
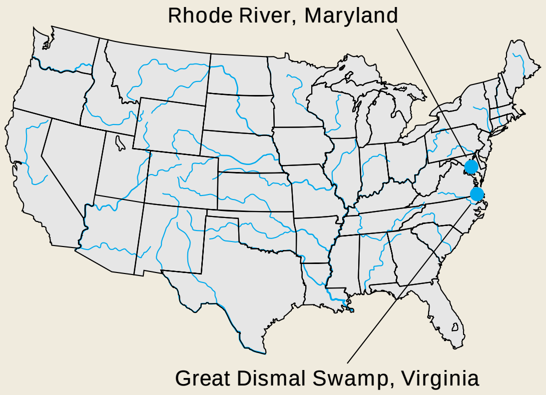


Figure J–1. Ground-water discharge at the edge of the Great Dismal Swamp in Virginia provides an example of local ground-water flow systems associated with coastal scarps (A). The vertical components of flow can change direction seasonally, partly because evapotranspiration discharges shallower ground water during part of the year (B). (Modified from Carter, Virginia, 1990, The Great Dismal Swamp—An illustrated case study, chapter 8, in Lugo, A.E., Brinson, Mark, and Brown, Sandra, eds., Ecosystems of the world, 15: Forested wetlands, Elsevier, Amsterdam, p. 201–211.) (Reprinted with permission from Elsevier Science-NL, Amsterdam, The Netherlands.)



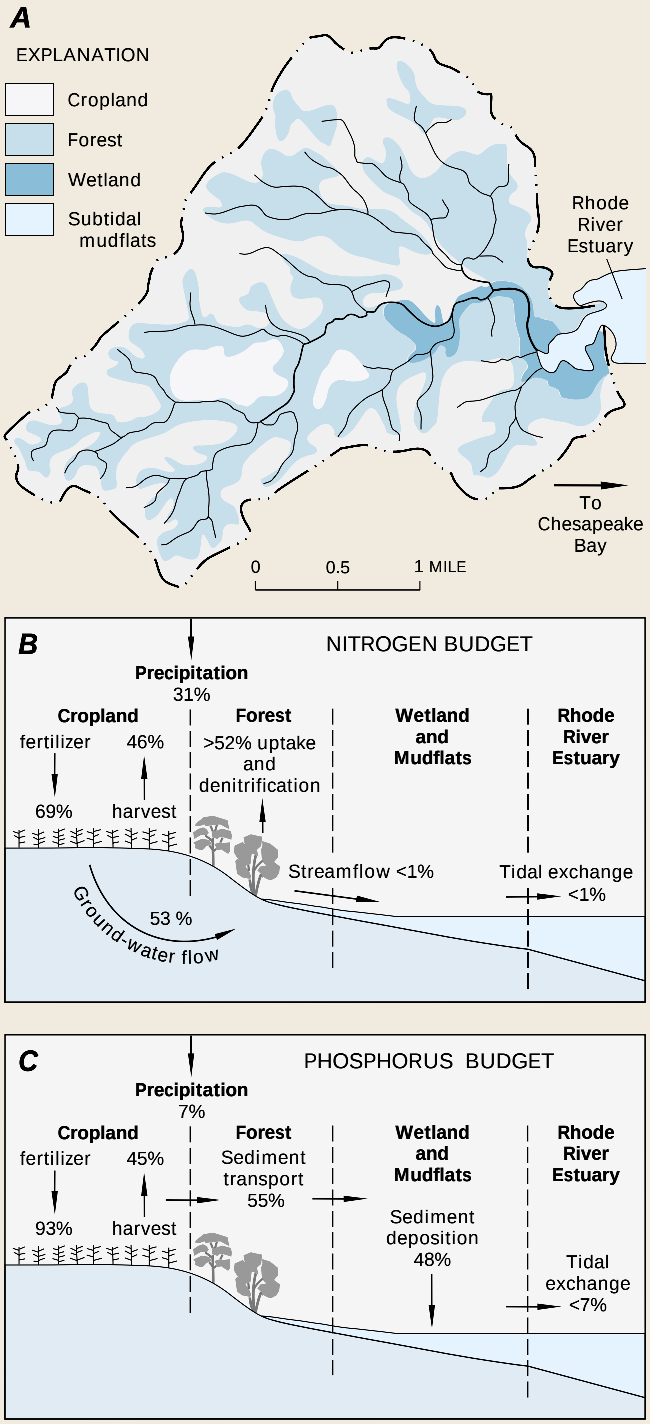


Figure J–2. Forests and wetlands separate cropland from streams in the Rhode River watershed in Maryland (A). More than half of the nitrogen applied to cropland is transported by ground water toward riparian forests and wetlands (B). More than half of the total phosphorus applied to cropland is transported by streams to wetlands and mudflats, where most is deposited in sediments (C). (Modified from Correll, D.L., Jordan, T.E., and Weller, D.E., 1992, Nutrient flux in a landscape—Effects of coastal land use and terrestrial community mosaic on nutrient transport to coastal waters: Estuaries, v. 15, no. 4, p. 431–442.) (Reprinted by permission of the Estuarine Research Federation.)

The gentle relief and sandy, well-drained soils of coastal terrain are ideal for agriculture. Movement of excess nutrients to estuaries are a particular problem in coastal areas because the slow rate of flushing of coastal bays and estuaries can cause them to retain nutrients. At high concentrations, nutrients can cause increased algal production, which results in overabundance of organic matter. This, in turn, can lead to reduction of dissolved oxygen in surface water to the extent that organisms are killed throughout large areas of estuaries and coastal bays.

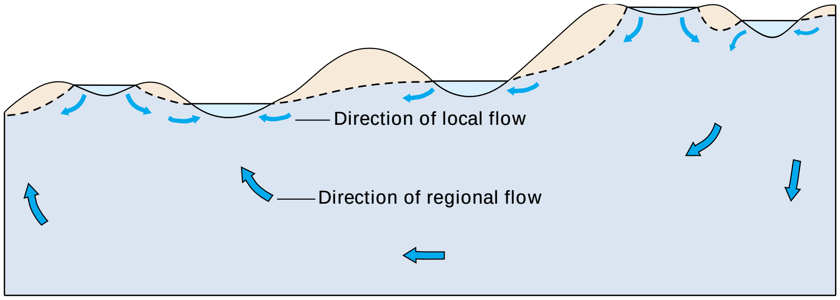
Movement of nutrients from agricultural fields has been documented for the Rhode River watershed in Maryland (Figure J–2). Application of fertilizer accounts for 69 percent of nitrogen and 93 percent of phosphorus input to this watershed (Figure J–2*B* and J–2*C*). Almost all of the nitrogen that is not removed by harvested crops is transported in ground water and is taken up by trees in riparian forests and wetlands or is denitrified to nitrogen gas in ground water before it reaches streams. On the other hand, most of the phosphorus not removed by harvested crops is attached to soil particles and is transported only during heavy precipitation when sediment from fields is transported into streams and deposited in wetlands and subtidal mudflats at the head of the Rhode River estuary. Whether phosphorus is retained in sediments or is released to the water column depends in part on whether sediments are exposed to oxygen. Thus, the uptake of nutrients and their storage in riparian forests, wetlands, and subtidal mudflats in the Rhode River watershed has helped maintain relatively good water quality in the Rhode River estuary.

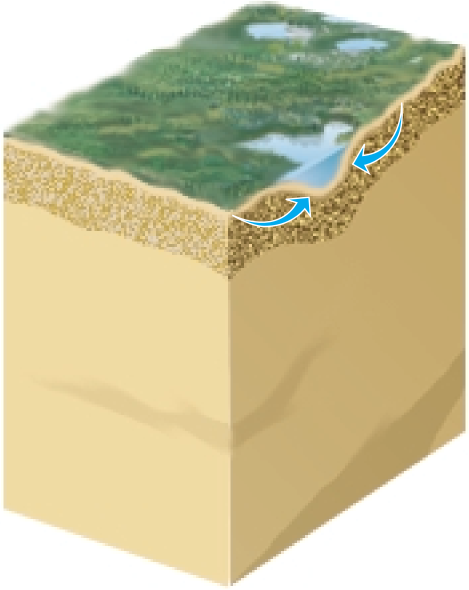
In other areas, however, agricultural runoff and input of nutrients have overwhelmed coastal systems, such as in the northern Gulf of Mexico near the mouth of the Mississippi River. The 1993 flood in the Mississippi River system delivered an enormous amount of nutrients to the Gulf of Mexico. Following the flood, oxygen-deficient sediments created areas of black sediment devoid of animal life in parts of the northern Gulf of Mexico.

GLACIAL AND DUNE TERRAIN

Glacial and dune terrain (area G of the conceptual landscape, Figure 2) is characterized by a landscape of hills and depressions. Although stream networks drain parts of these landscapes, many areas of glacial and dune terrain do not contribute runoff to an integrated surface drainage network. Instead, surface runoff from precipitation falling on the landscape accumulates in the depressions, commonly resulting in the presence of lakes and wetlands. Because of the lack of stream outlets, the water balance of these “closed” types of lakes and wetlands is controlled largely by exchange of water with the atmosphere (precipitation and evapotranspiration) and with ground water (see Box K).

Figure 24. *In glacial and dune terrain, local, intermediate, and regional ground-water flow systems interact with lakes and wetlands. It is not uncommon for wetlands that recharge local ground-water flow systems to be present in lowlands and for wetlands that receive discharge from local ground water to be present in uplands.*





Lakes and wetlands in glacial and dune terrain can have inflow from ground water, outflow to ground water, or both (Figure 16). The interaction between lakes and wetlands and ground water is determined to a large extent by their position with respect to local and regional ground-water flow systems. A common conception is that lakes and wetlands that are present in topographically high areas recharge ground water, and that lakes and wetlands that are present in low areas receive discharge from ground water. However, lakes and wetlands underlain by deposits having low permeability can receive discharge from local ground-water flow systems even if they are located in a regional ground-water recharge area. Conversely, they can lose water to local ground-water flow systems even if they are located in a regional ground-water discharge area (Figure 24).

Lakes and wetlands in glacial and dune terrain underlain by highly permeable deposits commonly have ground-water seepage into one side and seepage to ground water on the other side. This relation is relatively stable because the water-table gradient between surface-water bodies in this type of setting is relatively constant. However, the boundary between inflow to the lake or wetland and outflow from it, termed the hinge line, can move up and down along the shoreline. Movement of the hinge line between inflow and outflow is a result of the changing slope of the water table in response to changes in ground-water recharge in the adjacent uplands.

Transpiration directly from ground water has a significant effect on the interaction of lakes and wetlands with ground water in glacial and dune terrain. Transpiration from ground water (Figure 7) has perhaps a greater effect on lakes and wetlands underlain by low-permeability deposits than in any other landscape. The lateral movement of ground water in low-permeability deposits may not be fast enough to supply the quantity of water at the rate it is removed by transpiration, resulting in deep and steep-sided cones of depression. These cones of depression commonly are present around the perimeter of the lakes and wetlands (Figure 7 and Box K).

In the north-central United States, cycles in the balance between precipitation and evapotranspiration that range from 5 to 30 years can result in large changes in water levels, chemical concentrations, and major-ion water type of individual wetlands. In some settings, repeated cycling of water between the surface and subsurface in the same locale results in evaporative concentration of solutes and eventually in mineral precipitation in the subsurface. In addition, these dynamic hydrological and chemical conditions can cause significant changes in the types, number, and distribution of wetland plants and invertebrate animals within wetlands. These changing hydrological conditions that range from seasons to decades are an essential process for rejuvenating wetlands that provide ideal habitat and feeding conditions for migratory waterfowl.

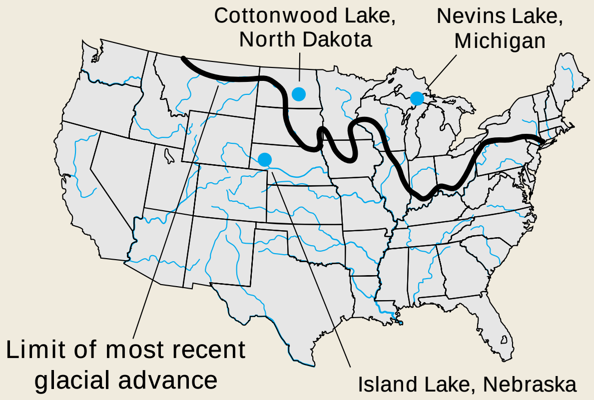
"The hydrological and chemical characteristics of lakes and wetlands in glacial and dune terrain are determined to a large extent by their position with respect to local and regional ground-water flow systems"

K

Field Studies of Glacial and Dune Terrain

Glacial terrain and dune terrain are characterized by land-surface depressions, many of which contain lakes and wetlands. Although much of the glacial terrain covering the north-central United States (see index map) has low topographic relief, neighboring lakes and wetlands are present at a sufficiently wide range of altitudes to result in many variations in how they interact with ground water, as evidenced by the following examples.

The Cottonwood Lake area, near Jamestown, North Dakota, is within the prairie-pothole region of North America. The hydrologic functions of these small depressional wetlands are highly variable in space and time. With respect to spatial variation, some wetlands recharge ground water, some receive ground-water inflow and have outflow to ground water, and some receive ground-water discharge. Wetland P1 provides an example of how their functions can vary in time. The wetland receives ground-water discharge most of the time; however, transpiration of ground water by plants around the perimeter of the wetland can cause water to seep from the wetland. Seepage from wetlands commonly is assumed to be ground-water recharge, but in cases like Wetland P1, the water is actually lost to transpiration. This process results in depressions in the water table around the perimeter of the wetland at certain times, as shown in



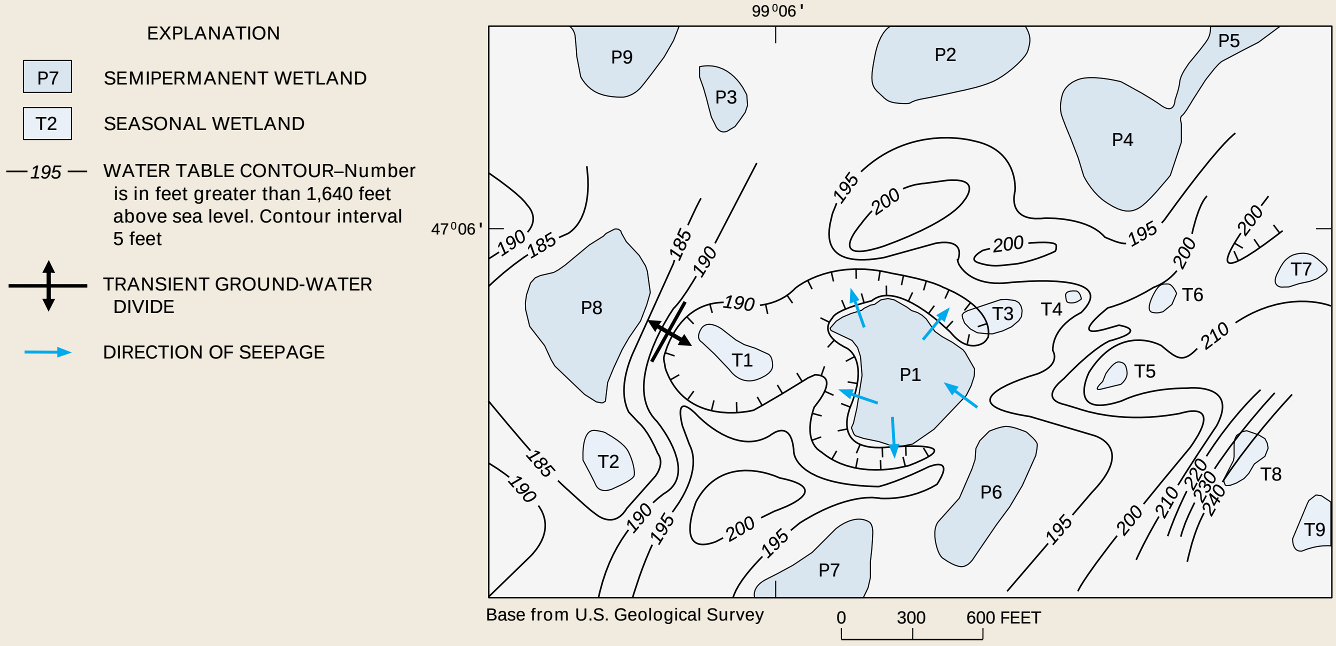


Figure K–1. Transpiration directly from ground water causes cones of depression to form by late summer around the perimeter of prairie pothole Wetland P1 in the Cottonwood Lake area in North Dakota. (Modified from Winter, T.C., and Rosenberry, D.O., 1995, The interaction of ground water with prairie pothole wetlands in the Cottonwood Lake area, east-central North Dakota, 1979–1990: Wetlands, v. 15, no. 3, p. 193–211.) (Used with permission.)

Figure K–1. Transpiration-induced depressions in the water table commonly are filled in by recharge during the following spring, but then form again to some extent by late summer nearly every year.

Nevins Lake, a closed lake in the Upper Peninsula of Michigan, illustrates yet another type of interaction of lakes with ground water in glacial terrain. Water-chemistry studies of Nevins Lake indicated that solutes such as calcium provide an indicator of ground-water inflow to the lake. Immediately following spring snowmelt, the mass of dissolved calcium in the lake increased rapidly because of increased ground-water inflow. Calcium then decreased steadily throughout the summer and early fall as the lake received less ground-water inflow (Figure K–2). This pattern varied annually depending on the amount of ground-water recharge from snowmelt and spring rains. The chemistry of water in the pores of the lake sediments was used to determine the spatial variability in the direction of seepage on the side of the lake that had the most ground-water inflow. Seepage was always out of the lake at the sampling site farthest from shore and was always upward into the lake at the site nearest to shore. Flow reversals were documented at sites located at intermediate distances from shore.

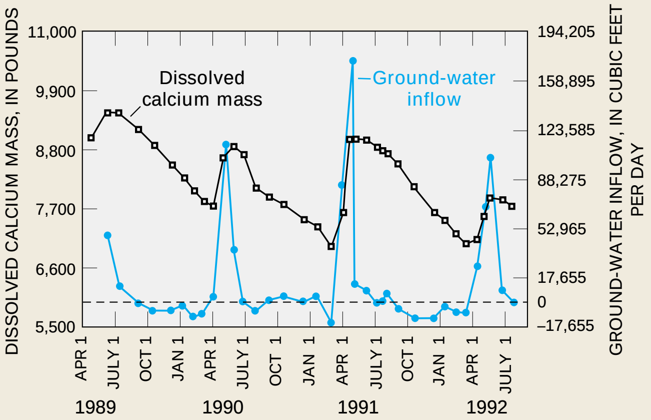


Figure K–2. A large input of ground water during spring supplies the annual input of calcium to Nevins Lake in the Upper Peninsula of Michigan. (Modified from Krabbenhoft, D.P., and Webster, K.E., 1995, Transient hydrogeological controls on the chemistry of a seepage lake: Water Resources Research, v. 31, no. 9, p. 2295–2305.)

Dune terrain also commonly contains lakes and wetlands. Much of the central part of western Nebraska, for example, is covered by sand dunes that have lakes and wetlands in most of the lowlands between the dunes. Studies of the interaction of lakes and wetlands with ground water at the Crescent Lake National Wildlife Refuge indicate that most of these lakes have seepage inflow from ground water and seepage outflow to ground water. The chemistry of inflowing ground water commonly has an effect on lake water chemistry. However, the chemistry of lake water can also affect ground water in areas of seepage from lakes. In the Crescent Lake area, for example, plumes of lake water were detected in ground water downgradient from the lakes, as indicated by the plume of dissolved organic carbon downgradient from Roundup Lake and Island Lake (Figure K–3).

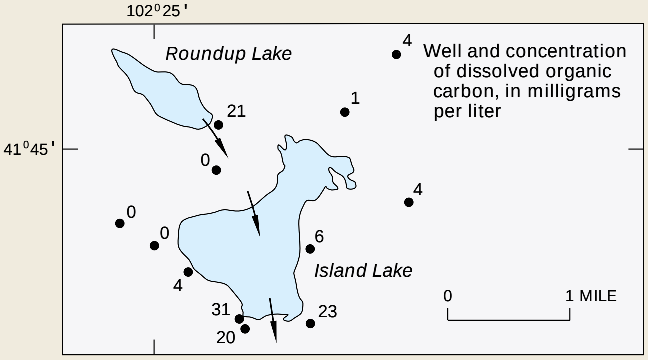
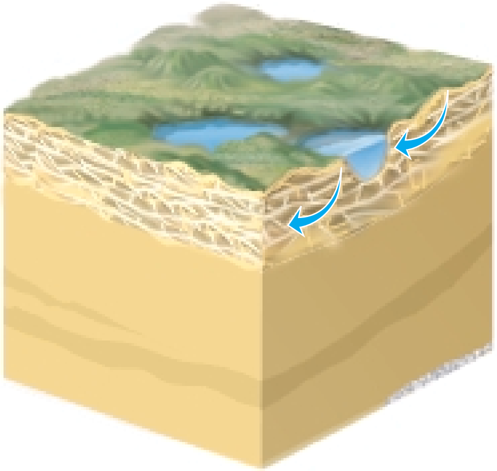


Figure K–3. Seepage from lakes in the sandhills of Nebraska causes plumes of dissolved organic carbon to be present in ground water on the downgradient sides of the lakes. (Modified from LaBaugh, J.W., 1986, Limnological characteristics of selected lakes in the Nebraska sandhills, U.S.A., and their relation to chemical characteristics of adjacent ground water: Journal of Hydrology, v. 86, p. 279–298.) (Reprinted with permission of Elsevier Science-NL, Amsterdam, The Netherlands.)

KARST TERRAIN

Karst may be broadly defined as all landforms that are produced primarily by the dissolution of rocks, mainly limestone and dolomite. Karst terrains (area K of the conceptual landscape, Figure 2) are characterized by (1) closed surface depressions of various sizes and shapes known as sinkholes, (2) an underground drainage network that consists of solution openings that range in size from enlarged cracks in the rock to large caves, and (3) highly disrupted surface drainage systems, which relate directly to the unique character of the underground drainage system.

Dissolution of limestone and dolomite guides the initial development of fractures into solution holes that are diagnostic of karst terrain. Perhaps nowhere else is the complex interplay between hydrology and chemistry so important to changes in landform. Limestone and dolomite weather quickly, producing calcium and magnesium carbonate waters that are relatively high in ionic strength. The increasing size of solution holes allows higher ground-water flow rates across a greater surface area of exposed minerals, which stimulates the dissolution process further, eventually leading to development of caves. Development of karst terrain also involves biological processes. Microbial production of carbon dioxide in the soil affects the carbonate equilibrium of water as it recharges ground water, which then affects how much mineral dissolution will take place before solute equilibrium is reached.



Ground-water recharge is very efficient in karst terrain because precipitation readily infiltrates through the rock openings that intersect the land surface. Water moves at greatly different rates through karst aquifers; it moves slowly through fine fractures and pores and rapidly through solution-enlarged fractures and conduits. As a result, the water discharging from many springs in karst terrain may be a combination of relatively slow-moving water draining from pores and rapidly moving storm-derived water. The slow-moving component tends to reflect the chemistry of the aquifer materials, and the more rapidly moving water associated with recent rainfall tends to reflect the chemical characteristics of precipitation and surface runoff.

Water movement in karst terrain is especially unpredictable because of the many paths ground water takes through the maze of fractures and solution openings in the rock (see Box L). Because of the large size of interconnected openings in well-developed karst systems, karst terrain can have true underground streams. These underground streams can have high rates of flow, in some places as great as rates of flow in surface streams. Furthermore, it is not unusual for medium-sized streams to disappear into the rock openings, thereby completely disrupting the surface drainage system, and to reappear at the surface at another place. Seeps and springs of all sizes are characteristic features of karst terrains. Springs having sufficiently large ground-water recharge areas commonly are the source of small- to medium-sized streams and constitute a large part of tributary flow to larger streams. In addition, the location where the streams emerge can change, depending on the spatial distribution of ground-water recharge in relation to individual precipitation events. Large spring inflows to streams in karst terrain contrast sharply with the generally more diffuse ground-water inflow characteristic of streams flowing across sand and gravel aquifers.

Because of the complex patterns of surface-water and ground-water flow in karst terrain, many studies have shown that surface-water drainage divides and ground-water drainage divides do not coincide. An extreme example is a stream that disappears in one surface-water basin and reappears in another basin. This situation complicates the identification of source areas for water and associated dissolved constituents, including contaminants, in karst terrain.

Water chemistry is widely used for studying the hydrology of karst aquifers. Extensive tracer studies (see Box G) and field mapping to locate points of recharge and discharge have been used to estimate the recharge areas of springs, rates of ground-water movement, and the water balance of aquifers. Variations in parameters such as temperature, hardness, calcium/magnesium ratios, and other chemical characteristics have been used to identify areas of ground-water recharge, differentiate rapid- and slow-moving ground-water flow paths, and compare springflow characteristics in different regions. Rapid transport of contaminants within karst aquifers and to springs has been documented in many locations. Because of the rapid movement of water in karst aquifers, water-quality problems that might be localized in other aquifer systems can become regional problems in karst systems.

Some landscapes considered to be karst terrain do not have carbonate rocks at the land surface. For example, in some areas of the southeastern United States, surficial deposits overlie carbonate rocks, resulting in a “mantled” karst terrain. Lakes and wetlands in mantled karst terrain interact with shallow ground water in a manner similar to that in sandy glacial and dune terrains. The difference between how lakes and wetlands interact with ground water in sandy glacial and dune terrain and how they interact in the mantled karst is related to the buried carbonate rocks. If dissolution of the buried carbonate rocks causes slumpage of an overlying confining bed, such that water can move freely through the confining bed, the lakes and wetlands also can be affected by changing hydraulic heads in the aquifers underlying the confining bed (see Box L).

Field Studies of Karst Terrain

Karst terrain is characteristic of regions that are underlain by limestone and dolomite bedrock. In many karst areas, the carbonate bedrock is present at land surface, but in other areas it may be covered by other deposits and is referred to as “mantled” karst. The Edwards Aquifer in south-central Texas is an example of karst terrain where the limestones and dolomites are exposed at land surface (Figure L–1). In this outcrop area, numerous solution cavities along vertical joints and sinkholes provide an efficient link between the land surface and the water table. Precipitation on the outcrop area tends to infiltrate rapidly into the ground, recharging ground water. In addition, a considerable amount of recharge to the aquifer is provided by losing streams that cross the outcrop area. Even the largest streams that originate to the north are dry in the outcrop area for most of the year. The unusual highway signs in this area go beyond local pride in a prolific water supply—they reflect a clear understanding of how vulnerable this water supply is to contamination by human activities at the land surface.

Just as solution cavities are major avenues for ground-water recharge, they also are focal points for ground-water discharge from karst aquifers. For example, springs near the margin of the Edwards Aquifer provide a continuous source of water for streams to the south.

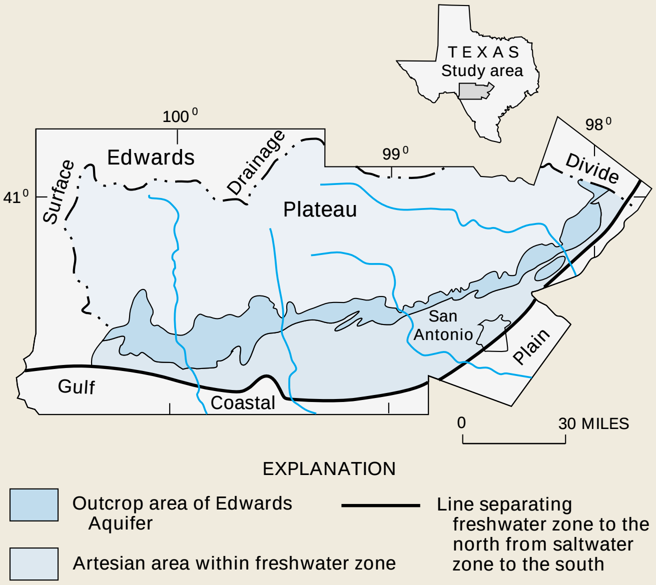
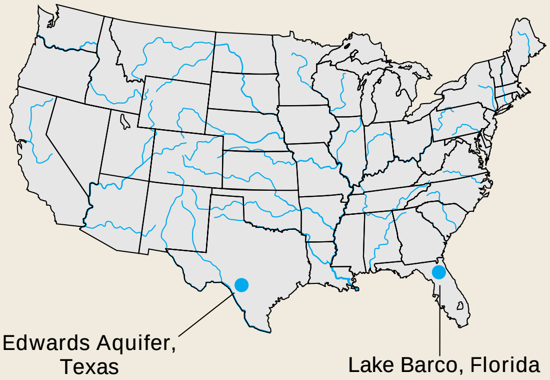


Figure L–1. A large area of karst terrain is associated with the Edwards Aquifer in south-central Texas. Large streams lose a considerable amount of water to ground water as they traverse the outcrop area of the Edwards Aquifer. (Modified from Brown, D.S., and Patton, J.T., 1995, Recharge to and discharge from the Edwards Aquifer in the San Antonio area, Texas, 1995: U.S. Geological Survey Open-File Report 96–181, 2 p.)

An example of mantled karst can be found in north-central Florida, a region that has many sinkhole lakes. In this region, unconsolidated deposits overlie the highly soluble limestone of the Upper Floridan aquifer. Most land-surface depressions containing lakes in Florida are formed when unconsolidated surficial deposits slump into sinkholes that form in the underlying limestone. Thus, although the lakes are not situated directly in limestone, the sinkholes in the bedrock underlying lakes commonly have a significant effect on the hydrology of the lakes.



Lake Barco is one of numerous lakes occupying depressions in northern Florida. Results of a study of the interaction of Lake Barco with ground water indicated that shallow ground water flows into the northern and northeastern parts of the lake, and lake water seeps out to shallow ground water in the western and southern parts (Figure L–2*A*). In addition, ground-water flow is downward beneath most of Lake Barco (Figure L–2*B*).

The studies of lake and ground-water chemistry included the use of tritium, chlorofluorocarbons (CFCs), and isotopes of oxygen (see Box G). The results indicated significant differences in the chemistry of (1) shallow ground water flowing into Lake Barco, (2) Lake Barco water, (3) shallow ground water downgradient from Lake Barco, and (4) deeper ground water beneath Lake Barco. Oxygen-rich lake water moving through the organic-rich lake sediments is reduced, resulting in discharge of oxygen-depleted water into the ground water beneath Lake Barco. This downward-moving ground water may have an undesired effect on the chemical quality of ground water in the underlying Upper Floridan aquifer, which is the principal source of water supply for the region. The patterns of ground-water movement determined from hydraulic-head data were corroborated by chemical tracers. For example, the dates that ground water in different parts of the flow system was recharged, as determined from CFC dating, show a fairly consistent increase in the length of time since recharge with depth (Figure L–2*C*).

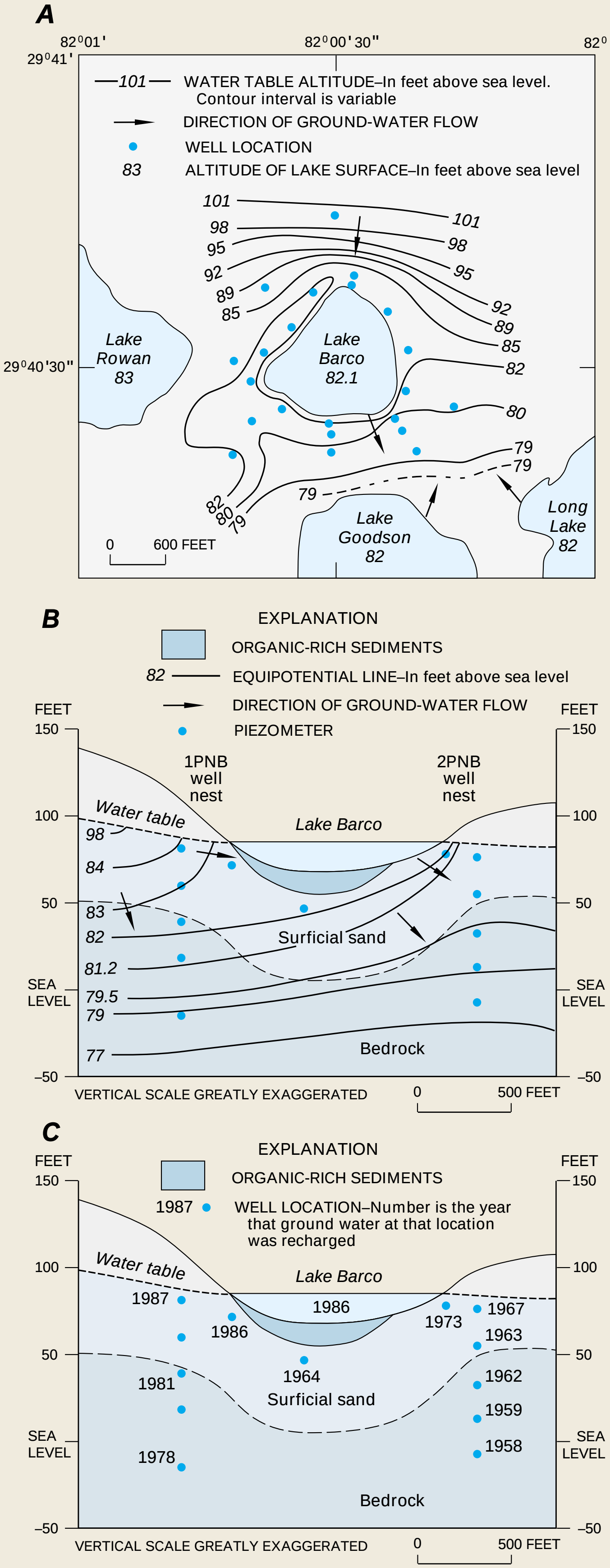


Figure L–2. Lake Barco, in northern Florida, is a flow-through lake with respect to ground water (A and B). The dates that ground water in different parts of the ground-water system was recharged indicate how long it takes water to move from the lake or water table to a given depth (C). (Modified from Katz, B.G., Lee, T.M., Plummer, L.N., and Busenberg, E., 1995, Chemical evolution of groundwater near a sinkhole lake, northern Florida, 1. Flow patterns, age of groundwater, and influence of lake water leakage: Water Resources Research, v. 31, no. 6, p. 1549–1564.)