

NATURAL PROCESSES OF GROUND-WATER AND SURFACE-WATER INTERACTION

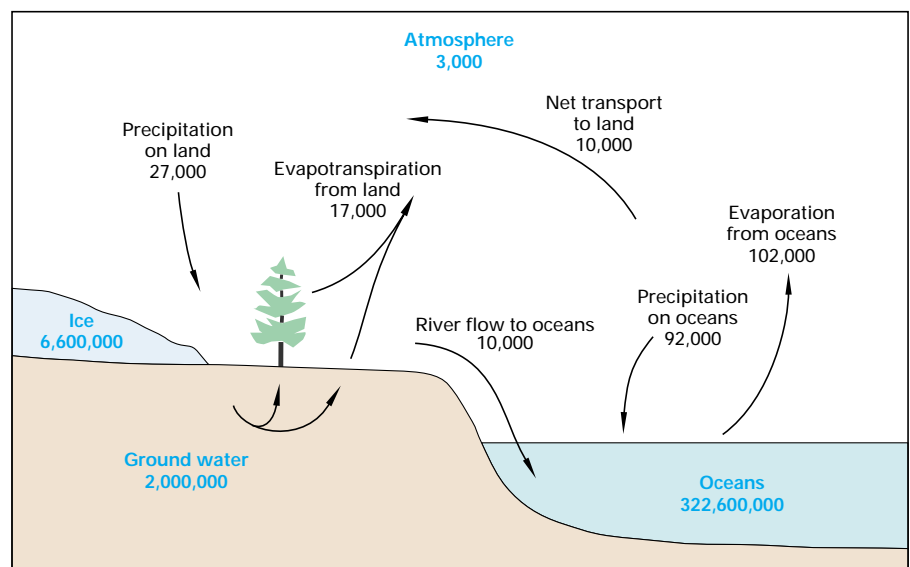
The Hydrologic Cycle and Interactions of Ground Water and Surface Water

The hydrologic cycle describes the continuous movement of water above, on, and below the surface of the Earth. The water on the Earth's surface—surface water—occurs as streams, lakes, and wetlands, as well as bays and oceans. Surface water also includes the solid forms of water—snow and ice. The water below the surface of the Earth primarily is ground water, but it also includes soil water.

The hydrologic cycle commonly is portrayed by a very simplified diagram that shows only major transfers of water between continents and oceans, as in Figure 1. However, for understanding hydrologic processes and managing water resources, the hydrologic cycle needs to be viewed at a wide range of scales and as having a great deal of vari-

ability in time and space. Precipitation, which is the source of virtually all freshwater in the hydrologic cycle, falls nearly everywhere, but its distribution is highly variable. Similarly, evaporation and transpiration return water to the atmosphere nearly everywhere, but evaporation and transpiration rates vary considerably according to climatic conditions. As a result, much of the precipitation never reaches the oceans as surface and subsurface runoff before the water is returned to the atmosphere. The relative magnitudes of the individual components of the hydrologic cycle, such as evapotranspiration, may differ significantly even at small scales, as between an agricultural field and a nearby woodland.

Figure 1. Ground water is the second smallest of the four main pools of water on Earth, and river flow to the oceans is one of the smallest fluxes, yet ground water and surface water are the components of the hydrologic system that humans use most. (Modified from Schelesinger, W.H., 1991, Biogeochemistry—An analysis of global change: Academic Press, San Diego, California.) (Used with permission.)



Pools are in cubic miles
Fluxes are in cubic miles per year

To present the concepts and many facets of the interaction of ground water and surface water in a unified way, a conceptual landscape is used (Figure 2). The conceptual landscape shows in a very general and simplified way the interaction of ground water with all types of surface water, such as streams, lakes, and wetlands, in many different terrains from the mountains to the oceans. The intent of Figure 2 is to emphasize that ground water and surface water interact at many places throughout the landscape.

Movement of water in the atmosphere and on the land surface is relatively easy to visualize, but the movement of ground water is not. Concepts related to ground water and the movement of ground water are introduced in Box A. As illustrated in Figure 3, ground water moves along flow paths of varying lengths from areas of recharge to areas of discharge. The generalized flow paths in Figure 3 start at the water table, continue through the ground-water system, and terminate at the stream or at the pumped well. The source of water to the water table (ground-water recharge) is infiltration of precipitation through the unsaturated zone. In the uppermost, unconfined aquifer, flow paths near the stream can be tens to hundreds of feet in length and have corresponding travel times of days to a few years. The longest and deepest flow paths in Figure 3 may be thousands of feet to tens of miles in length, and travel times may range from decades to millennia. In general, shallow ground water is more susceptible to contamination from human sources and activities because of its close proximity to the land surface. Therefore, shallow, local patterns of ground-water flow near surface water are emphasized in this Circular.

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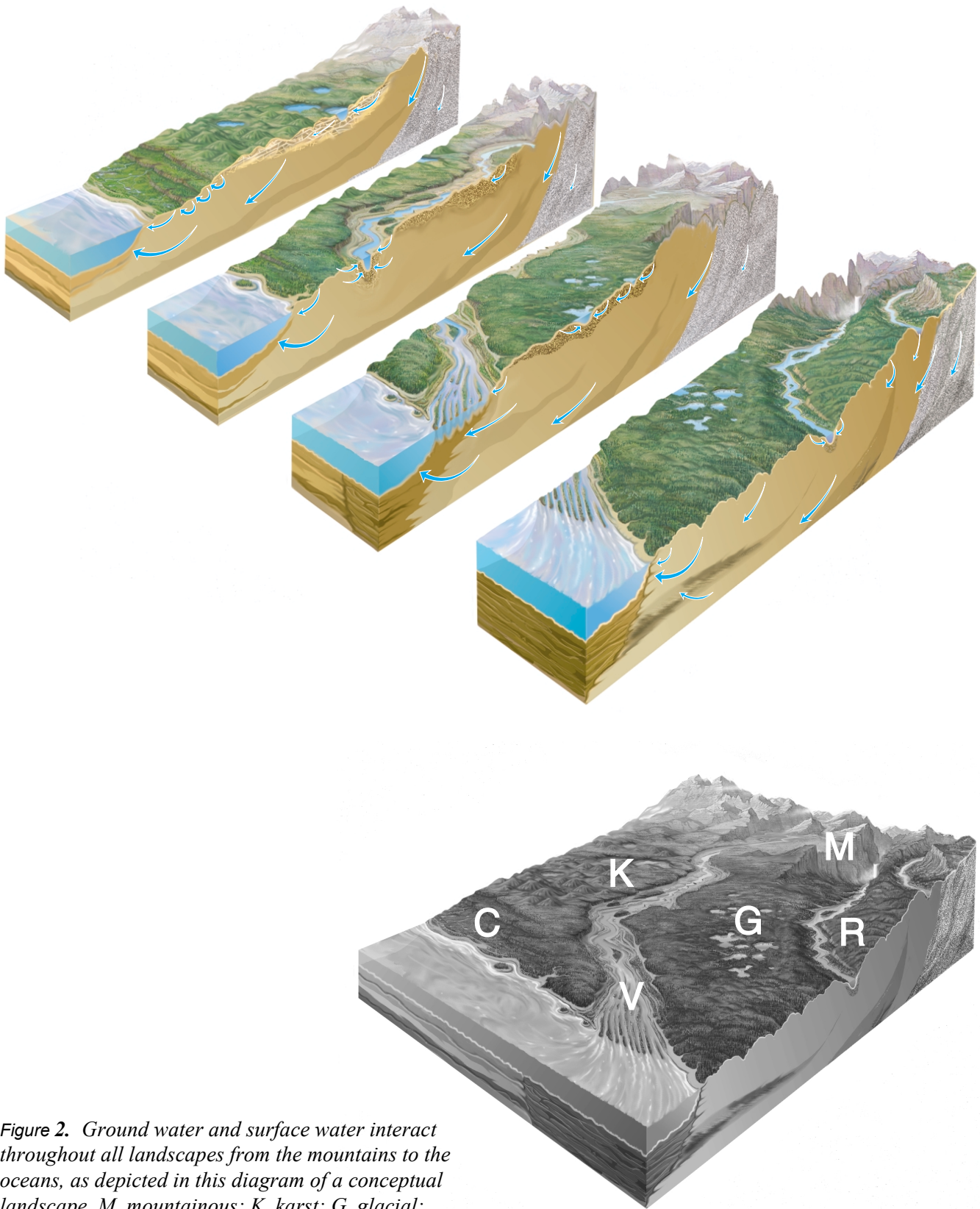


Figure 2. Ground water and surface water interact throughout all landscapes from the mountains to the oceans, as depicted in this diagram of a conceptual landscape. M, mountainous; K, karst; G, glacial; R, riverine (small); V, riverine (large); C, coastal.

Small-scale geologic features in beds of surface-water bodies affect seepage patterns at scales too small to be shown in Figure 3. For example, the size, shape, and orientation of the sediment grains in surface-water beds affect seepage patterns. If a surface-water bed consists of one sediment type, such as sand, inflow seepage is greatest at the shoreline, and it decreases in a nonlinear pattern away from the shoreline (Figure 4). Geologic units having different permeabilities also affect seepage distribution in surface-water beds. For example, a highly permeable sand layer within a surface-water bed consisting largely of silt will transmit water preferentially into the surface water as a spring (Figure 5).

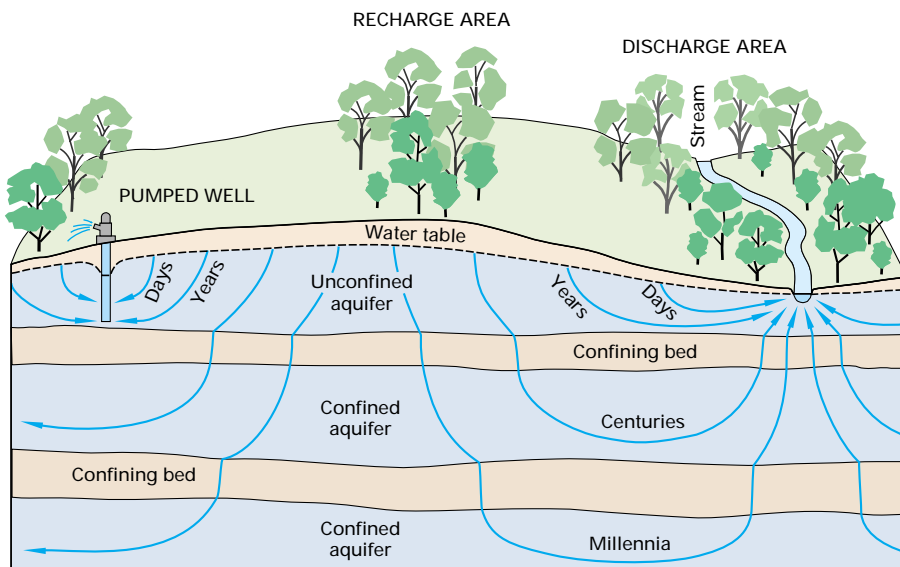


Figure 3. Ground-water flow paths vary greatly in length, depth, and traveltime from points of recharge to points of discharge in the ground-water system.

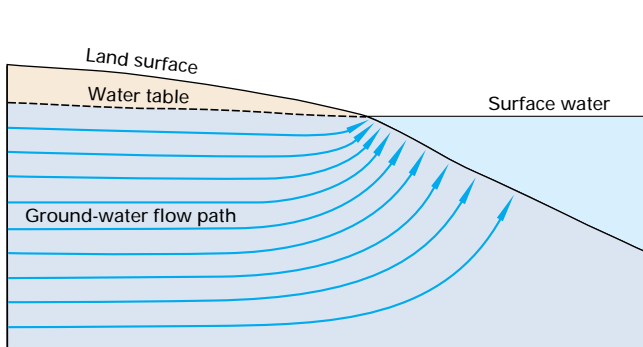


Figure 4. Ground-water seepage into surface water usually is greatest near shore. In flow diagrams such as that shown here, the quantity of discharge is equal between any two flow lines; therefore, the closer flow lines indicate greater discharge per unit of bottom area.

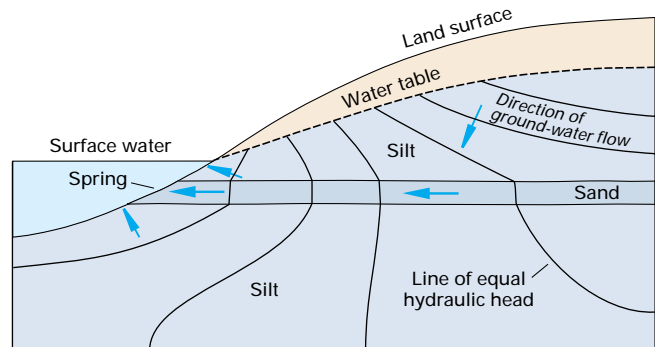


Figure 5. Subaqueous springs can result from preferred paths of ground-water flow through highly permeable sediments.

Concepts of Ground Water, Water Table, and Flow Systems

SUBSURFACE WATER

Water beneath the land surface occurs in two principal zones, the unsaturated zone and the saturated zone (Figure A-1). In the unsaturated zone, the voids—that is, the spaces between grains of gravel, sand, silt, clay, and cracks within rocks—contain both air and water. Although a considerable amount of water can be present in the unsaturated zone, this water cannot be pumped by wells because it is held too tightly by capillary forces. The upper part of the unsaturated zone is the soil-water zone. The soil zone is crisscrossed by roots, voids left by decayed roots, and animal and worm burrows, which enhance the infiltration of precipitation into the soil zone. Soil water is used by plants in life functions and transpiration, but it also can evaporate directly to the atmosphere.

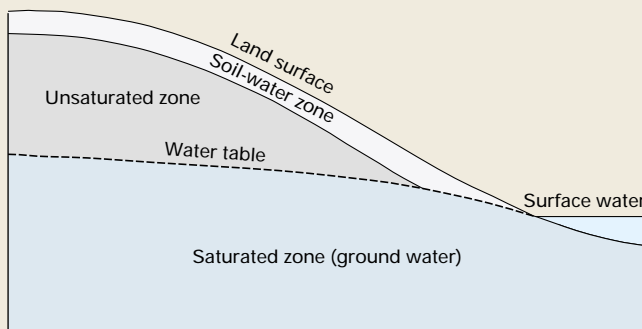


Figure A-1. The water table is the upper surface of the saturated zone. The water table meets surface-water bodies at or near the shoreline of surface water if the surface-water body is connected to the ground-water system.

In contrast to the unsaturated zone, the voids in the saturated zone are completely filled with water. Water in the saturated zone is referred to as ground water. The upper surface of the saturated zone is referred to as the water table. Below the water table, the water pressure is great enough to allow water to enter wells, thus permitting ground water to be withdrawn for use. A well is constructed by inserting a pipe into a drilled hole; a screen is attached, generally at its base, to prevent earth materials from entering the pipe along with the water pumped through the screen.

The depth to the water table is highly variable and can range from zero, when it is at land surface, to hundreds or even thousands of feet in some types of landscapes. Usually, the depth to the water table is small near permanent bodies of surface water such as streams, lakes, and wetlands. An important characteristic of the water table is that its configuration varies seasonally and from year to year because ground-water recharge, which is the accretion of water to the upper surface of the saturated zone, is related to the wide variation in the quantity, distribution, and timing of precipitation.

THE WATER TABLE

The depth to the water table can be determined by installing wells that penetrate the top of the saturated zone just far enough to hold standing water. Preparation of a water-table map requires that only wells that have their well screens placed near the water table be used. If the depth to water is measured at a number of such wells throughout an area of study, and if those water levels are referenced to a common datum such as sea level, the data can be contoured to indicate the configuration of the water table (Figure A-2).

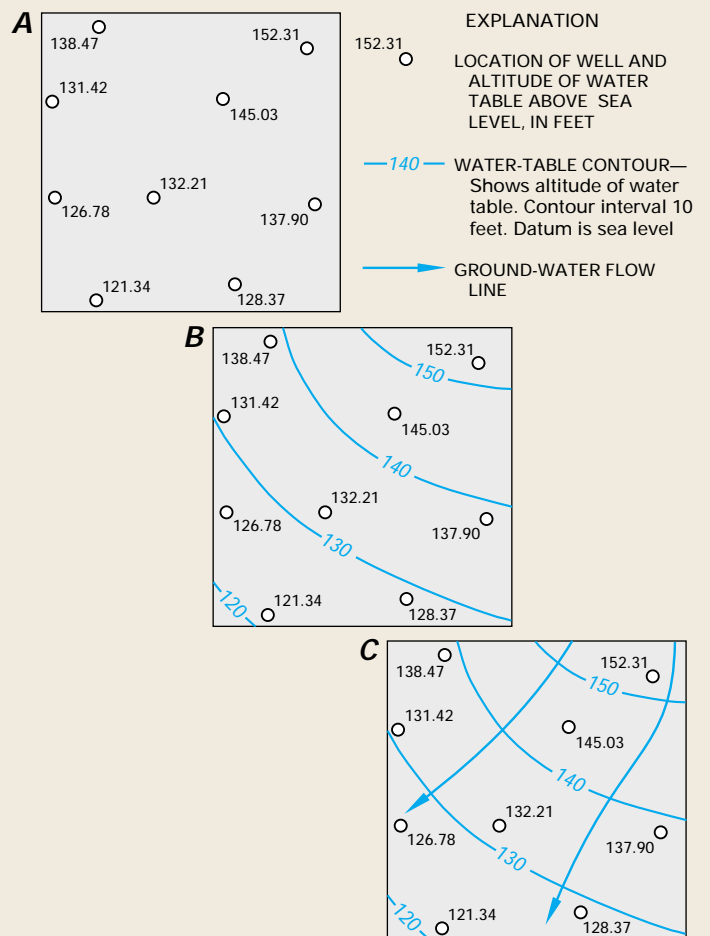


Figure A-2. Using known altitudes of the water table at individual wells (A), contour maps of the water-table surface can be drawn (B), and directions of ground-water flow along the water table can be determined (C) because flow usually is approximately perpendicular to the contours.

In addition to various practical uses of a water-table map, such as estimating an approximate depth for a proposed well, the configuration of the water table provides an indication of the approximate direction of ground-water flow at any location

on the water table. Lines drawn perpendicular to water-table contours usually indicate the direction of ground-water flow along the upper surface of the ground-water system. The water table is continually adjusting to changing recharge and discharge patterns. Therefore, to construct a water-table map, water-level measurements must be made at approximately the same time, and the resulting map is representative only of that specific time.

GROUND-WATER MOVEMENT

The ground-water system as a whole is actually a three-dimensional flow field; therefore, it is important to understand how the vertical components of ground-water movement affect the interaction of ground water and surface water. A vertical section of a flow field indicates how potential energy is distributed beneath the water table in the ground-water system and how the energy distribution can be used to determine vertical components of flow near a surface-water body. The term hydraulic head, which is the sum of elevation and water pressure divided by the weight density of water, is used to describe potential energy in ground-water flow systems. For example, Figure A-3 shows a generalized vertical section of subsurface water flow. Water that infiltrates at land surface moves vertically downward to the water table to become ground water. The ground water then moves both vertically and laterally within the ground-water system. Movement is downward and lateral on the right side of the diagram, mostly lateral in the center, and lateral and upward on the left side of the diagram.

Flow fields such as these can be mapped in a process similar to preparing water-table maps, except that vertically distributed piezometers need to be used instead of water-table wells. A piezometer is a well that has a very short screen so the water level represents hydraulic head in only a very small part of the ground-water system. A group of piezometers completed at different depths at the same location is referred to as a piezometer nest. Three such piezometer nests are shown in Figure A-3 (locations A, B, and C). By starting at a water-table contour, and using the water-level data from the piezometer nests, lines of equal hydraulic head can be drawn. Similar to drawing flow direction on water-table maps, flow lines can be drawn approximately perpendicular to these lines of equal hydraulic head, as shown in Figure A-3.

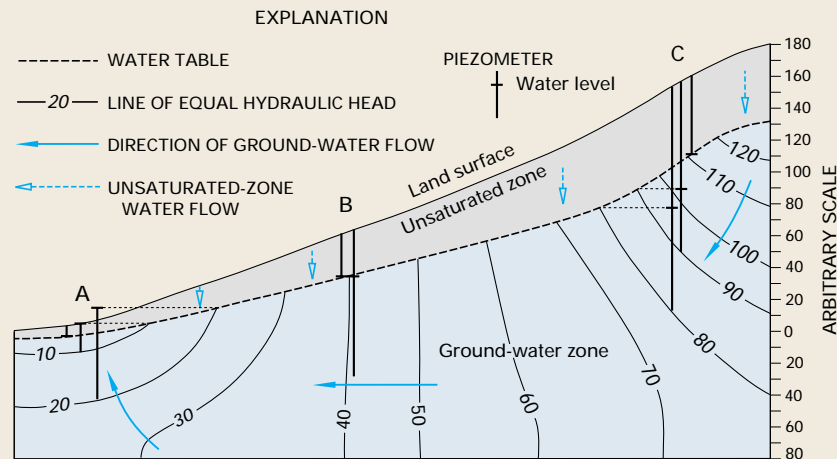


Figure A-3. If the distribution of hydraulic head in vertical section is known from nested piezometer data, zones of downward, lateral, and upward components of ground-water flow can be determined.

Actual flow fields generally are much more complex than that shown in Figure A-3. For example, flow systems of different sizes and depths can be present, and they can overlie one another, as indicated in Figure A-4. In a local flow system, water that recharges at a water-table high discharges to an adjacent lowland. Local flow systems are the most dynamic and the shallowest flow systems; therefore, they have the greatest interchange with surface water. Local flow systems can be underlain by intermediate and regional flow systems. Water in deeper flow systems have longer flow paths and longer contact time with subsurface materials; therefore, the water generally contains more dissolved chemicals. Nevertheless, these deeper flow systems also eventually discharge to surface water, and they can have a great effect on the chemical characteristics of the receiving surface water.

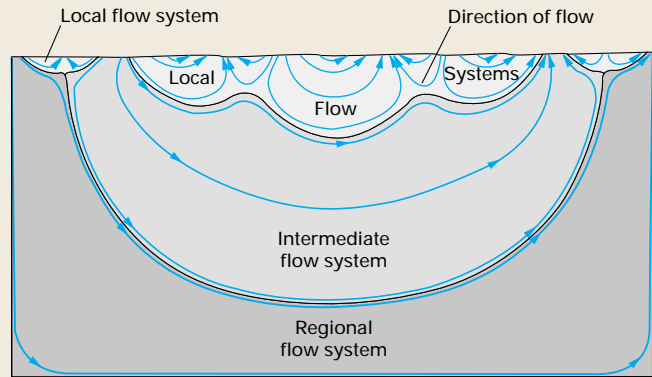


Figure A-4. Ground-water flow systems can be local, intermediate, and regional in scale. Much ground-water discharge into surface-water bodies is from local flow systems. (Figure modified from Toth, J., 1963, *A theoretical analysis of groundwater flow in small drainage basins*: p. 75-96 in *Proceedings of Hydrology Symposium No. 3, Groundwater*, Queen's Printer, Ottawa, Canada.)

GROUND-WATER DISCHARGE

The quantity of ground-water discharge (flux) to and from surface-water bodies can be determined for a known cross section of aquifer by multiplying the hydraulic gradient, which is determined from the hydraulic-head measurements

in wells and piezometers, by the permeability of the aquifer materials. Permeability is a quantitative measure of the ease of water movement through aquifer materials. For example, sand is more permeable than clay because the pore spaces between sand grains are larger than pore spaces between clay particles.

Figure A-3. If the distribution of hydraulic head in vertical section is known from nested piezometer data, zones of downward, lateral, and upward components of ground-water flow can be determined.

Changing meteorological conditions also strongly affect seepage patterns in surface-water beds, especially near the shoreline. The water table commonly intersects land surface at the shoreline, resulting in no unsaturated zone at this point. Infiltrating precipitation passes rapidly through a thin unsaturated zone adjacent to the shoreline, which causes water-table mounds to form quickly adjacent to the surface water (Figure 6). This process, termed focused recharge, can result in increased ground-water inflow to surface-water bodies, or it can cause inflow to surface-water bodies that normally have seepage to ground water. Each precipitation event has the potential to cause this highly transient flow condition near shorelines as well as at depressions in uplands (Figure 6).

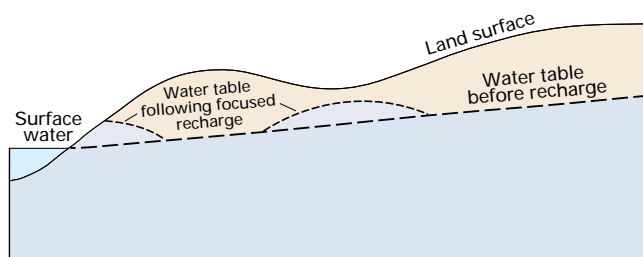


Figure 6. Ground-water recharge commonly is focused initially where the unsaturated zone is relatively thin at the edges of surface-water bodies and beneath depressions in the land surface.

Transpiration by nearshore plants has the opposite effect of focused recharge. Again, because the water table is near land surface at edges of surface-water bodies, plant roots can penetrate into the saturated zone, allowing the plants to transpire water directly from the ground-water system (Figure 7). Transpiration of ground water commonly results in a drawdown of the water table much like the effect of a pumped well. This highly variable daily and seasonal transpiration of ground water may significantly reduce ground-water discharge to a surface-water body or even cause movement of surface water into the subsurface. In many places it is possible to measure diurnal changes in the direction of flow during seasons of active plant growth; that is, ground water moves into the surface water during the night, and surface water moves into shallow ground water during the day.

These periodic changes in the direction of flow also take place on longer time scales: focused recharge from precipitation predominates during wet periods and drawdown by transpiration predominates during dry periods. As a result, the two processes, together with the geologic controls on seepage distribution, can cause flow conditions at the edges of surface-water bodies to be extremely variable. These “edge effects” probably affect small surface-water bodies more than large surface-water bodies because the ratio of edge length to total volume is greater for small water bodies than it is for large ones.

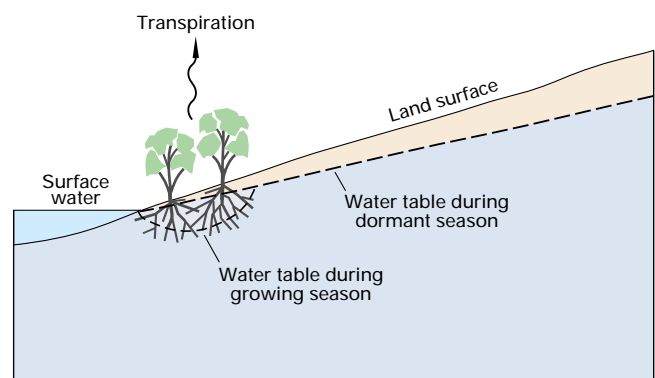


Figure 7. Where the depth to the water table is small adjacent to surface-water bodies, transpiration directly from ground water can cause cones of depression similar to those caused by pumping wells. This sometimes draws water directly from the surface water into the subsurface.

INTERACTION OF GROUND WATER AND STREAMS

Streams interact with ground water in all types of landscapes (see Box B). The interaction takes place in three basic ways: streams gain water from inflow of ground water through the streambed (gaining stream, Figure 8A), they lose water to ground water by outflow through the streambed (losing stream, Figure 9A), or they do both, gaining in some reaches and losing in other reaches. For ground water to discharge into a stream channel, the altitude of the water table in the vicinity of the stream must be higher than the alti-

tude of the stream-water surface. Conversely, for surface water to seep to ground water, the altitude of the water table in the vicinity of the stream must be lower than the altitude of the stream-water surface. Contours of water-table elevation indicate gaining streams by pointing in an upstream direction (Figure 8B), and they indicate losing streams by pointing in a downstream direction (Figure 9B) in the immediate vicinity of the stream.

Losing streams can be connected to the ground-water system by a continuous saturated zone (Figure 9A) or can be disconnected from

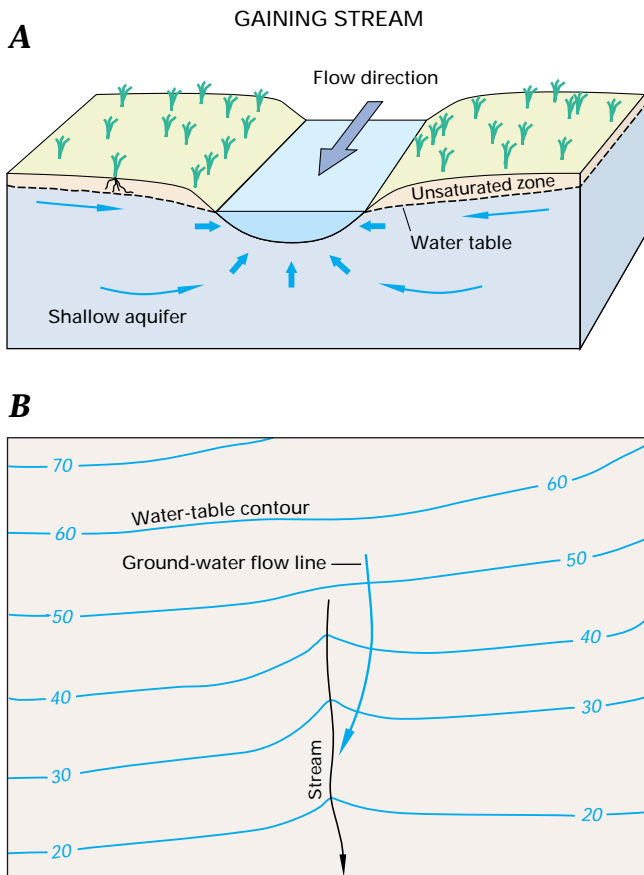


Figure 8. Gaining streams receive water from the ground-water system (A). This can be determined from water-table contour maps because the contour lines point in the upstream direction where they cross the stream (B).

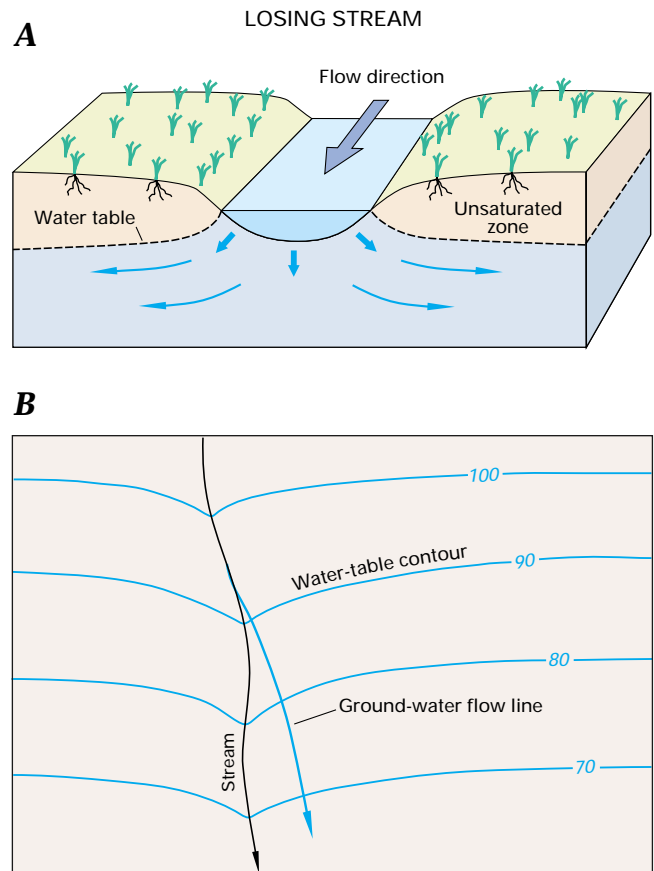


Figure 9. Losing streams lose water to the ground-water system (A). This can be determined from water-table contour maps because the contour lines point in the downstream direction where they cross the stream (B).

the ground-water system by an unsaturated zone. Where the stream is disconnected from the ground-water system by an unsaturated zone, the water table may have a discernible mound below the stream (Figure 10) if the rate of recharge through the streambed and unsaturated zone is greater than the rate of lateral ground-water flow away from the water-table mound. An important feature of streams that are disconnected from ground water is that pumping of shallow ground water near the stream does not affect the flow of the stream near the pumped wells.

In some environments, streamflow gain or loss can persist; that is, a stream might always gain water from ground water, or it might always lose water to ground water. However, in other envi-

ronments, flow direction can vary a great deal along a stream; some reaches receive ground water, and other reaches lose water to ground water. Furthermore, flow direction can change in very short timeframes as a result of individual storms causing focused recharge near the streambank, temporary flood peaks moving down the channel, or transpiration of ground water by streamside vegetation.

A type of interaction between ground water and streams that takes place in nearly all streams at one time or another is a rapid rise in stream stage that causes water to move from the stream into the streambanks. This process, termed bank storage (Figures 11 and 12B), usually is caused by storm precipitation, rapid snowmelt, or release of water

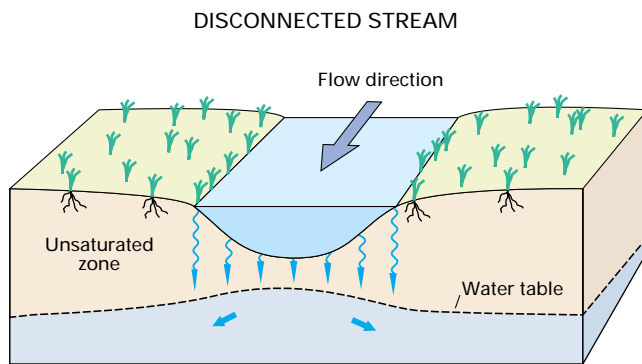


Figure 10. Disconnected streams are separated from the ground-water system by an unsaturated zone.

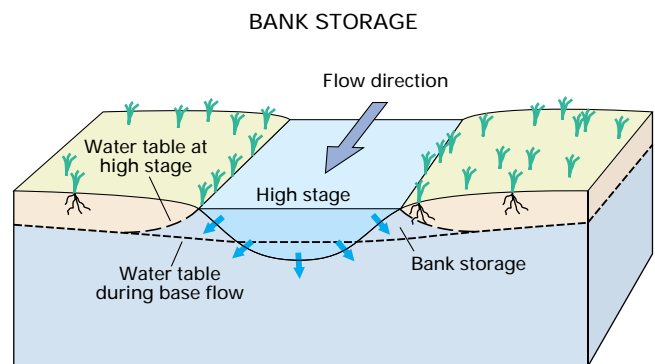


Figure 11. If stream levels rise higher than adjacent ground-water levels, stream water moves into the streambanks as bank storage.

“Streams interact with ground water in three basic ways: streams gain water from inflow of ground water through the streambed (gaining stream), they lose water to ground water by outflow through the streambed (losing stream), or they do both, gaining in some reaches and losing in other reaches”

from a reservoir upstream. As long as the rise in stage does not overtop the streambanks, most of the volume of stream water that enters the streambanks returns to the stream within a few days or weeks. The loss of stream water to bank storage and return of this water to the stream in a period of days or weeks tends to reduce flood peaks and later supplement stream flows. If the rise in stream stage is sufficient to overtop the banks and flood large areas of the land surface, widespread recharge to the water table can take place throughout the flooded area (Figure 12C). In this case, the time it takes for the recharged floodwater to return to the stream by ground-water flow may be weeks, months, or years because the lengths of the ground-water flow paths are much longer than those resulting from local bank storage. Depending on the frequency, magnitude, and intensity of storms and on the related magnitude of increases in stream stage, some streams and adjacent shallow aquifers may be in a continuous readjustment from interactions related to bank storage and overbank flooding.

In addition to bank storage, other processes may affect the local exchange of water between streams and adjacent shallow aquifers. Changes in streamflow between gaining and losing conditions can also be caused by pumping ground water

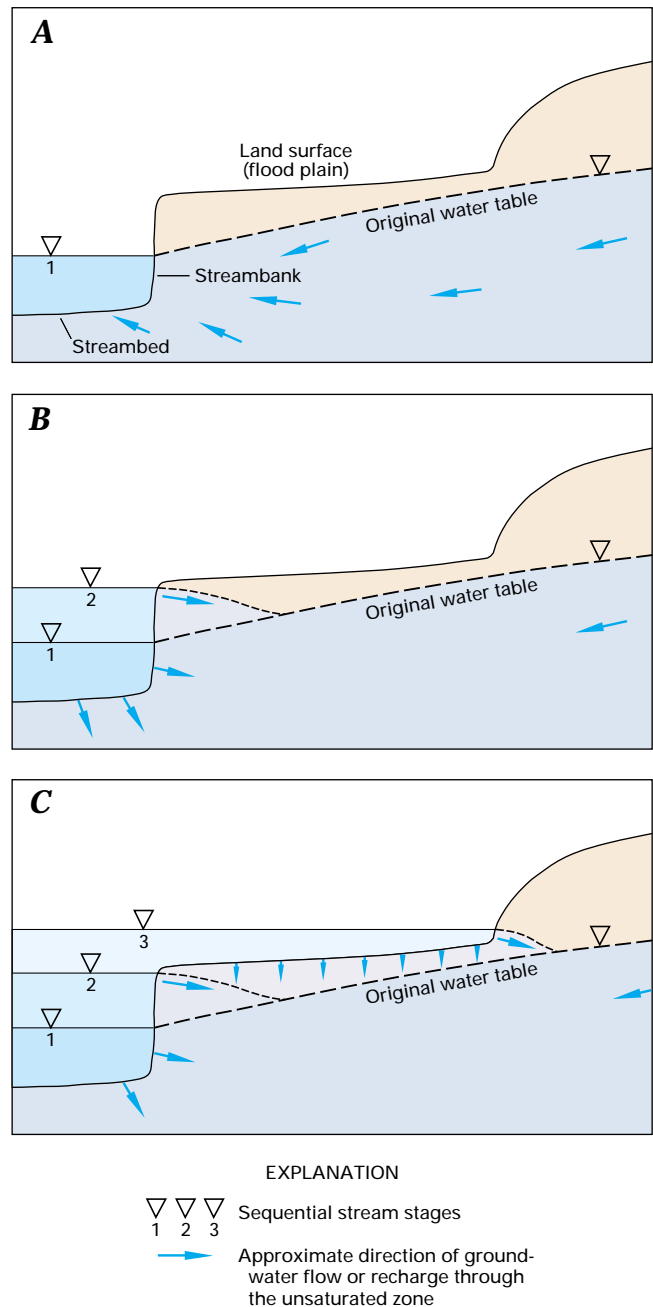


Figure 12. If stream levels rise higher than their streambanks (C), the floodwaters recharge ground water throughout the flooded areas.

near streams (see Box C). Pumping can intercept ground water that would otherwise have discharged to a gaining stream, or at higher pumping rates it can induce flow from the stream to the aquifer.

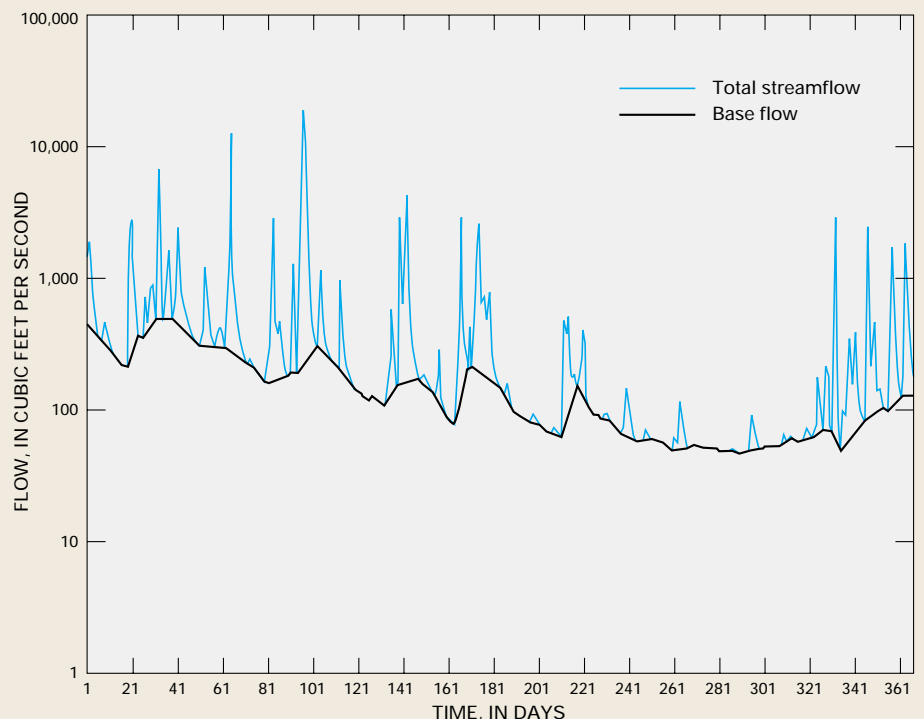
The Ground-Water Component of Streamflow

Ground water contributes to streams in most physiographic and climatic settings. Even in settings where streams are primarily losing water to ground water, certain reaches may receive ground-water inflow during some seasons. The proportion of stream water that is derived from ground-water inflow varies across physiographic and climatic settings. The amount of water that ground water contributes to streams can be estimated by analyzing streamflow hydrographs to determine the ground-water component, which is termed base flow (Figure B-1). Several different methods of analyzing hydrographs have been used by hydrologists to determine the base-flow component of streamflow.

One of the methods, which provides a conservative estimate of base flow, was used to determine the ground-water contribution to streamflow in 24 regions in the conterminous United States. The regions, delineated on the basis of physiography and climate, are believed to have common characteristics with respect to the interactions of ground water and surface water (Figure B-2). Fifty-four streams were selected for the analysis, at least two in each of the

24 regions. Streams were selected that had drainage basins less than 250 square miles and that had less than 3 percent of the drainage area covered by lakes and wetlands. Daily streamflow values for the 30-year period, 1961–1990, were used for the analysis of each stream. The analysis indicated that, for the 54 streams over the 30-year period, an average of 52 percent of the streamflow was contributed by ground water. Ground-water contributions ranged from 14 percent to 90 percent, and the median was 55 percent. The ground-water contribution to streamflow for selected streams can be compared in Figure B-2. As an example of the effect that geologic setting has on the contribution of ground water to streamflow, the Forest River in North Dakota can be compared to the Sturgeon River in Michigan. The Forest River Basin is underlain by poorly permeable silt and clay deposits, and only about 14 percent of its average annual flow is contributed by ground water; in contrast, the Sturgeon River Basin is underlain by highly permeable sand and gravel, and about 90 percent of its average annual flow is contributed by ground water.

Figure B-1. The ground-water component of streamflow was estimated from a streamflow hydrograph for the Homochitto River in Mississippi, using a method developed by the institute of Hydrology, United Kingdom. (Institute of Hydrology, 1980, Low flow studies: Wallingford, Oxon, United Kingdom, Research Report No. 1.)



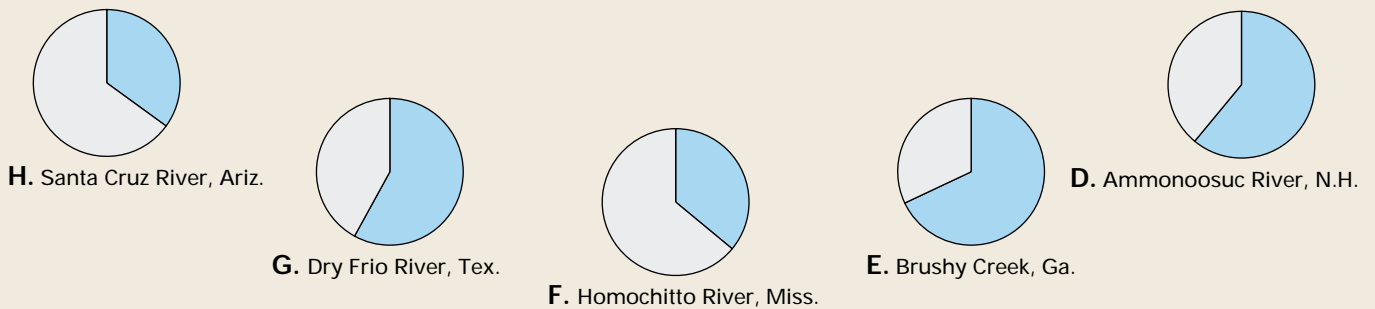
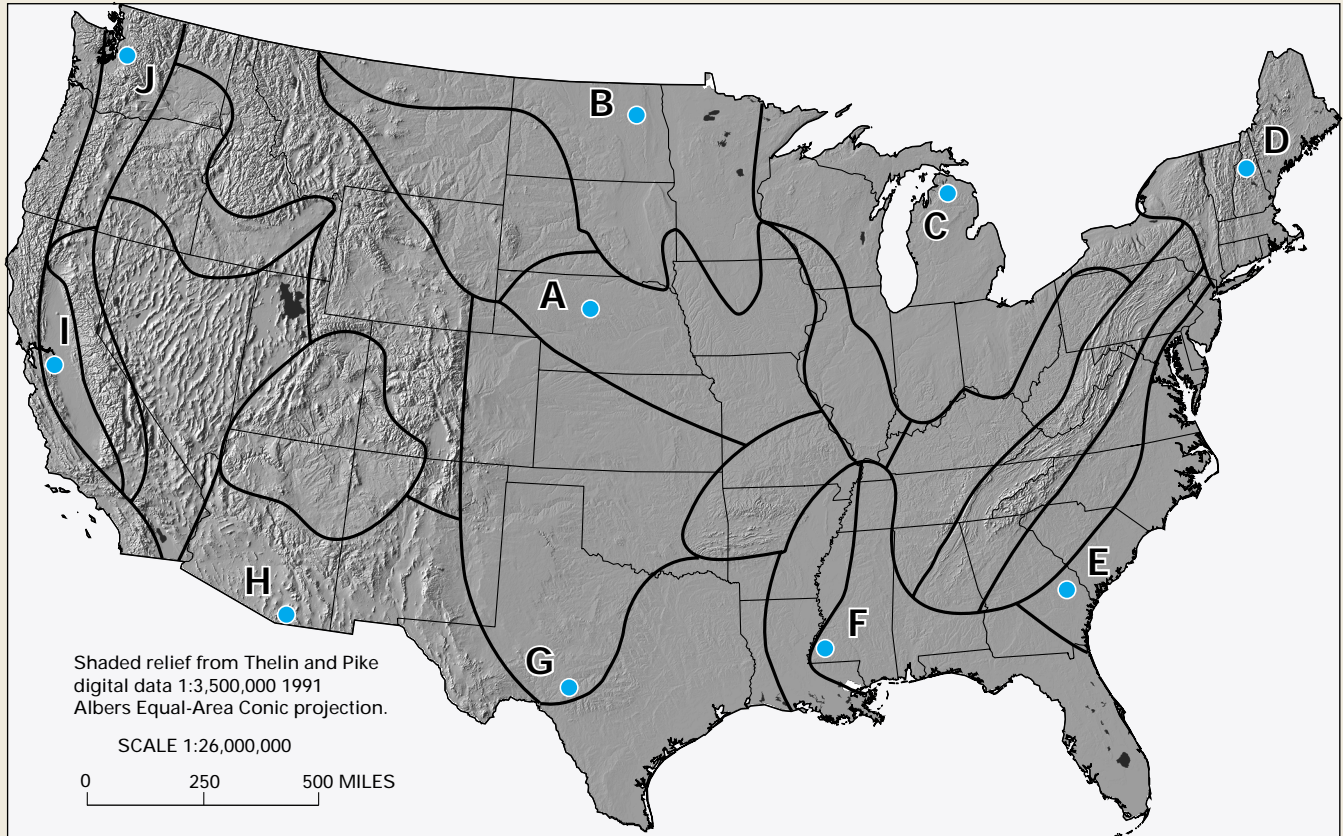
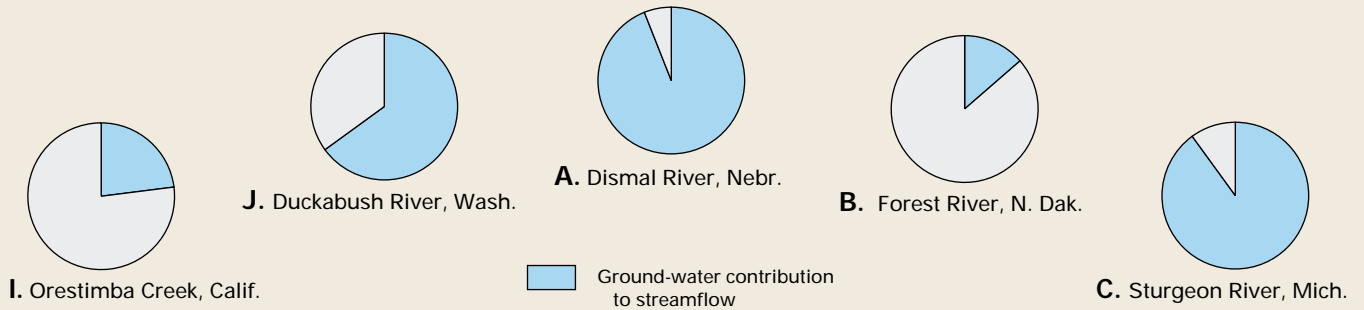


Figure B-2. In the conterminous United States, 24 regions were delineated where the interactions of ground water and surface water are considered to have similar characteristics. The estimated ground-water contribution to streamflow is shown for specific streams in 10 of the regions.

The Effect of Ground-Water Withdrawals on Surface Water

Withdrawing water from shallow aquifers that are directly connected to surface-water bodies can have a significant effect on the movement of water between these two water bodies. The effects of pumping a single well or a small group of wells on the hydrologic regime are local in scale. However, the effects of many wells withdrawing water from an aquifer over large areas may be regional in scale.

Withdrawing water from shallow aquifers for public and domestic water supply, irrigation, and industrial uses is widespread. Withdrawing water from shallow aquifers near surface-water bodies can diminish the available surface-water supply by capturing some of the ground-water flow that otherwise would have discharged to surface water or by inducing flow from surface water into the surrounding aquifer system. An analysis of the sources of water to a pumping well in a shallow aquifer that discharges to a stream is provided here to gain insight into how a pumping well can change the quantity and direction of flow between the shallow aquifer and the stream. Furthermore, changes in the direction of flow between the two water bodies can affect transport of contaminants associated with the moving water. Although a stream is used in the example, the results apply to all surface-water bodies, including lakes and wetlands.

A ground-water system under predevelopment conditions is in a state of dynamic equilibrium—for example, recharge at the water table is equal to ground-water discharge to a stream (Figure C-1A). Assume a well is installed and is pumped continually at a rate, Q_1 . After a new state of dynamic equilibrium is achieved, inflow to the ground-water system

from recharge will equal outflow to the stream plus the withdrawal from the well. In this new equilibrium, some of the ground water that would have discharged to the stream is intercepted by the well, and a ground-water divide, which is a line separating directions of flow, is established locally between the well and the stream (Figure C-1B). If the well is pumped at a higher rate, Q_2 , at a later time a new equilibrium is reached. Under this condition, the ground-water divide between the well and the stream is no longer present and withdrawals from the well induce movement of water from the stream into the aquifer (Figure C-1C). Thus, pumpage reverses the hydrologic condition of the stream in this reach from a ground-water discharge feature to a ground-water recharge feature.

In the hydrologic system depicted in Figures C-1A and C-1B, the quality of the stream water generally will have little effect on the quality of the shallow ground water. However, in the case of the well pumping at the higher rate, Q_2 (Figure C-1C), the quality of the stream water, which locally recharges the shallow aquifer, can affect the quality of ground water between the well and the stream as well as the quality of the ground water withdrawn from the well.

This hypothetical withdrawal of water from a shallow aquifer that discharges to a nearby surface-water body is a simplified but compelling illustration of the concept that ground water and surface water are one resource. In the long term, the quantity of ground water withdrawn is approximately equal to the reduction in streamflow that is potentially available to downstream users.

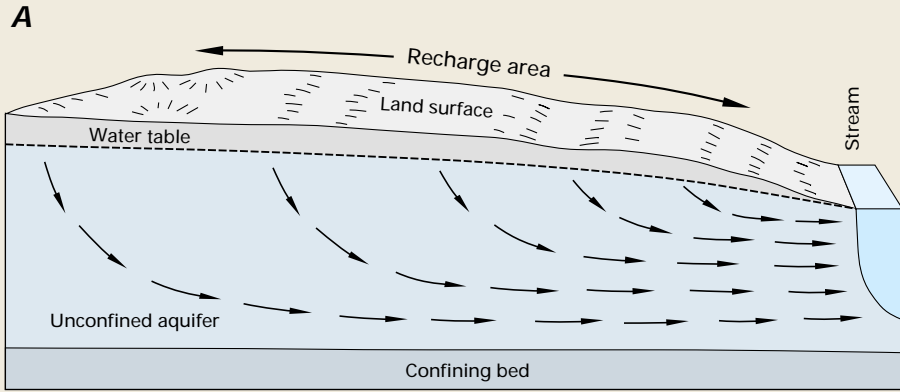
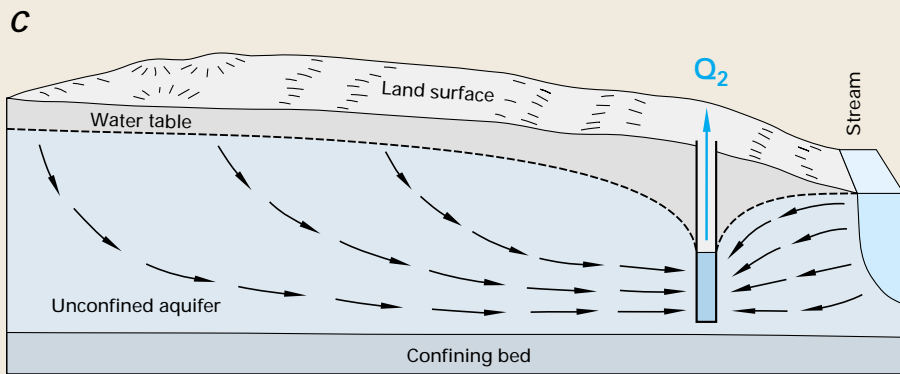
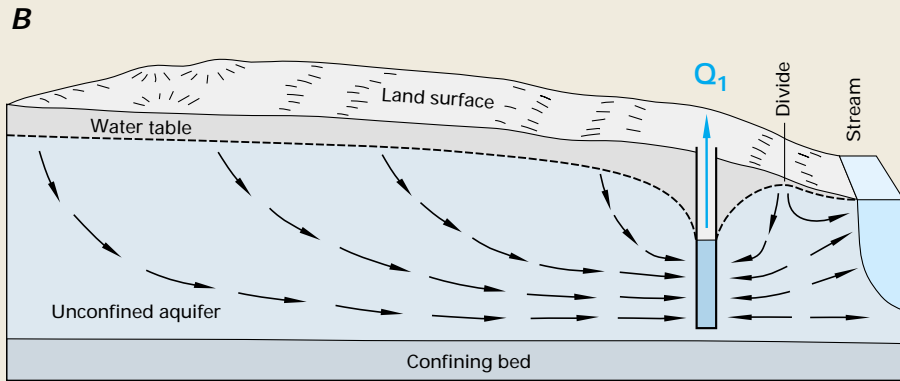


Figure C-1. In a schematic hydrologic setting where ground water discharges to a stream under natural conditions (A), placement of a well pumping at a rate (Q_1) near the stream will intercept part of the ground water that would have discharged to the stream (B). If the well is pumped at an even greater rate (Q_2), it can intercept additional water that would have discharged to the stream in the vicinity of the well and can draw water from the stream to the well (C).



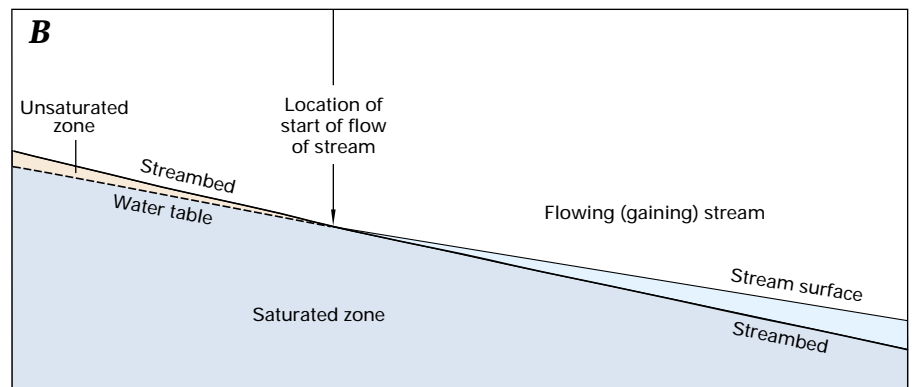
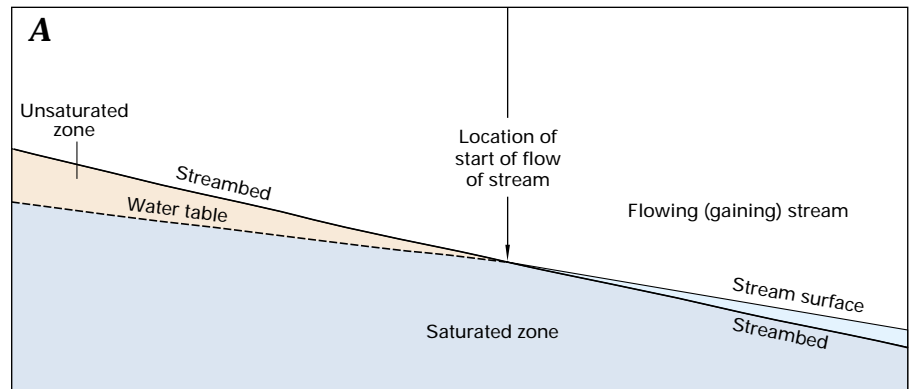
Where streamflow is generated in headwaters areas, the changes in streamflow between gaining and losing conditions may be particularly variable (Figure 13). The headwaters segment of streams can be completely dry except during storm events or during certain seasons of the year when snowmelt or precipitation is sufficient to maintain continuous flow for days or weeks. During these times, the stream will lose water to the unsaturated zone beneath its bed. However, as the water table rises through recharge in the headwaters area, the losing reach may become a gaining reach as the water table rises above the level of the stream. Under these conditions, the point where ground water first contributes to the stream gradually moves upstream.

Some gaining streams have reaches that lose water to the aquifer under normal conditions of streamflow. The direction of seepage through the bed of these streams commonly is related to abrupt changes in the slope of the streambed (Figure 14A) or to meanders in the stream channel (Figure 14B). For example, a losing stream reach

usually is located at the downstream end of pools in pool and riffle streams (Figure 14A), or upstream from channel bends in meandering streams (Figure 14B). The subsurface zone where stream water flows through short segments of its adjacent bed and banks is referred to as the hyporheic zone. The size and geometry of hyporheic zones surrounding streams vary greatly in time and space. Because of mixing between ground water and surface water in the hyporheic zone, the chemical and biological character of the hyporheic zone may differ markedly from adjacent surface water and ground water.

Ground-water systems that discharge to streams can underlie extensive areas of the land surface (Figure 15). As a result, environmental conditions at the interface between ground water and surface water reflect changes in the broader landscape. For example, the types and numbers of organisms in a given reach of streambed result, in part, from interactions between water in the hyporheic zone and ground water from distant sources.

Figure 13. The location where perennial streamflow begins in a channel can vary depending on the distribution of recharge in headwaters areas. Following dry periods (A), the start of streamflow will move up-channel during wet periods as the ground-water system becomes more saturated (B).



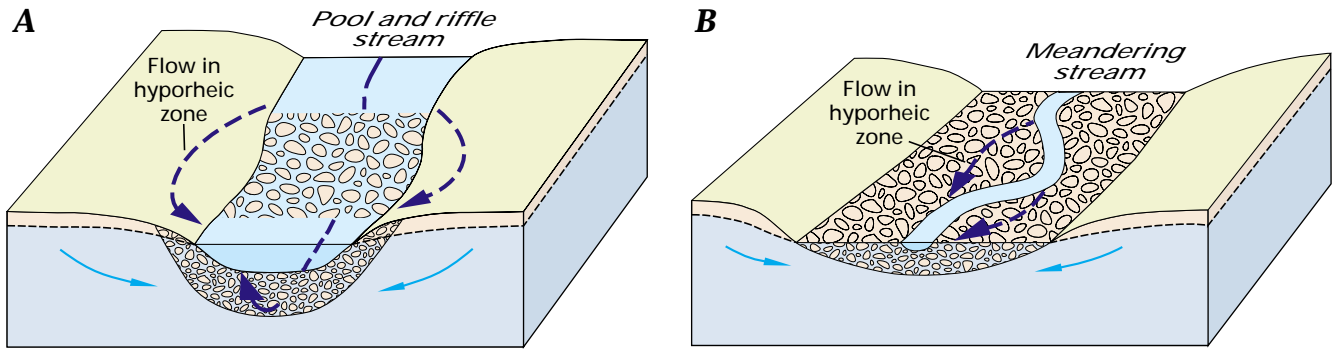


Figure 14. Surface-water exchange with ground water in the hyporheic zone is associated with abrupt changes in streambed slope (A) and with stream meanders (B).

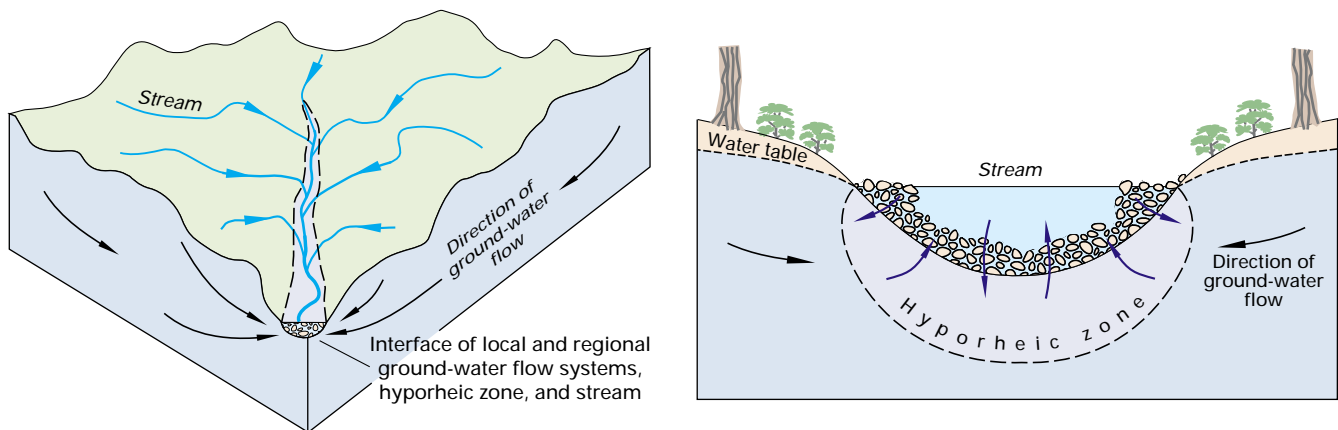


Figure 15. Streambeds and banks are unique environments because they are where ground water that drains much of the subsurface of landscapes interacts with surface water that drains much of the surface of landscapes.

INTERACTION OF GROUND WATER AND LAKES

Lakes interact with ground water in three basic ways: some receive ground-water inflow throughout their entire bed; some have seepage loss to ground water throughout their entire bed; but perhaps most lakes receive ground-water inflow through part of their bed and have seepage loss to ground water through other parts (Figure 16). Although these basic interactions are the same for lakes as they are for streams, the interactions differ in several ways.

The water level of natural lakes, that is, those not controlled by dams, generally does not change as rapidly as the water level of streams; therefore, bank storage is of lesser importance in lakes than it is in streams. Evaporation generally has a greater effect on lake levels than on stream levels because the surface area of lakes is generally larger and less shaded than many reaches of streams, and because lake water is not replenished as readily as a reach of a stream. Lakes can be present in many different parts of the landscape and can have complex ground-water flow systems associated with them. This is especially true for lakes in glacial and dune terrain, as is discussed in a later section of this Circular. Furthermore, lake sediments commonly have greater volumes of organic deposits than streams. These poorly permeable organic deposits can affect the distribution of seepage and biogeochemical exchanges of water and solutes more in lakes than in streams.

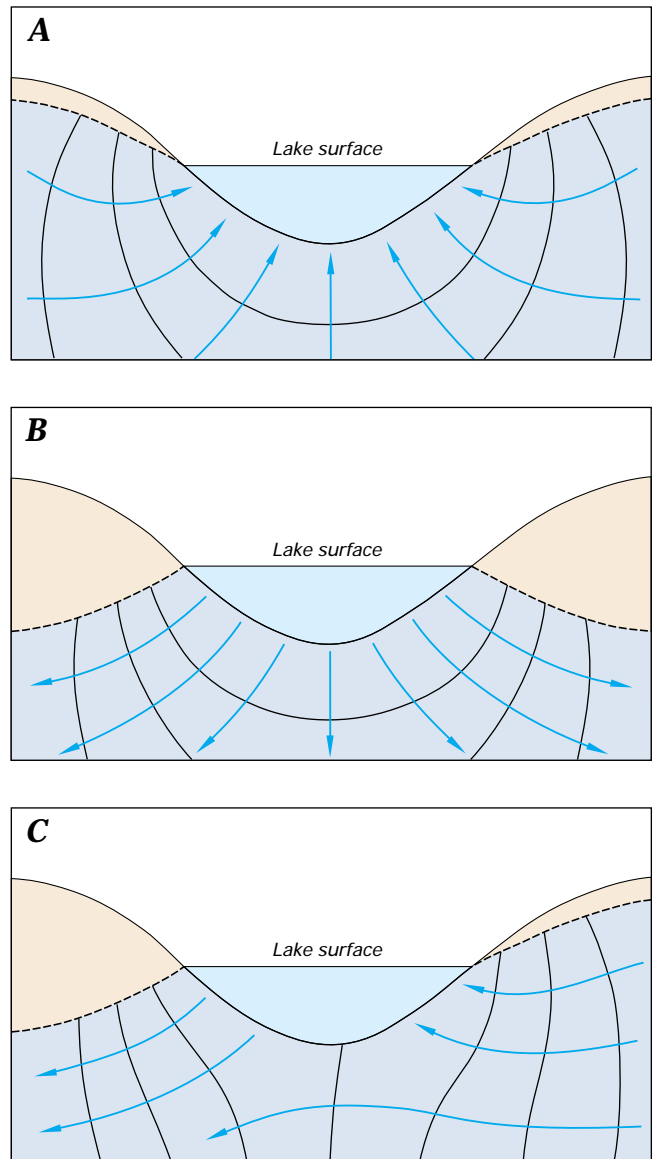


Figure 16. Lakes can receive ground-water inflow (A), lose water as seepage to ground water (B), or both

Reservoirs are human-made lakes that are designed primarily to control the flow and distribution of surface water. Most reservoirs are constructed in stream valleys; therefore, they have some characteristics both of streams and lakes. Like streams, reservoirs can have widely fluctuating levels, bank storage can be significant, and they commonly have a continuous flushing of water through them. Like lakes, reservoirs can have significant loss of water by evaporation, significant cycling of chemical and biological materials within their waters, and extensive biogeochemical exchanges of solutes with organic sediments.

“Lakes and wetlands can receive ground-water inflow throughout their entire bed, have outflow throughout their entire bed, or have both inflow and outflow at different localities”

INTERACTION OF GROUND WATER AND WETLANDS

Wetlands are present in climates and landscapes that cause ground water to discharge to land surface or that prevent rapid drainage of water from the land surface. Similar to streams and lakes, wetlands can receive ground-water inflow, recharge ground water, or do both. Those wetlands that occupy depressions in the land surface have interactions with ground water similar to lakes and streams. Unlike streams and lakes, however, wetlands do not always occupy low points and depressions in the landscape (Figure 17A); they also can be present on slopes (such as fens) or even on drainage divides (such as some types of bogs). Fens are wetlands that commonly receive ground-water discharge (Figure 17B); therefore, they receive a continuous supply of chemical constituents dissolved in the ground water. Bogs are wetlands that occupy uplands (Figure 17D) or extensive flat areas, and they receive much of their water and chemical constituents from precipitation. The distribution of major wetland areas in the United States is shown in Figure 18.

In areas of steep land slopes, the water table sometimes intersects the land surface, resulting in ground-water discharge directly to the land surface. The constant source of water at these seepage faces (Figure 17B) permits the growth of wetland plants. A constant source of ground water to wetland plants is also provided to parts of the landscape that are downgradient from breaks in slope of the water table (Figure 17B), and where

subsurface discontinuities in geologic units cause upward movement of ground water (Figure 17A). Many wetlands are present along streams, especially slow-moving streams. Although these riverine wetlands (Figure 17C) commonly receive ground-water discharge, they are dependent primarily on the stream for their water supply.

Wetlands in riverine and coastal areas have especially complex hydrological interactions because they are subject to periodic water-level changes. Some wetlands in coastal areas are affected by very predictable tidal cycles. Other coastal wetlands and riverine wetlands are more affected by seasonal water-level changes and by flooding. The combined effects of precipitation, evapotranspiration, and interaction with surface water and ground water result in a pattern of water depths in wetlands that is distinctive.

Hydroperiod is a term commonly used in wetland science that refers to the amplitude and frequency of water-level fluctuations. Hydroperiod affects all wetland characteristics, including the type of vegetation, nutrient cycling, and the types of invertebrates, fish, and bird species present.

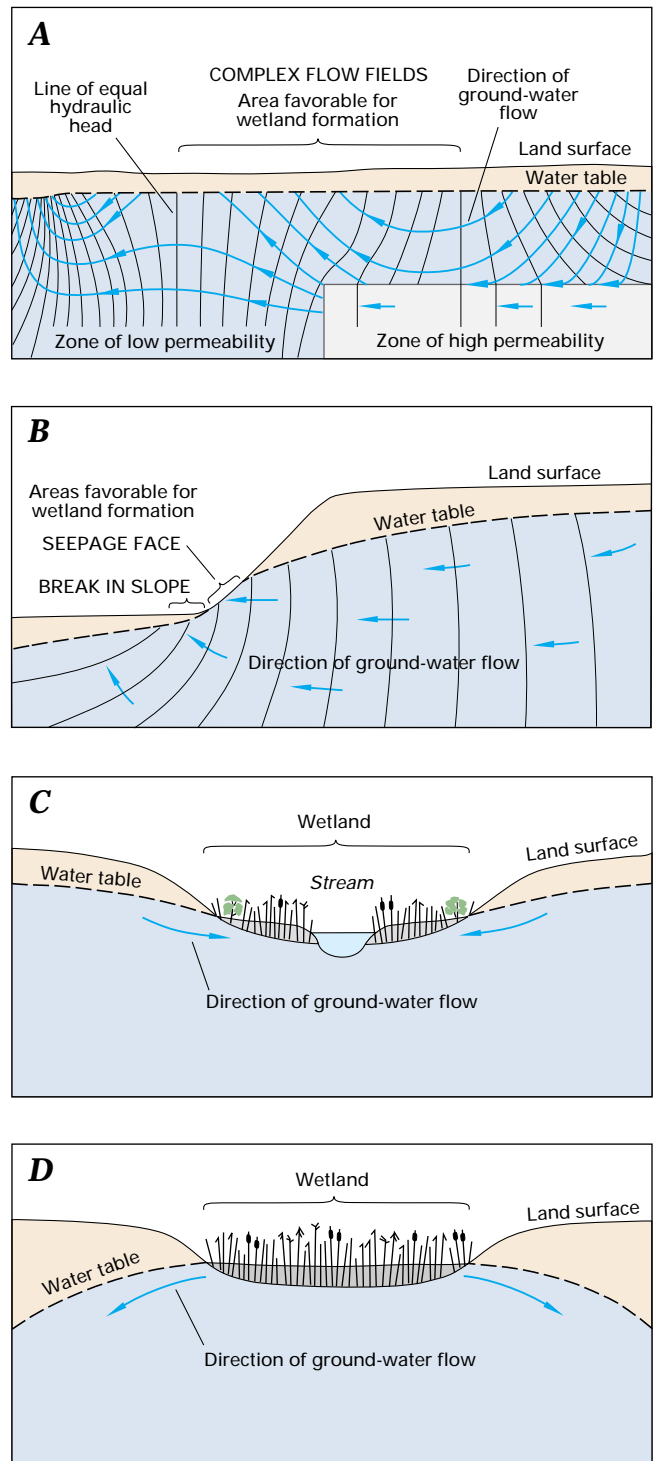


Figure 17. The source of water to wetlands can be from ground-water discharge where the land surface is underlain by complex ground-water flow fields (A), from ground-water discharge at seepage faces and at breaks in slope of the water table (B), from streams (C), and from precipitation in cases where wetlands have no stream inflow and ground-water gradients slope away from the wetland (D).

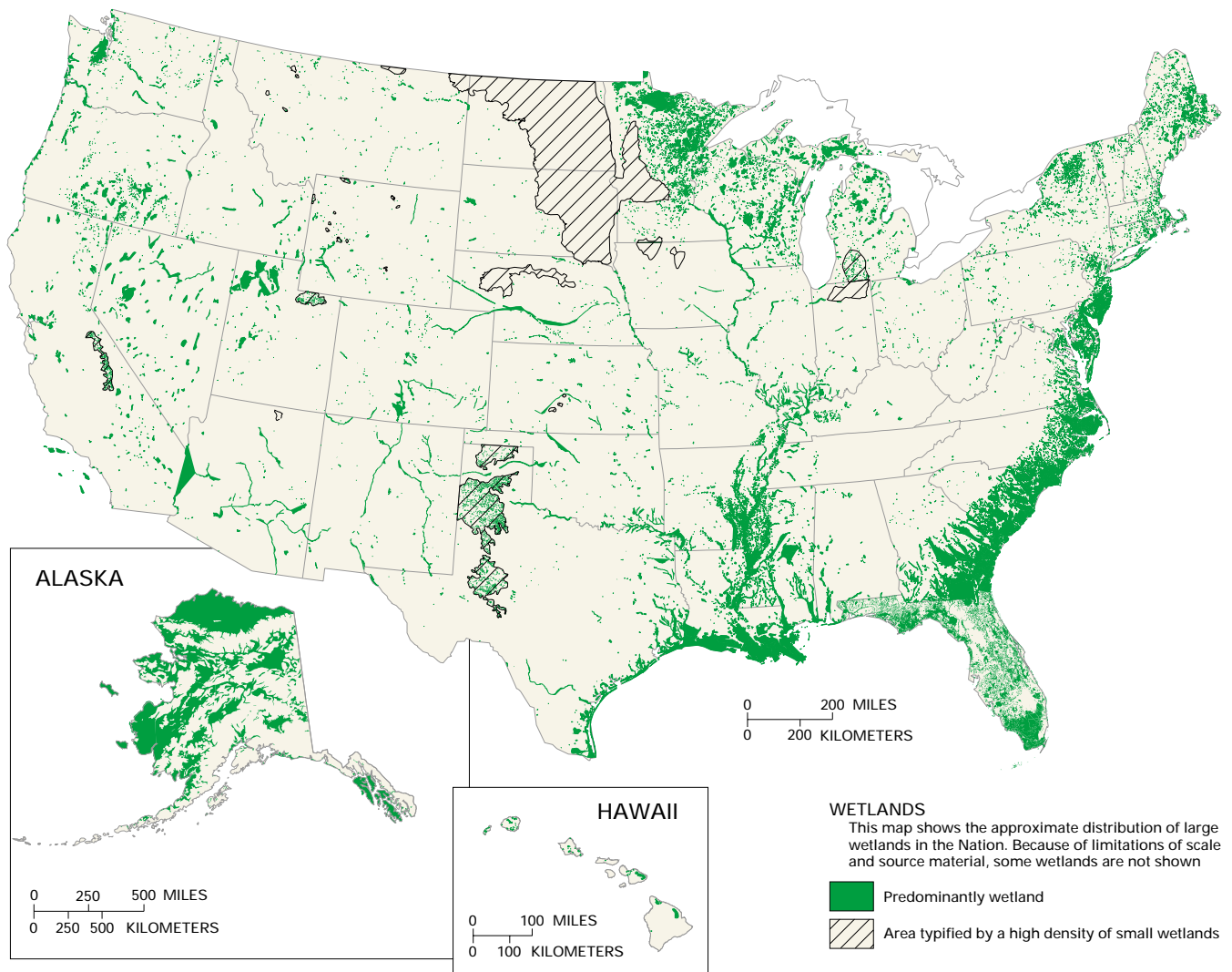


Figure 18. Wetlands are present throughout the Nation, but they cover the largest areas in the glacial terrain of the north-central United States, coastal terrain along the Atlantic and gulf coasts, and riverine terrain in the lower Mississippi River Valley.

A major difference between lakes and wetlands, with respect to their interaction with ground water, is the ease with which water moves through their beds. Lakes commonly are shallow around their perimeter where waves can remove fine-grained sediments, permitting the surface water and ground water to interact freely. In wetlands, on the other hand, if fine-grained and highly decomposed organic sediments are present near the wetland edge, the transfer of water and solutes between ground water and surface water is likely to be much slower.

Another difference in the interaction between ground water and surface water in wetlands compared to lakes is determined by rooted vegetation in wetlands. The fibrous root mat in wetland soils is highly conductive to water flow; therefore, water uptake by roots of emergent plants results in significant interchange between surface water and pore water of wetland sediments. The water exchanges in this upper soil zone even if exchange between surface water and ground water is restricted at the base of the wetland sediments.

Chemical Interactions of Ground Water and Surface Water

EVOLUTION OF WATER CHEMISTRY IN DRAINAGE BASINS

Two of the fundamental controls on water chemistry in drainage basins are the type of geologic materials that are present and the length of time that water is in contact with those materials. Chemical reactions that affect the biological and geochemical characteristics of a basin include (1) acid-base reactions, (2) precipitation and dissolution of minerals, (3) sorption and ion exchange, (4) oxidation-reduction reactions, (5) biodegradation, and (6) dissolution and exsolution of gases (see Box D). When water first infiltrates the land surface, microorganisms in the soil have a significant effect on the evolution of water chemistry. Organic matter in soils is degraded by

microbes, producing high concentrations of dissolved carbon dioxide (CO_2). This process lowers the pH by increasing the carbonic acid (H_2CO_3) concentration in the soil water. The production of carbonic acid starts a number of mineral-weathering reactions, which result in bicarbonate (HCO_3^-) commonly being the most abundant anion in the water. Where contact times between water and minerals in shallow groundwater flow paths are short, the dissolved-solids concentration in the water generally is low. In such settings, limited chemical changes take place before ground water is discharged to surface water.

“Two of the fundamental controls on water chemistry in drainage basins are the type of geologic materials that are present and the length of time that water is in contact with those materials”

In deeper ground-water flow systems, the contact time between water and minerals is much longer than it is in shallow flow systems. As a result, the initial importance of reactions relating to microbes in the soil zone may be superseded over time by chemical reactions between minerals and water (geochemical weathering). As weathering progresses, the concentration of dissolved solids increases. Depending on the chemical composition of the minerals that are weathered, the relative abundance of the major inorganic chemicals dissolved in the water changes (see Box E).

Surface water in streams, lakes, and wetlands can repeatedly interchange with nearby ground water. Thus, the length of time water is in contact with mineral surfaces in its drainage basin can continue after the water first enters a stream, lake, or wetland. An important consequence of these continued interchanges between surface water and ground water is their potential to further increase the contact time between water and chemically reactive geologic materials.

CHEMICAL INTERACTIONS OF GROUND WATER AND SURFACE WATER IN STREAMS, LAKES, AND WETLANDS

Ground-water chemistry and surface-water chemistry cannot be dealt with separately where surface and subsurface flow systems interact. The movement of water between ground water and surface water provides a major pathway for chemical transfer between terrestrial and aquatic systems (see Box F). This transfer of chemicals affects the supply of carbon, oxygen, nutrients such as nitrogen and phosphorus, and other chemical constituents that enhance biogeochemical processes on both sides of the interface. This transfer can ultimately affect the biological and chemical characteristics of aquatic systems downstream.

“The movement of water between ground water and surface water provides a major pathway for chemical transfer between terrestrial and aquatic systems”

Some Common Types of Biogeochemical Reactions Affecting Transport of Chemicals in Ground Water and Surface Water

ACID-BASE REACTIONS

Acid-base reactions involve the transfer of hydrogen ions (H^+) among solutes dissolved in water, and they affect the effective concentrations of dissolved chemicals through changes in the H^+ concentration in water. A brief notation for H^+ concentration (activity) is pH, which represents a negative logarithmic scale of the H^+ concentration. Smaller values of pH represent larger concentrations of H^+ , and larger values of pH represent smaller concentrations of H^+ . Many metals stay dissolved when pH values are small; increased pH causes these metals to precipitate from solution.

PRECIPITATION AND DISSOLUTION OF MINERALS

Precipitation reactions result in minerals being formed (precipitated) from ions that are dissolved in water. An example of this type of reaction is the precipitation of iron, which is common in areas of ground-water seeps and springs. At these locations, the solid material iron hydroxide is formed when iron dissolved in ground water comes in contact with oxygen dissolved in surface water. The reverse, or dissolution reactions, result in ions being released into water by dissolving minerals. An example is the release of calcium ions (Ca^{++}) and bicarbonate ions (HCO_3^-) when calcite ($CaCO_3$) in limestone is dissolved.

SORPTION AND ION EXCHANGE

Sorption is a process in which ions or molecules dissolved in water (solutes) become attached to the surfaces (or near-surface parts) of solid materials, either temporarily or permanently. Thus, solutes in ground water and surface water can be sorbed either to the solid materials that comprise an aquifer or streambed or to particles suspended in ground water or surface water. The attachments of positively charged ions to clays and of pesticides to solid surfaces are examples of sorption. Release of sorbed chemicals to water is termed desorption.

When ions attached to the surface of a solid are replaced by ions that were in water, the process is known as ion exchange. Ion exchange is the process that takes place in water softeners; ions that contribute to water hardness—calcium and magnesium—are exchanged for sodium on the surface of the solid. The result of this process is that the amount of calcium and magnesium in the water declines and the amount of sodium increases. The opposite takes place when saltwater enters an aquifer; some of the sodium in the saltwater is exchanged for calcium sorbed to the solid material of the aquifer.

OXIDATION-REDUCTION REACTIONS

Oxidation-reduction (redox) reactions take place when electrons are exchanged among solutes. In these reactions, oxidation (loss of electrons) of certain elements is accompanied by the reduction (gain of electrons) of other elements.

For example, when iron dissolved in water that does not contain dissolved oxygen mixes with water that does contain dissolved oxygen, the iron and oxygen interact by oxidation and reduction reactions. The result of the reactions is that the dissolved iron loses electrons (the iron is oxidized) and oxygen gains electrons (the oxygen is reduced). In this case, the iron is an electron donor and the oxygen is an electron acceptor. Bacteria can use energy gained from oxidation-reduction reactions as they decompose organic material. To accomplish this, bacterially mediated oxidation-reduction reactions use a sequence of electron acceptors, including oxygen, nitrate, iron, sulfate, and carbon dioxide. The presence of the products of these reactions in ground water and surface water can be used to identify the dominant oxidation-reduction reactions that have taken place in those waters. For example, the bacterial reduction of sulfate (SO_4^{2-}) to sulfide (HS^-) can result when organic matter is oxidized to CO_2 .

BIODEGRADATION

Biodegradation is the decomposition of organic chemicals by living organisms using enzymes. Enzymes are specialized organic compounds made by living organisms that speed up reactions with other organic compounds. Microorganisms degrade (transform) organic chemicals as a source of energy and carbon for growth. Microbial processes are important in the fate and transport of many organic compounds. Some compounds, such as petroleum

hydrocarbons, can be used directly by microorganisms as food sources and are rapidly degraded in many situations. Other compounds, such as chlorinated solvents, are not as easily assimilated. The rate of biodegradation of an organic chemical is dependent on its chemical structure, the environmental conditions, and the types of microorganisms that are present. Although biodegradation commonly can result in complete degradation of organic chemicals to carbon dioxide, water, and other simple products, it also can lead to intermediate products that are of environmental concern. For example, deethylatrazine, an intermediate degradation product of the pesticide atrazine (see Box P), commonly is detected in water throughout the corn-growing areas of the United States.

DISSOLUTION AND EXSOLUTION OF GASES

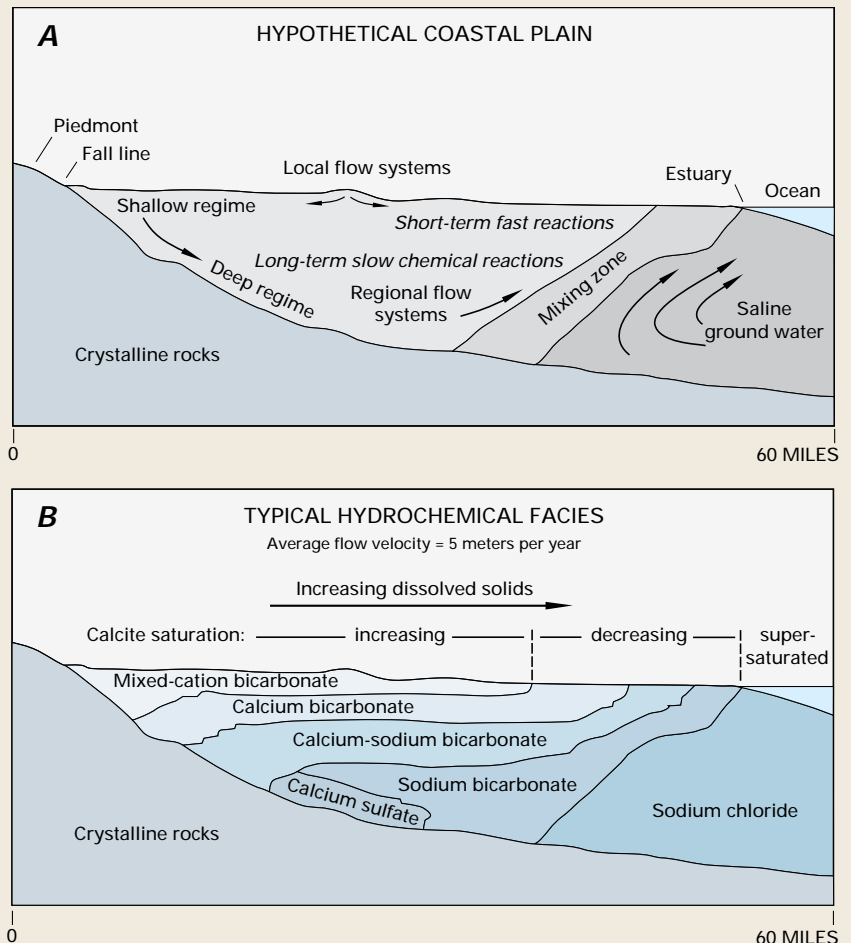
Gases are directly involved in many geochemical reactions. One of the more common gases is carbon dioxide (CO_2). For example, stalactites can form in caves when dissolved CO_2 exsolves (degasses) from dripping ground water, causing pH to rise and calcium carbonate to precipitate. In soils, the microbial production of CO_2 increases the concentration of carbonic acid (H_2CO_3), which has a major control on the solubility of aquifer materials. Other gases commonly involved in chemical reactions are oxygen, nitrogen, hydrogen sulfide (H_2S), and methane (CH_4). Gases such as chlorofluorocarbons (CFCs) and radon are useful as tracers to determine the sources and rates of ground-water movement (see Box G).

Evolution of Ground-Water Chemistry from Recharge to Discharge Areas in the Atlantic Coastal Plain

Changes in the chemical composition of ground water in sediments of the Atlantic Coastal Plain (Figure E-1) provide an example of the chemical evolution of ground water in a regional flow system. In the shallow regime, infiltrating water comes in contact with gases in the unsaturated zone and shallow ground water. As a result of this contact, localized, short-term, fast reactions take place that dissolve minerals and degrade organic material. In the deep regime, long-term, slower chemical reactions, such as precipitation and

dissolution of minerals and ion-exchange, add or remove solutes. These natural processes and reactions commonly produce a predictable sequence of hydrochemical facies. In the Atlantic Coastal Plain, ground water evolves from water containing abundant bicarbonate ions and small concentrations of dissolved solids near the point of recharge to water containing abundant chloride ions and large concentrations of dissolved solids where it discharges into streams, estuaries, and the Atlantic Ocean.

Figure E-1. In a coastal plain, such as along the Atlantic Coast of the United States, the interrelations of different rock types, shallow and deep ground-water flow systems (regimes), and mixing with saline water (A) results in the evolution of a number of different ground-water chemical types (B). (Modified from Back, William, Baedecker, M.J., and Wood, W.W., 1993, *Scales in chemical hydrogeology—A historical perspective*, in Alley, W.M., ed., *Regional Ground-Water Quality*: New York, van Nostrand Reinhold, p. 111–129.) (Reprinted by permission of John Wiley & Sons, Inc.)



Many streams are contaminated. Therefore, the need to determine the extent of the chemical reactions that take place in the hyporheic zone is widespread because of the concern that the contaminated stream water will contaminate shallow ground water (see Box G). Streams offer good examples of how interconnections between ground water and surface water affect chemical processes. Rough channel bottoms cause stream water to enter the streambed and to mix with ground water in the hyporheic zone. This mixing establishes sharp changes in chemical concentrations in the hyporheic zone.

A zone of enhanced biogeochemical activity usually develops in shallow ground water as a result of the flow of oxygen-rich surface water into the subsurface environment, where bacteria and geochemically active sediment coatings are abundant (Figure 19). This input of oxygen to the streambed stimulates a high level of activity by aerobic (oxygen-using) microorganisms if dissolved oxygen is readily available. It is not uncommon for dissolved oxygen to be completely used up in hyporheic flow paths at some distance into the streambed, where anaerobic microorganisms dominate microbial activity. Anaerobic bacteria can use nitrate, sulfate, or other solutes in place of oxygen for metabolism. The result of these processes is that many solutes are highly reactive

in shallow ground water in the vicinity of streambeds.

The movement of nutrients and other chemical constituents, including contaminants, between ground water and surface water is affected by biogeochemical processes in the hyporheic zone. For example, the rate at which organic contaminants biodegrade in the hyporheic zone can exceed rates in stream water or in ground water away from the stream. Another example is the removal of dissolved metals in the hyporheic zone. As water passes through the hyporheic zone, dissolved metals are removed by precipitation of metal oxide coatings on the sediments.

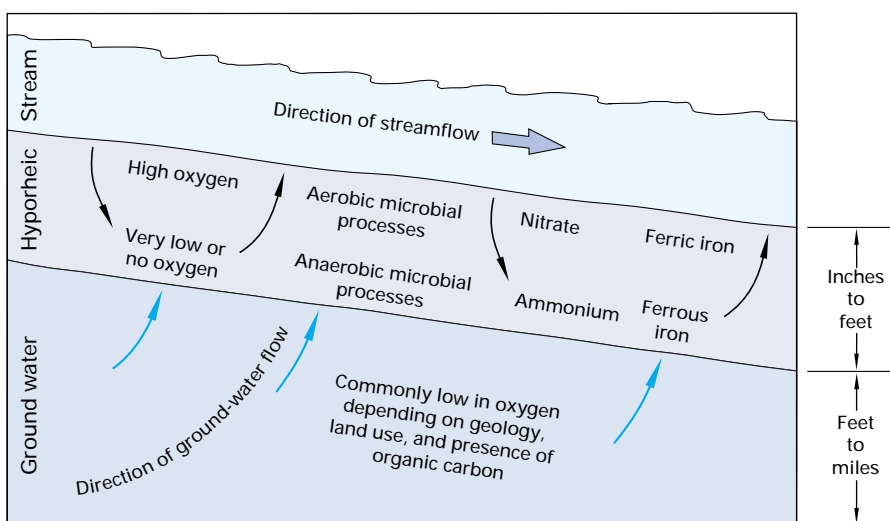


Figure 19. Microbial activity and chemical transformations commonly are enhanced in the hyporheic zone compared to those that take place in ground water and surface water. This diagram illustrates some of the processes and chemical transformations that may take place in the hyporheic zone. Actual chemical interactions depend on numerous factors including aquifer mineralogy, shape of the aquifer, types of organic matter in surface water and ground water, and nearby land use.

The Interface Between Ground Water and Surface Water as an Environmental Entity

In the bed and banks of streams, water and solutes can exchange in both directions across the streambed. This process, termed hyporheic exchange, creates subsurface environments that have variable proportions of water from ground water and surface water. Depending on the type of sediment in the streambed and banks, the variability in slope of the streambed, and the hydraulic gradients in the adjacent ground-water system, the hyporheic zone can be as much as several feet in depth and hundreds of feet in width. The dimensions of the hyporheic zone generally increase with increasing width of the stream and permeability of streambed sediments.

The importance of the hyporheic zone was first recognized when higher than expected abundances of aquatic insects were found in sediments where concentrations of oxygen were high. Caused by stream-water input, the high oxygen concentrations in the hyporheic zone make it possible for organisms to live in the pore spaces in the sediments, thereby providing a refuge for those organisms. Also, spawning success of salmon is greater where flow from the stream brings oxygen into contact with eggs that were deposited within the coarse sediment.



The hyporheic zone also can be a source of nutrients and algal cells to streams that foster the recovery of streams following catastrophic storms. For example, in a study of the ecology of Sycamore Creek in Arizona, it was found that the algae that grew in the top few inches of streambed sediment were quickest to recover following storms in areas where water in the sediments moved upward (Figure F-1).

These algae recovered rapidly following storms because concentrations of dissolved nitrogen were higher in areas of the streambed where water moved upward than in areas where water moved downward. Areas of streambed where water moved upward are, therefore, likely to be the first areas to return to more normal ecological conditions following flash floods in desert streams.

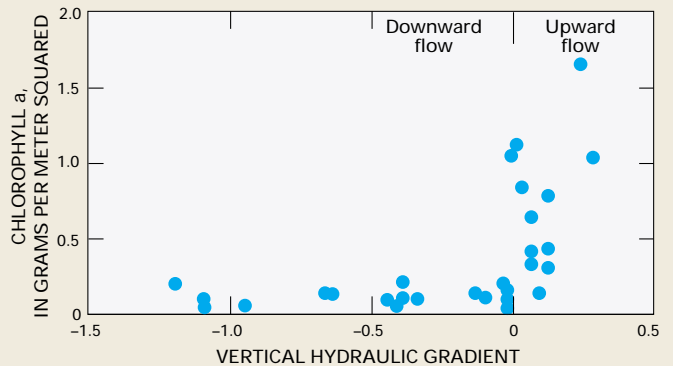


Figure F-1. Abundance of algae in streambed sediments, as indicated by concentration of chlorophyll a, was markedly greater in areas where water moved upward through the sediments than in areas where water moved downward through the sediments in Sycamore Creek in Arizona. (Modified from Valett, H.M., Fisher, S.G., Grimm, N.B., and Camill, P., 1994, Vertical hydrologic exchange and ecologic stability of a desert stream ecosystem: *Ecology*, v. 75, p. 548-560.) (Reprinted with permission.)

Hyporheic zones also serve as sites for nutrient uptake. A study of a coastal mountain stream in northern California indicated that transport of dissolved oxygen, dissolved carbon, and dissolved nitrogen in stream water into the hyporheic zone stimulated uptake of nitrogen by microbes and algae attached to sediment. A model simulation of nitrogen uptake (Figure F-2) indicated that both the physical process of water exchange between the stream and the hyporheic zone and the biological uptake of nitrate in the hyporheic zone affected the concentration of dissolved nitrogen in the stream.

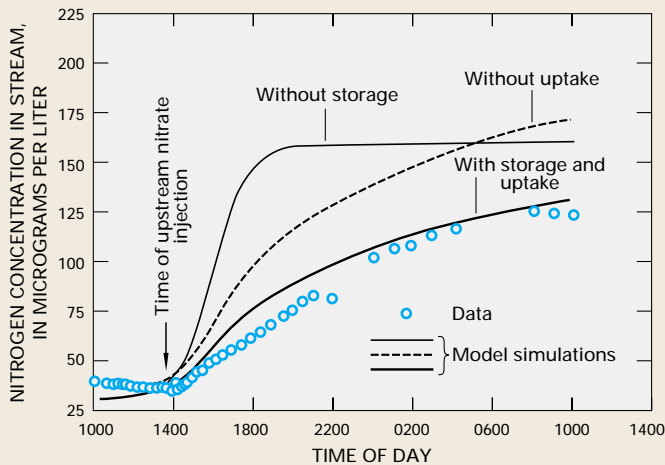


Figure F-2. Nitrate injected into Little Lost Man Creek in northern California was stored and taken up by algae and microbes in the hyporheic zone. (Modified from Kim, B.K.A., Jackman, A.P., and Triska, F.J., 1992, Modeling biotic uptake by periphyton and transient hyporheic storage of nitrate in a natural stream: *Water Resources Research*, v. 28, no. 10, p. 2743–2752.)

The importance of biogeochemical processes that take place at the interface of ground water and surface water in improving water quality for human consumption is shown by the following example. Decreasing metal concentrations (Figure F-3) in drinking-water wells adjacent to the River Glatt in Switzerland was attributed to the interaction of the river with subsurface water. The improvement in ground-water quality started with improved sewage-treatment plants, which lowered phosphate in the river. Lower phosphate concentrations lowered the amount of algal production in the river, which decreased the amount of dissolved organic carbon flowing into the riverbanks. These factors led to a decrease in the bacteria-caused dissolution of manganese and cadmium that were present as coatings on sediment in the aquifer. The result was substantially lower dissolved metal concentrations in ground water adjacent to the river, which resulted in an unexpected improvement in the quality of drinking water.

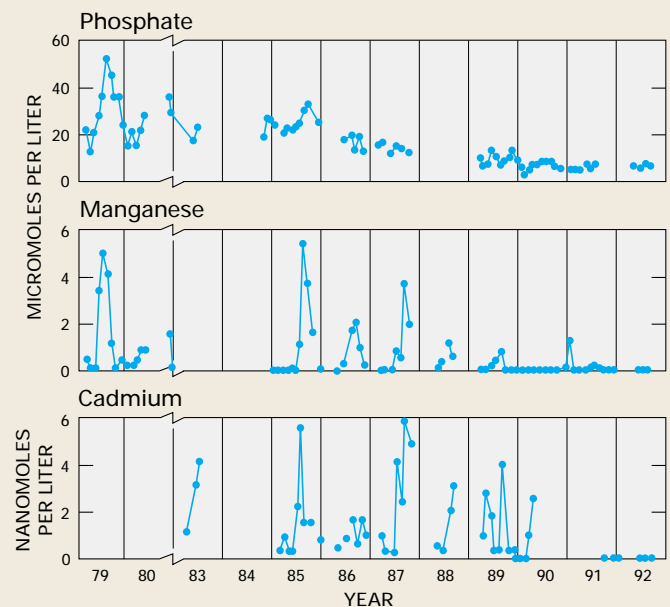
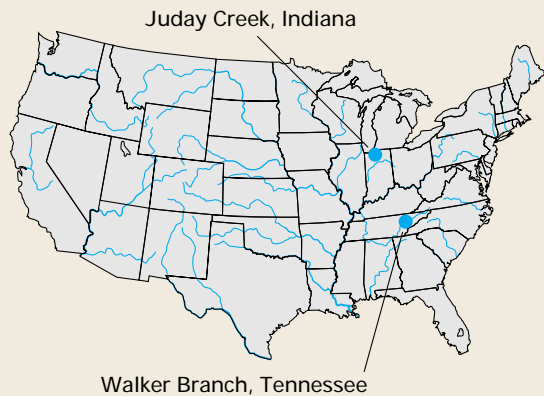


Figure F-3. A decline in manganese and cadmium concentrations after 1990 in drinking-water wells near the River Glatt in Switzerland was attributed to decreased phosphate in the river and hydrologic and biogeochemical interactions between river water and ground water. (Modified from von Gunten, H.R., and Lienert, Ch., 1993, Decreased metal concentrations in ground water caused by controls on phosphate emissions: *Nature*, v. 364, p. 220–222.) (Reprinted with permission from *Nature*, Macmillan Magazines Limited.)

Use of Environmental Tracers to Determine the Interaction of Ground Water and Surface Water

Environmental tracers are naturally occurring dissolved constituents, isotopes, or physical properties of water that are used to track the movement of water through watersheds. Useful environmental tracers include (1) common dissolved constituents, such as major cations and anions; (2) stable isotopes of oxygen (^{18}O) and hydrogen (^2H) in water molecules; (3) radioactive isotopes such as tritium (^3H) and radon (^{222}Rn); and (4) water temperature. When used in simple hydrologic transport calculations, environmental tracers can be used to (1) determine source areas of water and dissolved chemicals in drainage basins, (2) calculate hydrologic and chemical fluxes between ground water and surface water, (3) calculate water ages that indicate the length of time water and dissolved chemicals have been present in the drainage basin (residence times), and (4) determine average rates of chemical reactions that take place during transport. Some examples are described below.



Major cations and anions have been used as tracers in studies of the hydrology of small watersheds to determine the sources of water to streamflow during storms (see Figure G-1). In addition, stable isotopes of oxygen and hydrogen, which are part of water molecules, are useful for determining the mixing of waters from different source areas because of such factors as (1) differences in the isotopic composition of precipitation among recharge areas, (2) changes in the isotopic composition of shallow subsurface water caused by evaporation, and (3) temporal variability in the isotopic composition of precipitation relative to ground water.

Radioactive isotopes are useful indicators of the time that water has spent in the ground-water system. For example, tritium (^3H) is a well-known radioactive isotope of hydrogen that had peak concentrations in precipitation in the mid-1960s as a result of above-ground nuclear-bomb testing conducted at that time. Chlorofluorocarbons (CFCs), which

are industrial chemicals that are present in ground water less than 50 years old, also can be used to calculate ground-water age in different parts of a drainage basin.

^{222}Rn Radon is a chemically inert, radioactive gas that has a half-life of only 3.83 days. It is produced naturally in ground water as a product of the radioactive decay of ^{226}Ra radium in uranium-bearing rocks and sediment. Several studies have documented that radon can be used to identify locations of

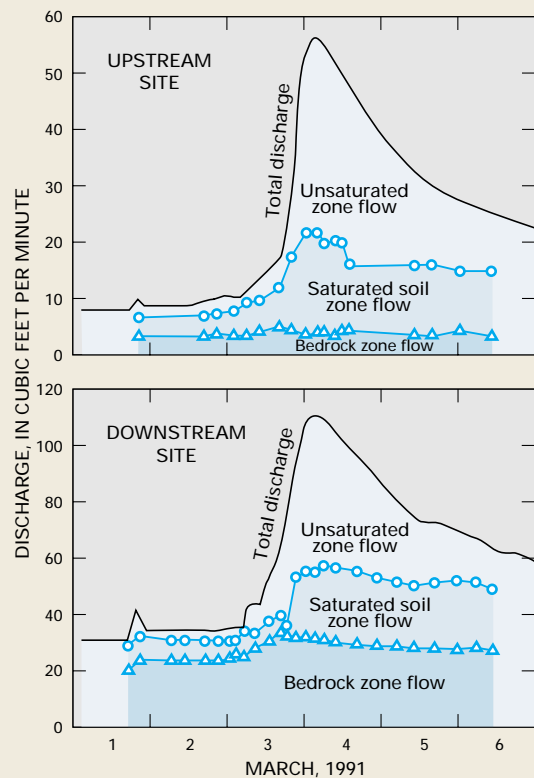


Figure G-1. The relative contributions of different subsurface water sources to streamflow in a stream in Tennessee were determined by analyzing the relative concentrations of calcium and sulfate. Note that increases in bedrock zone (ground water) flow appear to contribute more to the stormflow response at the downstream site than to the stormflow response at the upstream site in this small watershed. (Modified from Mulholland, P.J., 1993, *Hydrometric and stream chemistry evidence of three storm flowpaths in Walker Branch Watershed: Journal of Hydrology*, v. 151, p. 291–316.) (Reprinted with permission from Elsevier Science-NL, Amsterdam, The Netherlands.)

significant ground-water input to a stream, such as from springs. Radon also has been used to determine stream-water movement to ground water. For example, radon was used in a study in France to determine stream-water loss to ground water as a result of ground-water withdrawals. (See Figure G-2.)

An example of using stream-water temperature and sediment temperature for mapping gaining and losing reaches of a stream is shown in Figure G-3. In gaining reaches of the stream, sediment temperature and stream-water temperature are markedly different. In losing reaches of the stream, the diurnal fluctuations of temperature in the stream are reflected more strongly in the sediment temperature.

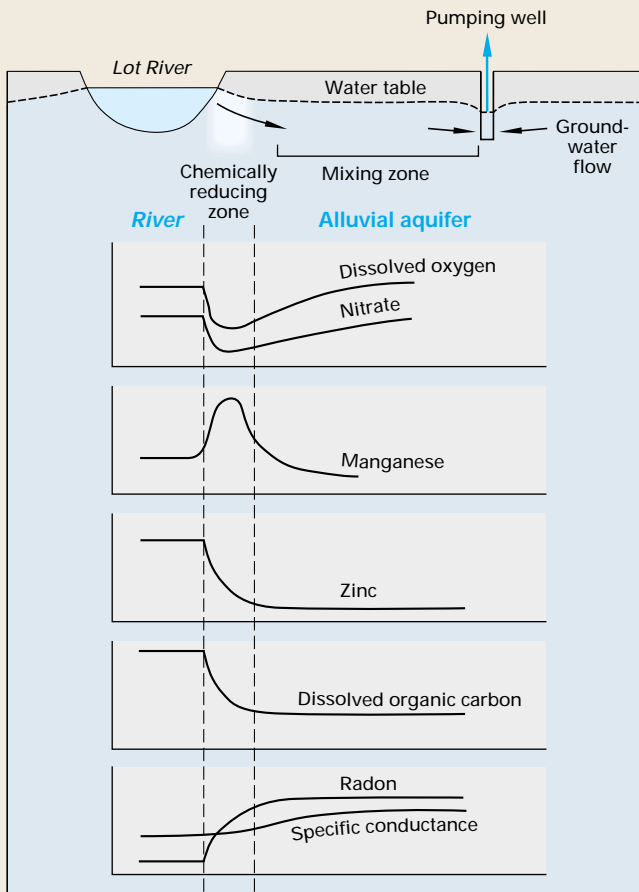


Figure G-2. Sharp changes in chemical concentrations were detected over short distances as water from the Lot River in France moved into its contiguous alluvial aquifer in response to pumping from a well. Specific conductance of water was used as an environmental tracer to determine the extent of mixing of surface water with ground water, and radon was used to determine the inflow rate of stream water. Both pieces of information were then used to calculate the rate at which dissolved metals reacted to form solid phases during movement of stream water toward the pumping well. (Modified from Bourg, A.C.M., and Bertin, C., 1993, *Biogeochemical processes during the infiltration of river water into an alluvial aquifer: Environmental Science and Technology*, v. 27, p. 661-666.) (Reprinted with permission from the American Chemical Society.)

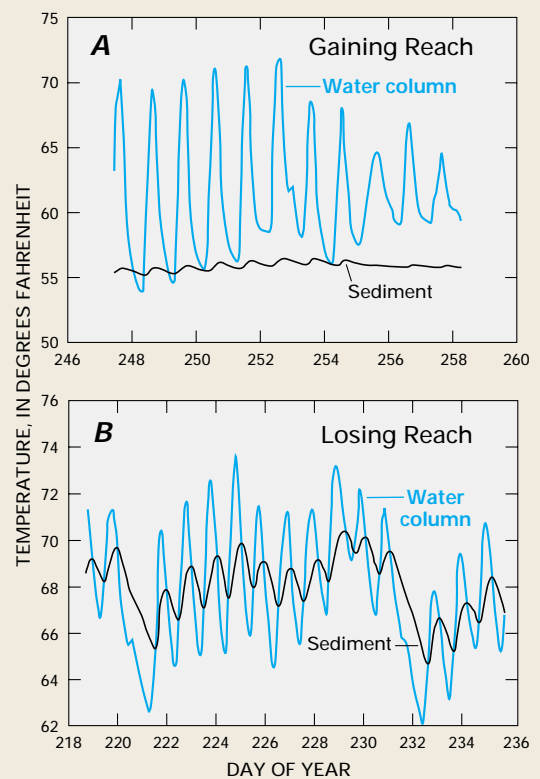


Figure G-3. Ground-water temperatures generally are more stable than surface-water temperatures. Therefore, gaining reaches of Juday Creek in Indiana are characterized by relatively stable sediment temperatures compared to stream-water temperatures (A). Conversely, losing reaches are characterized by more variable sediment temperatures caused by the temperature of the inflowing surface water (B). (Modified from Silliman, S.E., and Booth, D.F., 1993, *Analysis of time series measurements of sediment temperature for identification of gaining versus losing portions of Juday Creek, Indiana: Journal of Hydrology*, v. 146, p. 131-148.) (Reprinted with permission from Elsevier Science-NL, Amsterdam, The Netherlands.)

Lakes and wetlands also have distinctive biogeochemical characteristics with respect to their interaction with ground water. The chemistry of ground water and the direction and magnitude of exchange with surface water significantly affect the input of dissolved chemicals to lakes and wetlands. In general, if lakes and wetlands have little interaction with streams or with ground water, input of dissolved chemicals is mostly from precipitation; therefore, the input of chemicals is minimal. Lakes and wetlands that have a considerable amount of ground-water inflow generally have large inputs of dissolved chemicals. In cases where the input of dissolved nutrients such as phosphorus and nitrogen exceeds the output, primary production by algae and wetland plants is large. When this large amount of plant material dies, oxygen is used in the process of decomposition. In some cases the loss of oxygen from lake water can be large enough to kill fish and other aquatic organisms.

The magnitude of surface-water inflow and outflow also affects the retention of nutrients in wetlands. If lakes or wetlands have no stream outflow, retention of chemicals is high. The tendency to retain nutrients usually is less in wetlands that are flushed substantially by throughflow of surface water. In general, as surface-water inputs increase, wetlands vary from those that strongly retain nutrients to those that both import and export large amounts of nutrients. Furthermore, wetlands commonly have a significant role in altering the chemical form of dissolved constituents. For example, wetlands that have throughflow of surface water tend to retain the chemically oxidized forms and release the chemically reduced forms of metals and nutrients.

“The chemistry of ground water and the direction and magnitude of exchange with surface water significantly affect the input of dissolved chemicals to lakes and wetlands”
