# Water-Quality Principles
# QW1022

## Lesson 2—Groundwater as an Element in the Hydrological Cycle

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**U.S. Department of the Interior**

**U.S. Geological Survey**

# 145: Groundwater as an Element in the Hydrological Cycle

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*Groundwater commonly is portrayed in diagrams of the hydrological cycle as a single pool of water with simple transfers of water to and from the land surface, oceans, and atmosphere. In reality, groundwater is an integral part of a complex hydrological cycle that involves the continuous movement of water on Earth. We begin with a global perspective on the importance of groundwater in the hydrological cycle and then address the importance of understanding the dynamics of groundwater-flow systems. We next turn our attention to groundwater budgets and the effects of humans and climate on these budgets. The final section is devoted to a discussion of interactions between groundwater and surface water, and we highlight the importance of these interactions in the hydrological cycle.*

## INTRODUCTION

Groundwater occurs almost everywhere beneath the land surface. The widespread occurrence of potable groundwater is a major reason it is used as a source of water supply worldwide. Groundwater also plays a crucial role in sustaining streamflow during dry periods and is vital to many lakes and wetlands. Moreover, many plants and aquatic animals depend greatly upon the groundwater that discharges to streams, lakes, and wetlands.

Groundwater commonly is portrayed in diagrams of the hydrological cycle as a single pool of water with simple transfers of water to and from the land surface, oceans, and atmosphere. In reality, groundwater is an integral part of a complex hydrological cycle that involves the continuous movement of water on Earth. In this article, we begin with a global perspective on the importance of groundwater in the hydrological cycle. The importance of understanding the dynamics of groundwater-flow systems is then addressed. We next turn our attention to groundwater budgets and the effects of humans and climate on these budgets. The final section is devoted to a discussion of interactions between groundwater and surface water, and we highlight the importance of these interactions in the hydrological cycle.

## GROUNDWATER IN THE HYDROLOGICAL CYCLE OF THE EARTH

The volume of water contained in groundwater storage often is compared to other major global pools of water within the Earth's hydrological cycle. Published estimates of the volumes of water in major global pools are shown in Table 1. While the estimates of the volume of water in the Earth's oceans and atmosphere shown in Table 1 are relatively similar, the estimates for groundwater storage vary greatly. In part, this variability is due to different considerations of depth and salinity in defining the global groundwater pool, and, in part, the variability reflects less knowledge about groundwater than other global pools of water. The estimates of water contained as soil moisture also illustrate considerable uncertainty.

Note that early estimates by Kalle (1945) of the global groundwater pool greatly underestimated its volume. Kalle's estimates are shown for historical purposes, but are not considered further in the discussion below.

The two major pools of freshwater are groundwater and water contained in glaciers and polar ice. As shown in Table 1, and depending on its definition, the groundwater pool may constitute from less than one-third to more

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**Table 1** Volume of water (in thousand km$^3$) attributed to different parts of the world water balance (Significant figures largely retained from original sources)

| | Source | | | | |
|---|---|---|---|---|---|
| Water form | Kalle (1945)[f] | Nace (1967)[g] | Voskresensky (1978)[h] | L'vovich (1979)[i] | Schlesinger (1997)[j] |
| Ocean | 1 372 000 | 1 320 000 | 1 338 000 | 1 370 332 | 1 350 000 |
| Glaciers, Polar ice | 16 500 | 29 200 | 24 064 | 24 000 | 33 000 |
| Permafrost ice | – [a] | – | 300 | – | – |
| Groundwater | 250 | 8350[b] | 23 400 | 60 000 | 15 300 |
| *Fresh* | – | – | *10 530* | *4000*[c] | – |
| *Other* | – | – | *12 870* | – | – |
| Lakes | 250[d] | 229 | 176.4 | 280 | – |
| *Fresh* | – | *125* | *91* | *150* | – |
| *Saline* | – | *104*[e] | *85.4* | *125* | – |
| *Reservoirs* | – | – | – | *5* | – |
| Wetlands | – | – | 11.47 | – | – |
| Soil moisture | – | 67 | 16.5 | 83 | 122 |
| Rivers | [d] | 1.25 | 2.12 | 1.2 | 40 |
| Biota | – | – | 1.12 | – | 1 |
| Atmosphere | 13 | 13 | 12.9 | 14 | 13 |
| Total | 1 389 013 | 1 357 860 | 1 385 984 | 1 454 710 | 1 398 476 |

[a]Presence of hyphen indicates no data provided.
[b]Groundwater to a depth of 4000 m.
[c]Groundwater that is actively exchanged in hydrological cycle.
[d]Value for lakes includes lakes and rivers.
[e]Includes inland seas.
[f]Data from table on page 31 of Kalle (1945) were presented in kg cm$^{-2}$ of the Earth's surface area. These were converted to the values shown here using an estimate of the Earth's surface area of $510.1 \times 10^6$ km$^2$ from page 1 of Kalle and based on the specific gravity of water of 1.0 g per cm$^3$ at 4 °C.
[g]Data from Table 1 on page 2.
[h]Data from Table 9 on page 43.
[i]Data from Table 1 on page 15 and Table 2 on page 21 and text in Chapter 1.
[j]Data from Table 2.2 and Figure 10.1 on page 346.

than twice the volume of water contained in glaciers and polar ice.

Table 2 focuses on the global freshwater resource and shows the relative amounts of groundwater and other sources of freshwater (ignoring water frozen in glaciers and polar ice). Despite uncertainty in the values, it is clear that groundwater is the major pool of nonfrozen freshwater on Earth, likely composing more than 95% of this resource. Lakes and reservoirs contain much of the remaining nonfrozen (available) freshwater with considerable uncertainty about the volume contained as soil moisture. Rivers contain only a tiny fraction of the available freshwater.

Comparisons also can be made in the annual global fluxes of water as indicated in Table 3. The values for groundwater discharge to oceans shown in Table 3 include only estimates for direct discharge to bays, estuaries, and oceans. Groundwater also reaches coastal waters indirectly by discharge to streams and rivers that drain to coastal waters. The values for runoff to oceans shown in Table 3 include this indirect groundwater discharge to streams and rivers. For example, L'vovich (1979) estimates that 12 of the 41 thousand km$^3$ per year of runoff to oceans consists of groundwater discharge to streams and rivers.

The residence time of groundwater can vary from a few days to more than 10 000 years. It is difficult, however, to estimate an average global residence time for groundwater, because of the previously noted complexities in defining the global groundwater pool. In comparison, the average residence time for continental runoff in free-flowing rivers likely varies between 16 and 26 days, and regulation by reservoirs has increased the mean age of continental runoff to slightly longer than a month (Vörösmarty and Sahagian, 2000).

Some discussions of the global groundwater pool distinguish between groundwater in "active" exchange with the Earth's surface (commonly referred to as renewable groundwater) and "fossil" water accumulated over tens of thousands of years under different climatic conditions, which, once used, cannot be replenished readily (sometimes referred to as nonrenewable groundwater). Commonly, it is difficult to describe the two categories of groundwater reserves quantitatively and to draw boundaries between them.

Water withdrawals by humans are superimposed upon the natural pools and fluxes of groundwater. Groundwater withdrawals are large and have increased greatly during the past 50 years. Perhaps as many as two billion people depend directly upon groundwater for drinking water, and 40% of the world's food is produced by irrigated agriculture that relies largely on groundwater (United Nations Environment

**Table 2** Percent of total nonfrozen freshwater resources attributed to different components of the world water balance. Percent values were derived from volumes presented in Table 1

| | Source | | | |
|---|---|---|---|---|
| Freshwater form | Nace (1967) | Voskresensky (1978) | L'vovich (1979) | Schlesinger (1997) |
| Groundwater | 97.74 | 98.86 | 94.31[a] | 98.95 |
| Lakes | 1.46 | 0.85 | 3.54 | – |
| Reservoirs | – | – | 0.12 | – |
| Wetlands | – | 0.11 | – | – |
| Soil moisture | 0.78 | 0.15 | 1.96 | 0.79 |
| Rivers | 0.01 | 0.02 | 0.03 | 0.26 |
| Total | 100 | 100 | 100 | 100 |

[a]Groundwater that is actively exchanged in hydrological cycle.

**Table 3** Annual water fluxes (in thousand $km^3$ per year) between the atmosphere, land, and ocean, and between land and the ocean

| | Source | | | | | |
|---|---|---|---|---|---|---|
| Water flux form | Hutchinson (1957)[a] | Nace (1967)[b] | Budyko and Sokolov (1978)[c] | L'vovich (1979)[d] | Chahine (1992)[e] | Schlesinger (1997)[f] |
| Evaporation | | | | | | |
| From ocean surfaces | 383 | 350 | 505 | 452.6 | 434 | 425 |
| From land surfaces | 63 | 70 | 72 | 72 | 71 | 71 |
| Precipitation | | | | | | |
| On ocean surfaces | 347 | 320 | 458 | 411.6 | 398 | 385 |
| On land surfaces | 99 | 100 | 119 | 113.5 | 107 | 111 |
| Runoff/Discharge | | | | | | |
| Runoff to oceans | 36 | 38[g] | 44.7 | 41 | 36 | 40 |
| Direct groundwater discharge to oceans | – | 1.6[h] | 2.2 | 2.2 | – | – |

[a]Data from Table 16 calculated from amounts published in Kalle (1945).
[b]Data from Table 1 on page 2.
[c]Data from Table 189 on page 590.
[d]Data from Table 1 on page 15 and Table 2 on page 21 and text in Chapter 1; groundwater discharge value from page 30.
[e]5 Data from Figure 1.
[f]Data from Figure 10.1 on page 346.
[g]From rivers and icecaps.
[h]Arbitrarily set to about 5% of runoff.

Programme, 2003). Unfortunately, despite the importance of groundwater, estimates of its use in many countries are either unavailable or of poor quality.

Finally, it is worth noting that groundwater depletion may have affected global water balances enough to contribute to sea-level rise during the past century as a result of water pumped from wells that returns to the sea by runoff or evaporation/precipitation (Sahagian *et al.*, 1994). Extensive drainage of wetlands during the past century also may have contributed to sea-level rise.

## GROUNDWATER-FLOW SYSTEMS

The three-dimensional body of Earth material saturated with moving groundwater that extends from areas of recharge to areas of discharge is referred to as a groundwater-flow system (Figure 1). Perhaps the most obvious source of water to groundwater-flow systems (or more simply referred to as groundwater systems) is areal recharge from precipitation. Water from areal recharge flows from the water table where the recharge enters the saturated groundwater system through the flow system to the discharge area (which is a stream in Figure 1). Surface water features such as streams, wetlands, and lakes can either be the location of water entering the groundwater system (recharge area) or the location of water leaving the groundwater system (discharge area).

The areal extent of groundwater-flow systems varies from a few square kilometers or less to tens of thousands of square kilometers. The lengths of groundwater-flow paths range from a few meters to tens, and sometimes hundreds, of kilometers. A deep groundwater-flow system with long flow paths between areas of recharge and discharge may be overlain by, and in hydraulic connection with, several shallow, more local, flow systems.

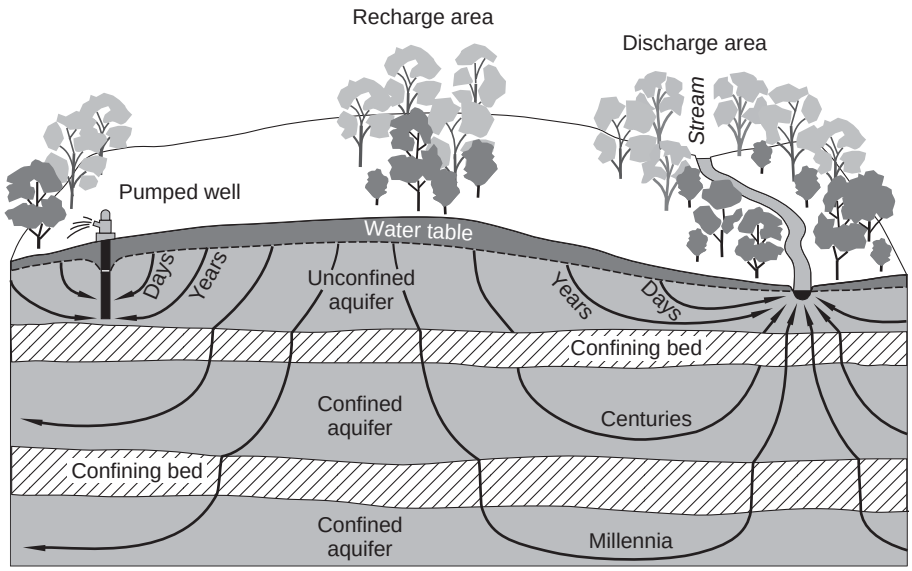


**Figure 1** Groundwater system showing generalized flow paths of groundwater movement and the relative age of the water since the time of recharge. (Modified from Heath, 1983 and Winter *et al.*, 1998)

## Factors that control Groundwater Movement

Two types of factors control groundwater movement: factors within the porous media and factors along the boundaries of the porous media where the water enters and leaves the groundwater system.

The equation that describes the movement of groundwater within a porous media is known as *Darcy's law*. In simplified terms, the average linear velocity (Freeze and Cherry, 1979) of the water is calculated from Darcy's law as:

$$v = -\left(\frac{K}{n}\right) i \qquad (1)$$

where $v$ is the average linear velocity (referred to as velocity for the remainder of the text) of the groundwater ($LT^{-1}$), $K$ is the hydraulic conductivity ($LT^{-1}$), $n$ is the effective porosity (dimensionless), and $i$ is the hydraulic gradient (the change in head per unit of distance; dimensionless), which is a negative number in the direction of flow. As a result, the hydraulic conductivity, effective porosity, and hydraulic gradient are all important in the determination of the movement of groundwater. The hydraulic conductivity, which represents the ability of the geologic framework to transmit water, is a property of the porous media and the fluid contained therein. The hydraulic conductivity of a groundwater system can vary over many orders of magnitude (Figure 2). The larger the hydraulic conductivity of a porous media, the easier it is for water to flow through it. The porosity is the ratio of the volume of the voids divided by the total volume. The effective porosity is the volume of the voids that are interconnected and available for fluid transmission divided by the total volume. The hydraulic gradient is the change in static head per unit of distance in a given direction (Lohman *et al.*, 1972). The static head generally is measured as the water level in a well. The water-level distribution or potentiometric surface describes the hydraulic gradient within an aquifer, which, in turn, determines the direction and rate of groundwater flow. For groundwater systems in cavernous karst terrain or in fractured-rock systems, the validity of Darcy's law, which was developed for porous media, may not be strictly applicable (White, 1993), and other methods for determining the velocity distribution may be required.

Darcy's law describes the rates and directions that water will flow in a groundwater system. The other important factors that are necessary to adequately understand the movement of water in a groundwater system are the sources of water to the system and the stresses (including water use) on the system. The external boundaries of the groundwater system, such as streams and areal recharge, affect the paths of water movement and the hydraulic head distribution (Reilly, 2001). The location of wells and water use (for example, irrigation) also impact where water moves within the groundwater system.

## Time of Travel in a Groundwater Flow System

The age (time since recharge) of groundwater varies in different parts of groundwater-flow systems. The age of groundwater increases along a particular flow path through the groundwater-flow system from an area of recharge to

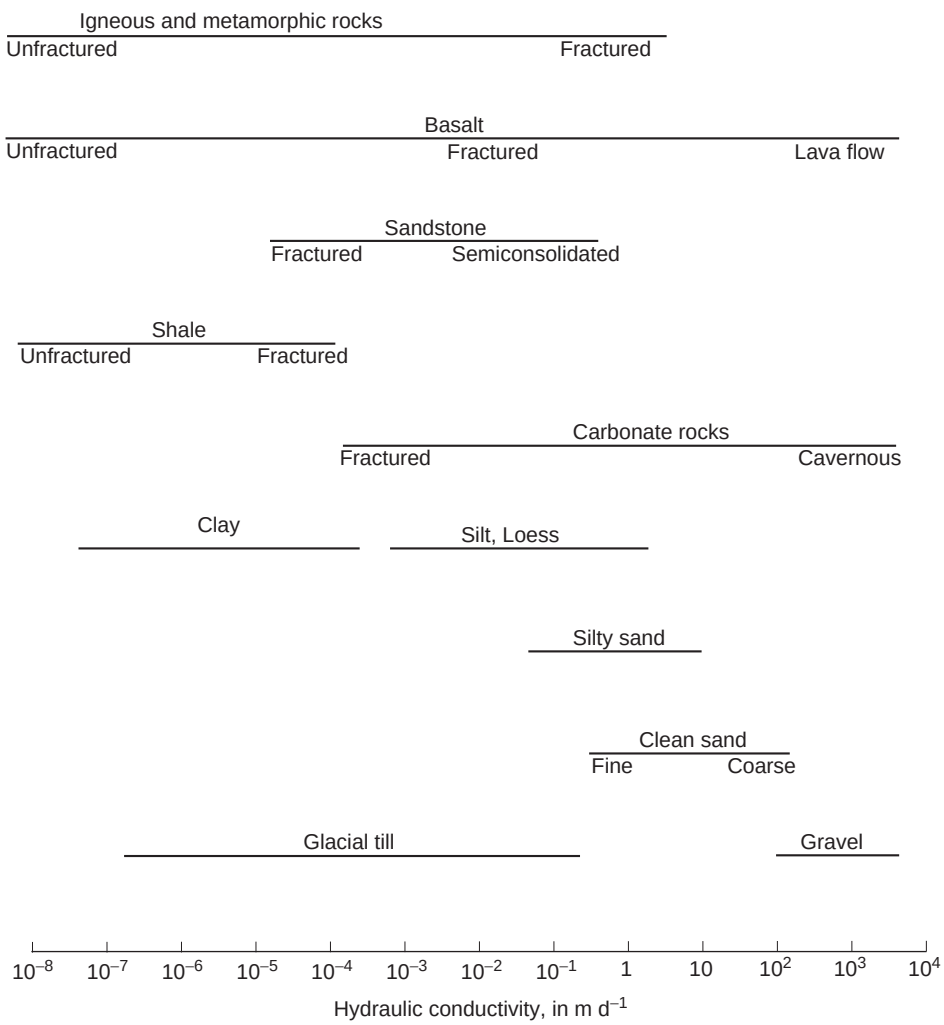


**Figure 2** Range of hydraulic conductivity for selected rock types (Modified from Heath, 1983)

an area of discharge. In general, Darcy's law determines the flow velocity, and, therefore, the time of travel. Travel times within groundwater systems can vary considerably (Figure 1). In shallow flow systems, ages of groundwater at areas of discharge can vary from less than a day to a few hundred years. In deep, regional flow systems with long flow paths (tens of kilometers), ages of groundwater may reach thousands or tens of thousands of years or more. The shallower and younger groundwater tends to be more vulnerable to contamination from human activities at the land surface (Focazio *et al.*, 2002).

Fractured-rock systems in bedrock usually have smaller effective porosities than unconsolidated porous media such as sands and gravels, and flow velocities through fractured rock can be relatively fast. For example, travel times of water over distances of several kilometers have been estimated at less than a year for municipal wells completed in fractured dolomite in Wisconsin (Rayne *et al.*, 2001). In such cases, seasonal variations in recharge and pumping affect the variability in travel times. In more sluggish groundwater systems, such as the Bangkok Basin in Thailand (Sanford and Buapeng, 1996), long-term climate and geologic change need to be considered in understanding the movement of groundwater over tens of thousands of years.

Tracer techniques have been applied widely to estimate the residence time of subsurface waters, as well as the amounts and timing of recharge and discharge (Cook and Böhlke, 1999). Most tracer techniques require knowledge (or assumption) of the time history of tracer input at the land surface or the water table. This temporal pattern then is correlated to a concentration-depth pattern in the subsurface at a point in time. Other approaches use information on decay products to determine age. Tracers can occur naturally (chloride, heat, the stable isotopes $^{2}H$ and $^{18}O$), occur in the atmosphere as a result of human activities (tritium, $^{36}Cl$, CFCs), or be applied intentionally on the land surface (fertilizers, pesticides). Isotopes of elements,

such as radon, dissolved from host rocks also can be used to estimate residence times and interactions with surface water. Over the past decade, research in age dating and tracking young groundwater (<50 years old) using multiple tracers has resulted in major breakthroughs in understanding the dynamics of groundwater systems.

The time required for water to move through a groundwater system also can be estimated for some systems on the basis of aquifer storage volumes. A three-dimensional groundwater system contains a specific volume of water in the pores and cracks of the media. This volume has been referred to here as the volume in storage or the "pool" of water in storage (V in units of $L^3$). The rate at which this volume is replaced is related to the rate of recharge entering the system that displaces the water in storage; this is the flux of water entering the system (Q in units of $L^3T^{-1}$). In an average (or simplified) sense that does not take into account differences in flow paths and the nature of the framework, the time required for water to move through the entire system is simply $VQ^{-1}$. This approach may yield a poor approximation in complex flow systems, such as karst or fractured rock.

## GROUNDWATER BUDGETS

Water budgets are used widely to account for flow and storage changes in various hydrological systems, including rivers, lakes, drainage basins, the land surface, and groundwater systems. In its most basic form, a water budget is expressed simply as:

$$\text{Inflow} - \text{Outflow} = \text{Change in Water Storage} \qquad (2)$$

Each of the three main terms in the water-budget equation can include both natural (e.g. precipitation) and human-induced (e.g. imported water) components. Quantities in the budget apply to a predefined volume and can either be flow rates [$L^3T^{-1}$] or volumes [$L^3$] for a specified time period.

Understanding water budgets for groundwater systems is both critically important and challenging. Groundwater is hidden from view, and groundwater divides do not necessarily coincide with surface-water divides (Winter *et al.*, 2003). Several aquifers may underlie the land surface at any given location with varying hydraulic properties and locations of recharge.

Under natural (predevelopment) conditions, a groundwater system is in long-term equilibrium. That is, averaged over some period of time (and in the absence of climate change), the amount of water entering or recharging the system is approximately equal to the amount of water leaving or discharging from the system. Inflows (recharge) to a groundwater system under natural conditions include areal recharge from precipitation on the land surface and recharge

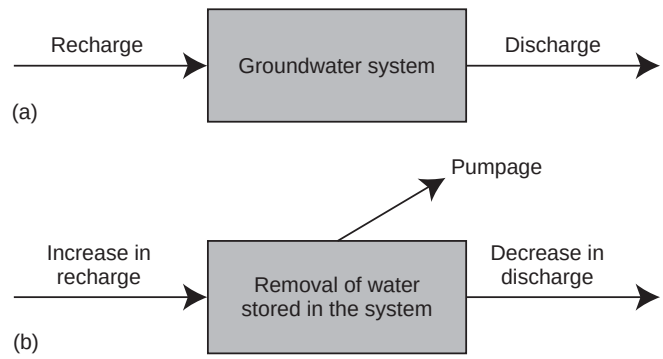


**Figure 3** (a) Predevelopment water-budget diagram illustrating that inflow equals outflow. (b) Water-budget diagram showing changes in flow for a groundwater system being pumped. The sources of water for the pumpage are changes in recharge, discharge, and the amount of water stored. (Alley *et al.,* 1999)

from losing streams, lakes, and wetlands. Outflows (discharge) from a groundwater system under natural conditions include discharge to surface-water bodies and groundwater evapotranspiration. Because the system is in equilibrium, the quantity of water stored is constant or varies about some average condition in response to annual or longer-term climatic variations. This predevelopment water budget is shown schematically in Figure 3(a).

### Effects of Human Activities on Groundwater Budgets

Withdrawals of groundwater by pumping change the natural (or predevelopment) flow system. Water that is withdrawn must be supplied by some combination of (i) more water entering the groundwater system (increased recharge), (ii) less water leaving the system (decreased discharge), and (iii) removal of water that was stored in the system. This statement, illustrated in Figure 3(b), can be written in terms of rates (or volumes over a specified period of time) as:

$$\text{Pumpage} = \text{Increased recharge} + \text{Water removed from storage} + \text{Decreased discharge} \qquad (3)$$

That is, the water pumped from the system must come from some change of flows and from removal of water stored in the predevelopment system (Theis, 1940; Lohman, 1972). Note that the initial predevelopment water-budget values do not enter directly into the budget calculation.

Regardless of the amount of water withdrawn, the system will undergo some drawdown in water levels in pumping wells to induce the flow of water to these wells, which means that some water initially is removed from storage. For most groundwater systems, the change in storage in response to pumping is a transient phenomenon that occurs as the system readjusts to the pumping stress. The relative

contributions of changes in storage, changes in recharge, and changes in discharge evolve with time. If the system can come to a new equilibrium, the changes in storage will diminish to zero and inflows will again balance outflows:

$$\text{Pumpage} = \text{Increased recharge} + \text{Decreased discharge} \tag{4}$$

Thus, the long-term source of water to discharging wells is typically a change in the amount of water entering or leaving the system.

The time that is required to bring a hydrologic system into equilibrium depends on the rate at which the discharge can be captured, which is a function of the characteristics of the aquifer system and the placement of pumping wells (Lohman, 1972). In many circumstances, the dynamics of the groundwater system are such that an approximate equilibrium condition would not be reached for decades or even centuries.

A prevalent view among many hydrologists and nonhydrologists alike is that the development of a groundwater system is considered to be "safe" if the rate of groundwater withdrawal does not exceed the rate of natural recharge. Bredehoeft *et al.* (1982) referred to this concept as the "Water-Budget Myth". It is an oversimplification of the information that is needed to understand the effects of developing a groundwater system. Natural recharge is a critical element for understanding water budgets, is an essential parameter in the modeling of aquifer systems, and its estimation is fundamental to understanding contaminant transport from the land surface. Nonetheless, an estimate of natural recharge, by itself, is of limited value in determining the amount of groundwater that can be withdrawn on a sustained basis. How much groundwater is available for use depends much more upon how changes in inflow and outflow, which result from pumping, affect the surrounding environment and the acceptable trade-off between groundwater use and these changes. Achieving this trade-off in the long term is a central theme in the evolving concept of sustainability (Alley *et al.*, 1999; Sophocleous, 2000; Alley and Leake, 2004).

In determining the effects of pumping and the amount of water available for use, it is critical to recognize that not all the water pumped is necessarily consumed (Kendy, 2003). For example, some of the water pumped for irrigation is consumed by evapotranspiration and some of the water returns to the groundwater system by infiltration, canal leakage, and other means of irrigation return flow. Most other uses of groundwater are similar, in that some of the water pumped is not consumed, but is returned to the system.

Although withdrawal of groundwater is the most direct way in which humans affect groundwater budgets, many other human activities commonly must be considered, particularly those that affect recharge rates. Changes in the patterns and rates of recharge can be caused by irrigation and urban development, removal or changes in the type of vegetation, changes in surface-water flows and storage, and land drainage. Identifying human practices that influence recharge is straightforward, but quantifying the effects of these practices is very difficult. For example, built-up and paved areas promote runoff and inhibit infiltration. The enhanced runoff, however, may be channeled to a retention basin or infiltration gallery, resulting in the relocation of recharge areas and the transition from slow, diffuse recharge to rapid, localized recharge (Lerner, 2002). Wastewater infiltration also is often a major component of overall recharge to aquifers around urban areas, especially in more arid climates (Foster and Chilton, 2004).

Artificial recharge with excess surface water or reclaimed wastewater is increasing in many areas, thus becoming an increasingly important component of the hydrological cycle. In fact, with long-term intensive use, much of the local groundwater may be derived from artificial recharge (Shelton *et al.*, 2001)

Examples of large-scale changes that can occur in the water budgets of regional groundwater systems are illustrated by examining the simulated sources of water (increased recharge, decreased discharge, and changes in storage) that supplied withdrawals from nine major groundwater systems in the United States. These results, shown in Figure 4, are from numerical simulations of major regional aquifer systems for the period of development up through the mid to latter 1980s. The results illustrate large variability in the relative proportions of different sources of water to pumping wells. The Floridan and Edwards–Trinity aquifer systems, which equilibrate rapidly after pumping, were simulated as steady state with no long-term change in storage for the simulation period. In contrast, the Southern High Plains, where most natural discharge occurs far from pumping wells, and the deeply buried Great Plains aquifer system had substantial changes in groundwater storage. In some areas, such as the California Central Valley and Eastern Snake River Plain, return flow of excess irrigation water was the major change in the water budget during the simulation period. Note that the distinction between changes in recharge and changes in discharge commonly is a function of how the system was defined (i.e. a gain to one system may result in a loss from an adjoining system). For example, groundwater withdrawals from confined aquifers (Northern Atlantic Coastal Plain and Gulf Coastal Plain) can cause flow to be diverted (recharged) into the deeper regional flow regime that would otherwise discharge to streams in the outcrop areas.

## Effects of Climate on Groundwater Budgets

Climate affects groundwater budgets in several ways. Most directly, climate influences the rates and distribution of

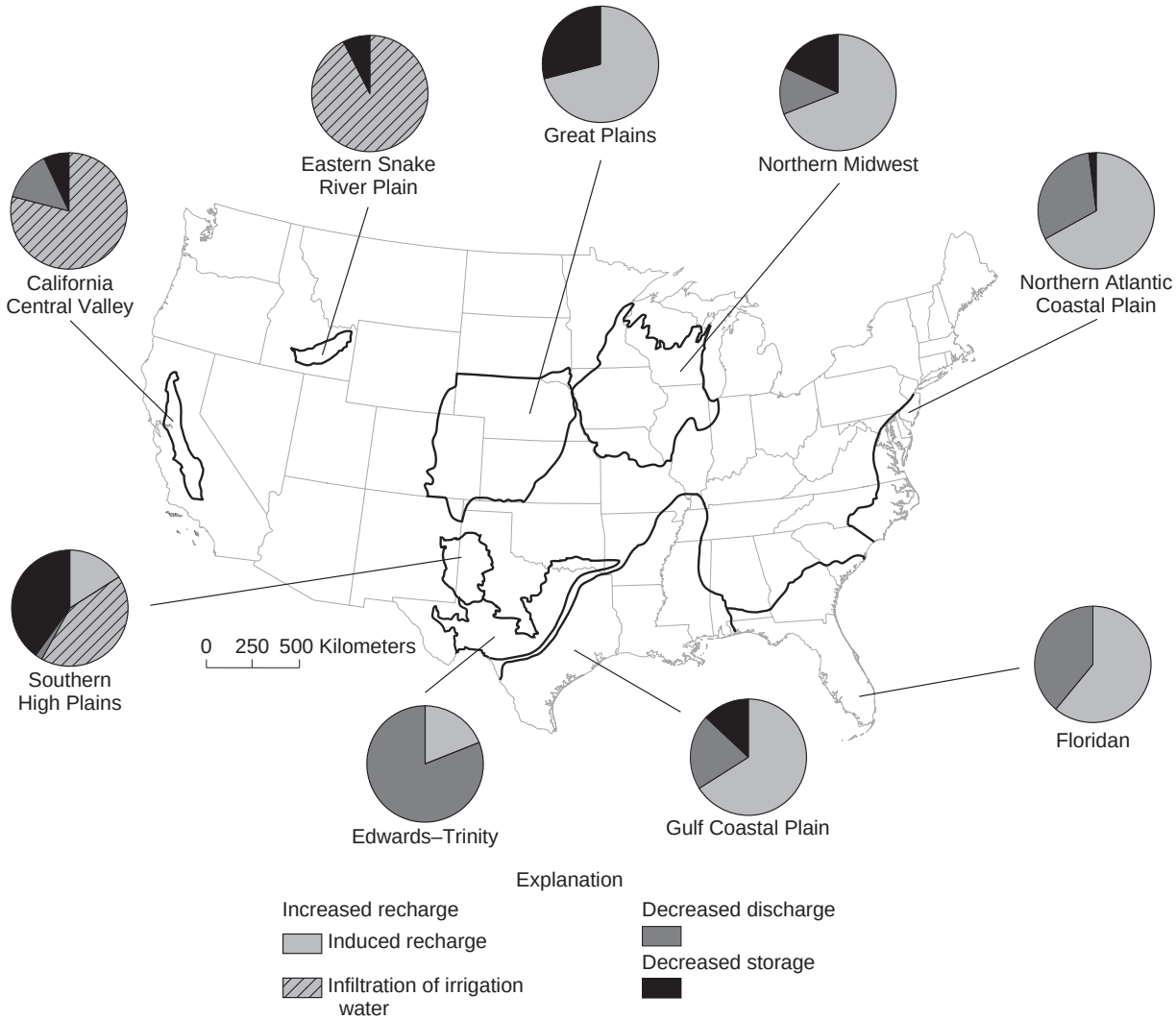


**Figure 4** Sources of water that supply withdrawals from major aquifer systems in the United States based on model simulations for various periods through the 1970s and 1980s. The results illustrate the variety of ways in which overall groundwater budgets can change in response to large-scale pumping. The simulations of the aquifer systems were done at different stages in their development, and, in many cases, are not representative of today's conditions given the dynamic nature of groundwater systems. (Data from Johnston, 1999; Modified from Alley *et al.,* 2002)

recharge. Climate also affects human demands for groundwater and affects plant transpiration from shallow groundwater in response to changing inputs of solar energy and changing depths to the water table. Some groundwater responses to climate variability and change may show considerable temporal lag, given the relatively long response times of groundwater systems. For example, long-term trends in the balance between precipitation and evapotranspiration – caused by either long-term variability or by anthropogenic global change – may affect groundwater discharge, but their effects may be attenuated and spread out over time. As an indication of the effects of long-term climate changes on groundwater systems, much of the water pumped today in arid regions comes from aquifers that were recharged at higher rates during wetter or cooler conditions in the past (e.g. Zhu *et al.*, 1998).

Long-term droughts may be viewed as a natural stress on groundwater systems that in many ways have effects similar to groundwater withdrawals; namely, reductions in groundwater storage and accompanying reductions in groundwater discharge to streams and other surface-water bodies. Because climate stress on the hydrologic system is added to the existing or projected human-derived stress, droughts represent extreme hydrologic conditions that may be important in long-term groundwater management.

Human-induced climate change in the coming decades may affect groundwater resources in several ways, including (i) long-term changes in groundwater recharge resulting

from changes in the annual and seasonal distribution of precipitation and temperature, (ii) more severe and longer lasting droughts, (iii) changes in evapotranspiration resulting from changes in vegetation, and (iv) possible increased demands for groundwater as a backup source of water supply. Surficial aquifers that supply much of the water in streams, lakes, and wetlands are likely to be the part of the groundwater system most sensitive to climate change. Climate not only affects groundwater budgets, but to the extent that groundwater systems influence evapotranspiration and runoff, changes in groundwater fluxes also may affect climate (National Research Council, 2004).

Several studies have used atmospheric circulation models and future scenarios of precipitation and temperature to simulate the effects of climate change on groundwater recharge, storage, and discharge. For example, York *et al.* (2002) illustrate the potential importance of links among climate, shallow groundwater levels, and groundwater evapotranspiration using a coupled aquifer-land surface-atmosphere model.

## INTERACTIONS BETWEEN GROUNDWATER AND SURFACE WATER

Interactions between groundwater and surface water are an integral part of the hydrological cycle. The interactions need to be viewed at a range of scales and as having large variability in space and time. Many of the early studies of these interactions focused on streams. Once thought to be of little consequence and thus ignored, interactions of groundwater with lakes, wetlands, estuaries, and oceans are now recognized as important processes. The interactions of surface-water bodies with groundwater systems are governed by the positions of the water bodies relative to the groundwater-flow system, the characteristics of surface-water beds and underlying geologic materials, and the climatic setting (Winter *et al.*, 1998).

### Streams

Streams interact with groundwater in various ways. Depending upon the relation between groundwater head at the stream–channel interface and the stream stage, a stream may gain water from inflow of groundwater through the streambed (gaining stream; stream stage < groundwater head), lose water to groundwater by outflow through the streambed (losing stream; stream stage > groundwater head), gain water from groundwater on one side of the stream and lose water to groundwater on the other side (flow-through), or possibly have zero exchange (parallel-flow) when the channel stage and groundwater head are equal throughout a cross section (Woessner, 2000). The position of a channel within a floodplain can control the type of exchange. For example, flow-through reaches are most often found when a channel cuts perpendicular to the fluvial plain groundwater-flow field (Wroblicky *et al.*, 1998; Woessner, 2000). Larkin and Sharp (1992) classify stream-aquifer systems based on the predominant regional groundwater-flow component as (i) underflow component–dominated (groundwater flow is largely parallel to and in the same direction as the stream), (ii) baseflow component–dominated (groundwater flow is largely perpendicular to and from the stream), and (iii) mixed. They conclude that the dominant groundwater component (baseflow or underflow) can be inferred from geomorphic data such as channel slope, river sinuosity, degree of river incision through the alluvium, width-to-depth ratio of the bankfull river channel, and the character of the fluvial depositional system.

In some environments, streamflow gain or loss can persist; that is, a stream might always gain water from groundwater, or it might always lose water to groundwater. In many environments, the flow direction can vary along a stream; some reaches receive groundwater and other reaches lose groundwater. Flow directions between groundwater and surface water also can change on daily or shorter timeframes as a result of individual storms causing focused recharge near the streambank, flood peaks moving down the channel, or transpiration of groundwater by streamside vegetation.

Losing streams can be connected to the groundwater system by a continuous saturated zone or can be disconnected from the groundwater system by an unsaturated zone. An important feature of streams that are disconnected from groundwater is that pumping shallow groundwater near the stream does not affect the flow of the stream near the pumped wells.

Even in settings where streams are primarily losing water to groundwater, certain reaches may receive groundwater inflow during some seasons. The amount of water that groundwater contributes to streams can be estimated by analyzing streamflow hydrographs to determine the groundwater (baseflow) component (Rutledge, 1993; Halford and Mayer, 2000). The proportion of stream water derived from groundwater inflow varies considerably among different physiographic and climatic settings, as illustrated in Figure 5 for streams in 10 different regions in the United States.

Water exchange across the interface between groundwater and surface water has been explored in some detail in the past decade (Jones and Mulholland, 2000). Many studies have demonstrated that local geomorphic features such as streambed topography, streambed roughness, meandering, and heterogeneities in sediment hydraulic conductivities can give rise to localized flow systems within streambeds and banks (Harvey and Bencala, 1993; Winter *et al.*, 1998). The near-stream subsurface environment in which there is active

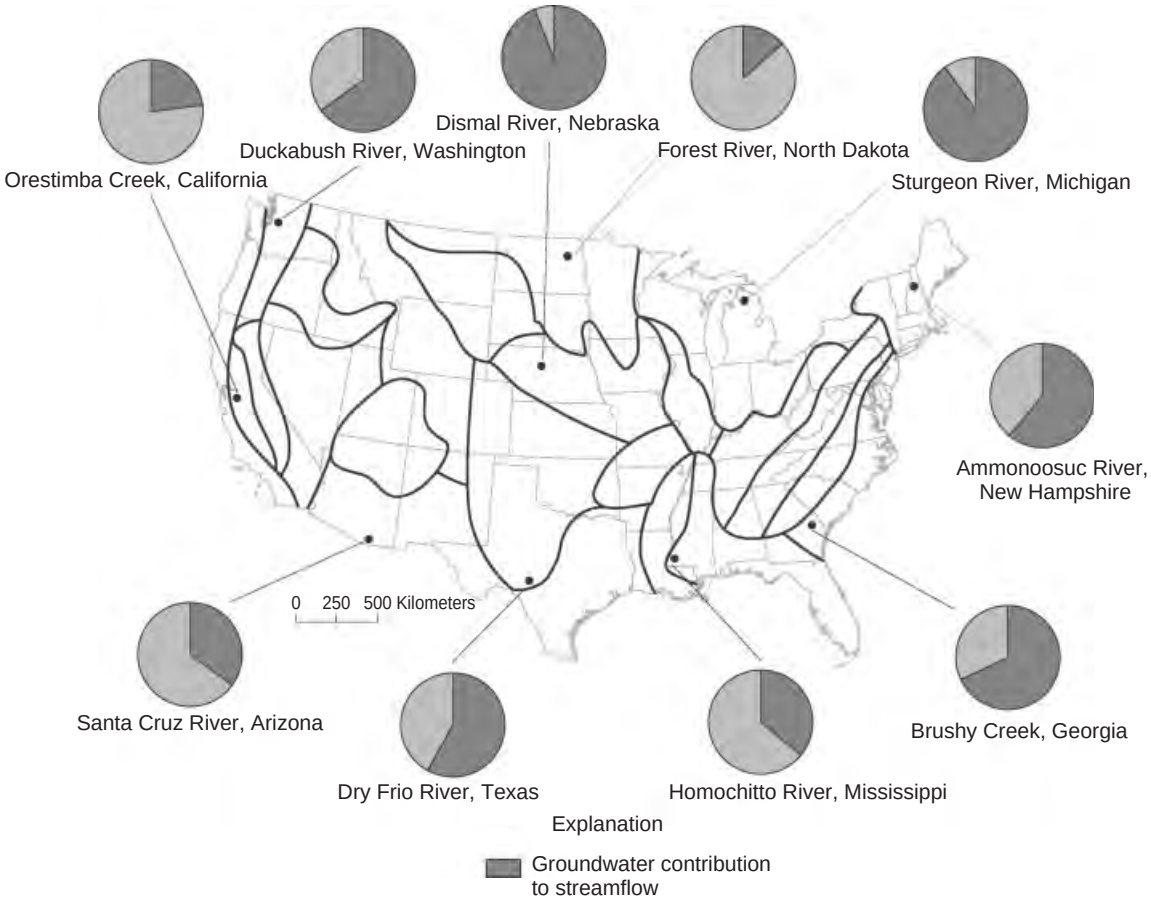


**Figure 5** Twenty-four regions were delineated in the conterminous United States where the interactions of groundwater and surface water are considered to have similar characteristics. The estimated groundwater contribution to streamflow is shown for specific streams in 10 of these regions. (Modified from Winter *et al.,*1998)

exchange between surface water and groundwater commonly is referred to as the hyporheic zone. The hyporheic zone can be viewed as a zone in which finer-scale interactions between the stream-channel water and groundwater occur within the context of larger-scale patterns of loss and gain of channel water in drainage basins. Understanding these very localized subsystems within the hydrological cycle is increasingly viewed as important to understanding the chemical composition of surface and subsurface water and stream and riparian ecology (Brunke and Gonser, 1997; Boulton, 2000). For example, an important feature of water exchange between streams and hyporheic zones is that surface water is kept in close contact with chemically reactive mineral coatings and microbial colonies in the subsurface, having resultant effects on biogeochemical processes such as nitrification (Dahm *et al.,* 1998; Hinkle *et al.,* 2001). Localized exchange of surface water and groundwater also can result in thermal effects that influence the distribution of biota and biogeochemical processes. For example, thermal effects of groundwater discharge have been directly related to fish habitat, both in terms of spawning areas and refuge for adults when ice forms in colder environments (Power *et al.,* 1999)

## Lakes and Wetlands

The relation of lakes and wetlands to groundwater can be viewed as occurring in three different ways: (i) lakes and wetlands may recharge groundwater without receiving groundwater input, (ii) they may both recharge groundwater and receive discharge water from groundwater, or (iii) they may receive groundwater discharge, but not discharge water to groundwater (Sloan, 1972; Winter, 1976; Boyle, 1994). These relations are influenced by the positions of the lakes or wetlands within regional flow systems (Smith and Townley, 2002). Groundwater-flow patterns in the vicinity of these water bodies can vary in response to shifts in subsurface-water divides that occur during wetter or drier conditions (Holzbecher, 2001).

The amount of groundwater moving into or through lakes and wetlands varies considerably. Negligible to relatively small amounts of groundwater move into or

through lakes and wetlands in poor hydraulically conductive materials, such as some crystalline rocks (Schindler *et al.*, 1976), clayey till (Shaw and Prepas, 1990), lacustrine clays (Zektzer and Kudelin, 1966), or dense organic peats. More considerable amounts of groundwater move into or through lakes and wetlands in more conductive materials such as sandy tills, sands, or sands, and gravels (Shaw and Prepas, 1990), as well as in karst (Lee, 2002).

Groundwater movement into and out of lakes and wetlands can be diffused, focused, or some combination thereof. Focused groundwater discharge into lakes and wetlands is in the form of springs. The direction of flow at the periphery of lakes and wetlands can be affected by transpiration. Flow reversals in response to transpiration have been documented in a variety of geologic materials, such as glacial till (Meyboom, 1967), organic peat (Fraser *et al.*, 2001), and sandy till (Doss, 1993).

### Springs

Springs typically are present where the water table intersects the land surface. They may form at topographic depressions; along lithologic contacts where permeable rocks overlie rocks of much lower permeability; along joints, fractures, or faults; or in karst areas (Fetter, 1988). Springs serve as important sources of water to streams and other surface-water features, as well as being important cultural and esthetic features in themselves. The important role of springs in human settlement is evident from the many localities named after the local springs. The constant source of water typical at many springs leads to abundant growth of plants and, in many cases, to unique habitats. Groundwater development can lead to reductions in spring flow, alterations of springs from perennial to ephemeral, or elimination of springs altogether. Springs typically represent points on the landscape where groundwater-flow paths from different sources converge. Groundwater development may affect the amount of flow from these different sources to varying extents, thus affecting the chemical composition of the spring water. Where springs discharge at land surface, they may offer a relatively easy place to measure natural groundwater discharge. However, because springs integrate the signal of hydrologic and geologic processes, sometimes over large spatial areas and long periods of time, they typically provide only an indirect source of information about groundwater systems (Manga, 2001).

### Bays, Estuaries, and Oceans

The interface between saltwater and freshwater in groundwater systems is important in determining the amount of freshwater available in the aquifer system. This boundary generally occurs as fresh groundwater discharges into salty bays, estuaries, and oceans; however, the boundary also occurs inland where deeper brines bound an active freshwater flow system. In relatively homogeneous porous media, the denser saltwater tends to remain separated from the overlying freshwater. A transition zone, however, known as the *zone of diffusion* or the *zone of dispersion*, forms between the two fluids. In coastal areas where the porous media is heterogeneous in nature, a system of layered mixing zones can form. Substantial groundwater withdrawals can cause surrounding saltwater to move into areas of groundwater use in coastal (Bear *et al.*, 1999; Barlow, 2003) and inland (Lahm and Bair, 2000) areas, decrease the volume of freshwater available for use, and reduce the amount of freshwater discharged into coastal ecosystems.

Most research to date has been on the landward movement of saltwater in response to pumping in coastal areas. More recently, the seaward flow of fresh groundwater to coastal ecosystems and the role of groundwater in delivering nutrients and other dissolved constituents to these systems have received increased attention (Burnett *et al.*, 2002). As with all the components of the hydrological cycle, this circulation of water from one part of the cycle to another part is important to many issues.

The time required for the saltwater–freshwater interface and freshwater discharge to respond to human and natural changes can range from almost instantaneous to thousands of years. In some coastal areas, such as New Jersey, United States, and Suriname, South America (Kooi and Groen, 2001), relatively fresh groundwater found far off the coast is hypothesized to be water that was recharged during the Pleistocene when sea levels were much lower and has not yet been discharged to the surface.

## CONCLUDING REMARKS

During the past 50 years, groundwater depletion has spread from isolated pockets to large areas in many countries throughout the world. Although sometimes viewed as a separate resource, groundwater is in fact an integral part of the hydrological cycle. Groundwater systems are dynamic and continually changing in response to human and climatic stresses. Linkages with surface water and the land surface are numerous and extend over many different timescales.

Future success in understanding the dynamic nature of groundwater systems will rely on continued and expanded data collection at various scales. Despite their importance, groundwater data have received little attention in concerns expressed about the continuity of global water data, primarily because such concerns have focused more on visible atmospheric and surface-water monitoring networks.

Water-level measurements from wells remain the principal source of information on the effects of hydrologic stresses on groundwater systems. Advances in instrumentation now enable the collection of real-time water-level data allowing scientists to observe diurnal and seasonal

trends from well networks across large areas (Cunningham, 2001). To understand the true nature of change in a groundwater system and to differentiate between natural and human-induced changes, records of water-level measurements over substantial periods are required (Taylor and Alley, 2001).

Surface-water depletion is viewed increasingly as the limiting factor to the long-term utilization of groundwater resources, yet the distinctly different temporal and spatial scales at which groundwater and surface-water systems operate present major challenges to their integrated analysis. The locations, quantity, and timing of reductions in surface-water flow resulting from groundwater development are fundamental questions at the scales of years to decades, whereas ecological issues require attention to seasonal and even diurnal changes in groundwater recharge and discharge. There is a continuing need to develop modeling tools that better represent groundwater as an integral part of the hydrological cycle; for example, coupled watershed and groundwater models are needed to better simulate the hydrological continuum from the atmosphere, to the land surface, to the unsaturated and saturated subsurface zones, and then back to surface waters or land surface.

Finally, groundwater systems have value not only as perennial sources of water supply, but also as reservoirs for cyclical injection and withdrawal to modulate the variability inherent in surface-water supplies. Management approaches increasingly involve the use of artificial recharge of excess surface water or recycled water by direct well injection, surface spreading, or induced recharge from streams. Many scientific challenges remain, however, to understand more fully the long-term responses of aquifer systems to alternate management approaches (Galloway *et al.*, 2003).

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