Water Budgets: Foundations for Effective Water-Resources and Environmental Management

Circular 1308

Photograph credits:

1. U.S. Department of Energy Atmospheric Radiation Measurement Program
2. U.S. Geological Survey
3. U.S. Geological Survey
4. U.S. Fish and Wildlife Service
5. U.S. Geological Survey

Water Budgets: Foundations for Effective Water-Resources and Environmental Management

By Richard W. Healy, Thomas C. Winter, James W. LaBaugh, and O. Lehn Franke

Circular 1308

U.S. Department of the Interior
DIRK KEMPTHORNE, Secretary

U.S. Geological Survey
Mark D. Myers, Director

U.S. Geological Survey, Reston, Virginia: 2007

Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

Foreword

Water availability is an important concern in the 21st century. Ensuring sustainable water supplies requires an understanding of the hydrologic cycle—how water moves through Earth's atmosphere, land surface, and subsurface. Water budgets are tools that water users and managers use to quantify the hydrologic cycle. A water budget is an accounting of the rates of water movement and the change in water storage in all or parts of the atmosphere, land surface, and subsurface. Although simple in concept, water budgets may be difficult to accurately determine. It is important for the public and decisionmakers to have an appreciation of the uncertainties that exist in water budgets and the relative importance of those uncertainties in evaluating how much water may be available for human and environmental needs.

As part of its mission, the U.S. Geological Survey (USGS) provides information that describes the Earth, its resources, and the processes that govern the availability and quality of those resources. This Circular provides an overview of the hydrologic cycle and a discussion of methods for determining water budgets and assessing the uncertainties in those determinations. Examples illustrate the importance of water budgets to humans and the environment and demonstrate how water budgets can be incorporated into management practices. Through this Circular, the USGS seeks to inform the public and decisionmakers about a scientific basis for water-resources and environmental management and to broaden awareness and understanding of water budgets and the hydrologic cycle so as to promote wise use and management of a most precious resource—water.

Robert M. Hirsch
Associate Director for Water

How long can the water needs of a growing urban area be sustained by an aquifer that contains a finite amount of water?

What are the ecological effects of withdrawing water from an aquifer that naturally discharges to a wetland? Will the withdrawal result in reduced discharge to and subsequent drying of the wetland? How will plants and animals be affected?

How will droughts affect agricultural and domestic water supplies? Will increased diversions reduce storage in surface-water reservoirs to the point where recreational uses are limited?

Preface

Water is the essence of life. Its availability determines where and how animals and plants exist on Earth. Humans need water for consumption, for producing food, and for manufacturing; we also are attracted to water for its esthetic value and for the recreational opportunities it offers. At the same time, all other life forms on Earth require water for their sustenance. Native plants in grasslands and forests; wheat and corn crops in agricultural fields; insects, amphibians, and birds in wetlands; fish in streams and lakes; wild mammals and reptiles; and domesticated pets and livestock—all depend on water.

Competition for water among humans and between humans and other life forms is the unavoidable outcome of burgeoning populations and a limited resource. Resolution of competing needs requires decisions based on science as well as societal values. Informed decisions are developed with an understanding of the hydrologic cycle—the process by which water moves from the atmosphere to land surface as precipitation, infiltrating the subsurface or flowing along land surface to the oceans, and eventually returning to the atmosphere by evaporation. All water on Earth resides in one of the three compartments of the hydrologic cycle: the atmosphere, the land surface, and the subsurface. A water budget is an accounting of water stored within and water exchanged among some subset of the compartments, such as a watershed, a lake, or an aquifer.

Throughout history, humans have managed water for their own needs. Ancient Mayan and Egyptian cultures prospered on crops produced with intricate irrigation systems. Remains of aqueducts built almost two thousand years ago by the Roman Empire can still be found throughout Europe. Early

explorers of the American West, such as John Wesley Powell, realized that civilization could flourish in this arid region only if water could be stored and distributed as needed. Today, population centers and agriculture thrive in the West, mainly because of the dams and reservoirs constructed on rivers such as the Colorado and Columbia. Design and operation of large reservoir projects rely on detailed water-budget analyses, examination of precipitation and evaporation rates, discharge rates of streams, rates of exchange between surface water and ground water, and factors such as climate, geology, vegetation, and soils that affect those rates. The story of water development in the Western United States is a story that has been repeated in various forms all over the Earth.

Reservoirs and ground-water wells are key features of the Nation's water supply infrastructure. They both provide great benefits in terms of the reliable delivery of water to users. However, it is well-recognized that they can also have adverse impacts on aquatic ecosystems. The needs and values of society determine whether or not the benefits of these systems outweigh their negative consequences and determine if changes in the design or operation of these systems should be made. Water needs of ecosystems have become an integral part of water management. Operators of reservoirs now take into account the health of downstream riparian ecosystems. Managers of aquifers are likely to consider the effects of ground-water withdrawals on the interactions between ground water and surface water and the organisms that depend on that interaction. These are but a few of the myriad issues that arise in balancing the water needs of humans and the environment. Water budgets form the foundations of informed management strategies for resolving these issues.

Can crops be matched to climate so as to minimize irrigation requirements?

Can streamflow in arid regions be increased by the removal of non-native phreatophytes that line channels, thus reducing evapotranspiration? How much water will be used by replacement vegetation?

Will dewatering of a surface mine have an effect on surface-water expressions many miles away?

Contents

Boxes

Rain is grace;
rain is the sky
condescending to the earth;
without rain, there would be no life.

John Updike (1989)

Conversion Factors and Datums

Multiply	By	To Obtain
Length		
centimeter (cm)	0.3937	inch
millimeter (mm)	0.03937	inch
meter (m)	3.281	foot (ft)
kilometer (km)	0.6214	mile (mi)
Area		
square meter (m^2)	0.0002471	acre
hectare (ha)	2.471	acre
square kilometer (km^2)	247.1	acre
square centimeter (cm^2)	0.001076	square foot (ft^2)
square meter (m^2)	10.76	square foot (ft^2)
square centimeter (cm^2)	0.1550	square inch (ft^2)
hectare (ha)	0.003861	square mile (mi^2)
square kilometer (km^2)	0.3861	square mile (mi^2)
Volume		
cubic meter (m^3)	264.2	gallon (gal)
cubic centimeter (cm^3)	0.06102	cubic inch (in^3)
cubic meter (m^3)	0.0008107	acre-foot (acre-ft)
Flow rate		
cubic meter per year (m^3/yr)	0.000811	acre-foot per year (acre-ft/yr)
cubic meter per second (m^3/s)	35.31	cubic foot per second (ft^3/s)
cubic meter per second (m^3/s)	22.83	million gallons per day (Mgal/d)

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:
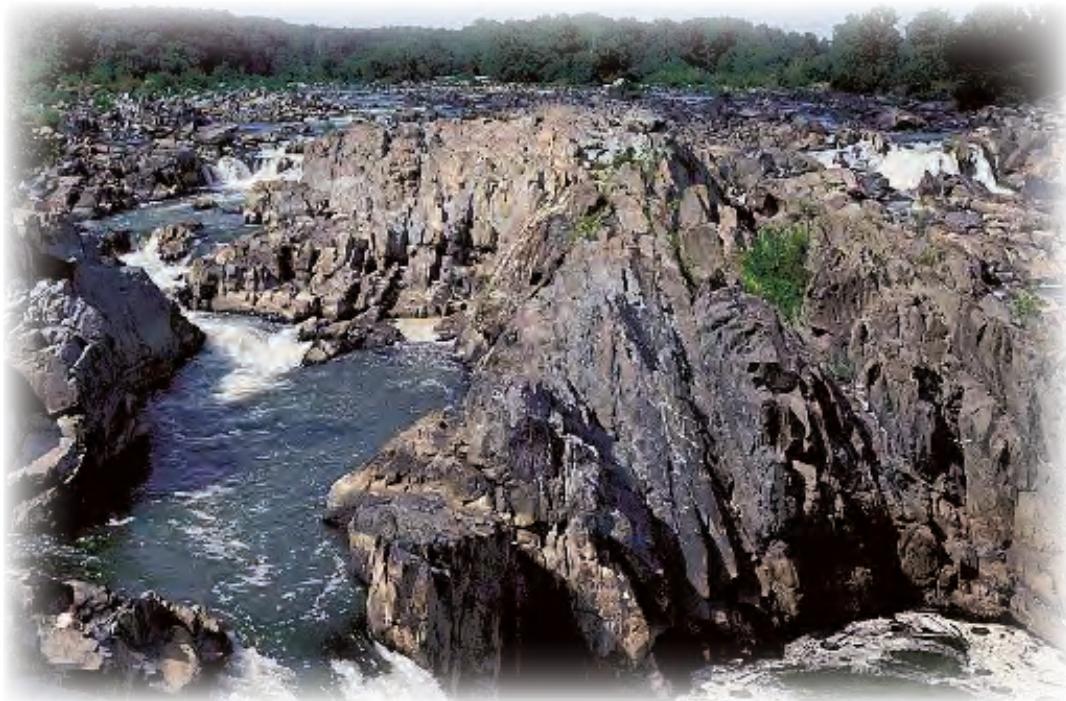
$$°F=(1.8×°C)+32$$

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

$$°C=(°F–32)/1.8$$

Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88).

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

I came where the river
Ran over stones
My ears knew
An early joy.
And all the waters
Of all the streams
Sang in my veins
That summer day

Theodore Roethke from "The Lost Son"
(1948)

Water Budgets: Foundations for Effective Water-Resources and Environmental Management

By Richard W. Healy, Thomas C. Winter, James W. LaBaugh, and O. Lehn Franke

Introduction

Water budgets provide a means for evaluating availability and sustainability of a water supply. A water budget simply states that the rate of change in water stored in an area, such as a watershed, is balanced by the rate at which water flows into and out of the area. An understanding of water budgets and underlying hydrologic processes provides a foundation for effective water-resource and environmental planning and management. Observed changes in water budgets of an area over time can be used to assess the effects of climate variability and human activities on water resources. Comparison of water budgets from different areas allows the effects of factors such as geology, soils, vegetation, and land use on the hydrologic cycle to be quantified.

Human activities affect the natural hydrologic cycle in many ways. Modifications of the land to accommodate agriculture, such as installation of drainage and irrigation systems, alter infiltration, runoff, evaporation, and plant transpiration rates. Buildings, roads, and parking lots in urban areas tend to increase runoff and decrease infiltration. Dams reduce flooding in many areas. Water budgets provide a basis for assessing how a natural or human-induced change in one part of the hydrologic cycle may affect other aspects of the cycle.

"Only from space can you see that our planet should not be called Earth, but rather Water, with speck-like islands of dryness on which people, animals, and birds surprisingly find a place to live."

Oleg Makarov (1988)

Natural Resources Conservation Service

This report provides an overview and qualitative description of water budgets as foundations for effective water-resources and environmental management of fresh-water hydrologic systems. Perhaps of most interest to the hydrologic community, the concepts presented are also relevant to the fields of agriculture, atmospheric studies, meteorology, climatology, ecology, limnology, mining, water supply, flood control, reservoir management, wetland studies, pollution control, and other areas of science, society, and industry. The first part of the report describes water storage and movement in the atmosphere, on land surface, and in the subsurface, as well as water exchange among these compartments. Our ability to measure these phenomena and inherent uncertainties in measurement techniques also are discussed. The latter part of the report presents a number of case studies that illustrate how water-budget studies are conducted, documents how human activities affect water budgets, and describes how water budgets are used to address water and environmental issues.
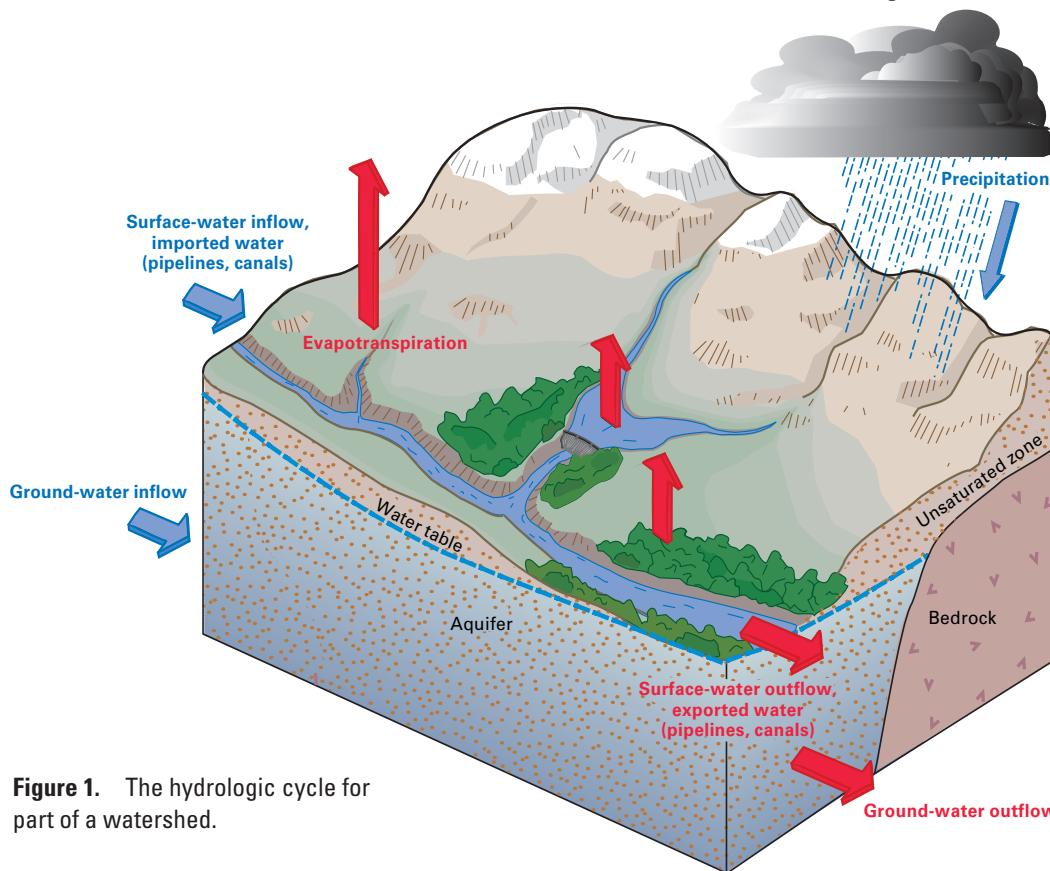
Natural Resources Conservation Service

Hydrologic Cycle

Earth's water exists on land surface in oceans, ice fields, lakes, rivers, streams, and wetlands; it also exists in the sub-surface as soil water and ground water and in the atmosphere (fig. 1). More than 97 percent of the Earth's water is in oceans (table 1). Of the inland water that resides on and beneath land surface, 77 percent is contained in icecaps and glaciers and for practical purposes is inaccessible. The remaining inland water is stored primarily in the subsurface as ground water. Water is constantly moving within the hydrologic cycle, and that movement takes place over many pathways (fig. 1). Water moves quickly through some pathways; for example, rain falling from the atmosphere to a field of corn in summer may return to the atmosphere in a matter of hours or days by evaporation. Traveltimes over other pathways are measured in years, decades, centuries, or more—ice fields in Greenland contain water that fell from the atmosphere thousands of years ago.

Figure 1. The hydrologic cycle for part of a watershed.

"The central concept in the science of hydrology is the so-called hydrologic cycle—a convenient term to denote the circulation of the water from the sea, through the atmosphere, to the land; and thence, with numerous delays, back to the sea by overland and subterranean routes, and in part, by way of the atmosphere; also, the many short circuits of the water that is returned to the atmosphere without reaching the sea***. The science of hydrology is especially concerned with the second phase of this cycle—that is, with the water in its course from the time it is precipitated upon the land until it is discharged into the sea or returned to the atmosphere. It involves the measurement of the quantities and rates of movement of water at all times and at every stage of its course***."

O.E. Meinzer (1942, p. 1)

Table 1. Estimated global water supply (from Nace, 1967).

[km³, cubic kilometers]

Water storage	Volume, in thousands of km³	Percentage of total water
Ocean water	1,320,000	97.1
Atmosphere	13	0.001
Water in land areas	37,800	2.8
Freshwater lakes	125	0.009
Saline lakes and inland seas	104	0.008
Rivers	1.25	0.0001
Icecaps and glaciers	29,200	2.14
Soil root zone	67	0.005
Ground water (to depth of 4,000 meters)	8,350	0.61

D.H. Campbell

"It is the sea that whitens the roof.
The sea drifts through the winter air.
It is the sea that the north wind makes.
The sea is in the falling snow."

Wallace Stevens from
"The Man With the Blue Guitar" (1937)

The atmosphere receives water through evaporation and loses it as precipitation, mostly in the form of rain or snow. The average residence time for water in the atmosphere is about 10 days. A drop of rain can have a multitude of fates, depending on where and when it falls. Some rainfall never reaches land surface; instead, it evaporates as it falls (a phenomenon known as virga) and returns to the atmospheric reservoir. A falling raindrop could land on a leaf of a tree, from where it might fall to the ground, evaporate, or perhaps be imbibed by the plant. Another drop might land directly on the ground. That water could puddle in a depression, travel over the surface to a lower elevation (runoff), or enter the subsurface (infiltrate). Water in a puddle will likely evaporate or infiltrate. Water that runs off may infiltrate at a more favorable location or travel to a stream and ultimately be transported to an ocean; at any point on this journey, that water can evaporate. The average residence time for water in free-flowing rivers ranges between 16 and 26 days (Vorosmarty and Sahagian, 2000). Streams that run through reservoirs can have substantially longer residence times. Not all surface water flows to oceans. Some lakes and wetlands have no surface drainage. They lose water to evaporation and to ground water. Humans withdraw water from streams and reservoirs, thus interrupting its migration to the ocean.

Water moves much more slowly in the subsurface than in the atmosphere or on land surface. Water that infiltrates the subsurface can remain in the unsaturated zone where it will most likely be returned to the atmosphere by evaporation or plant transpiration; it can discharge to the surface in a channel or depression, thus becoming surface flow; or it can traverse the unsaturated zone to recharge an underlying aquifer. Most water that infiltrates the subsurface is returned to the atmosphere by evaporation from bare soil or by plant transpiration (table 2). That returned water typically resides in the subsurface for less than a year. Discharge to land surface of unsaturated-zone water, sometimes referred to as interflow,

may occur days to months after that water has infiltrated, depending on the distance between the points of infiltration and discharge. Infiltrated water that travels downward past the depth of the root zone may eventually reach the saturated zone, thus becoming aquifer recharge. Traveltimes of water through the entire thickness of the unsaturated zone span a very large range: from hours, for thin unsaturated zones in humid regions (Freeze and Banner, 1970), to millennia, for thick unsaturated zones in arid regions (Phillips, 1994). Water that reaches the saturated zone may reside there for days to thousands of years (Alley and others, 2005). Under natural conditions, ground water discharges to surface-water bodies such as streams, wetlands, lakes, or oceans, or it is extracted by plants and returned to the atmosphere by transpiration. Humans also extract ground water for agricultural, domestic, and industrial uses; such water is ultimately reapplied to land surface, returned to the subsurface, or discharged to surface-water bodies.

Table 2. Water budget for global land mass (from Lvovitch, 1973). Evapotranspiration is the sum of evaporation and plant transpiration.

Water-budget component	Annual rate, in milli-meters	Percentage of annual precipitation
Precipitation	834	100
Evapotranspiration	540	65
Total discharge to oceans	294	35
Discharge to oceans from surface runoff	204	24
Discharge to oceans from base flow	90	11
Infiltration of precipitation	630	76

Spring at head of Paris Canyon, Bear Lake County, Idaho 1912.

Rainbow Falls on the Missouri River near Great Falls, Montana, 1904.

Precipitation in the form of snow can follow several courses. In many environments, snow accumulated on land surface melts in a few days or less. In other areas, a seasonal snowpack exists throughout winter and melts in the spring. Still other areas, such as Greenland and Antarctica, have snow and ice fields that are thousands of years old. In any of these cases, the melting water flows to a surface-water body, infiltrates into the subsurface, or is evaporated back into the atmosphere.

It is evident from the preceding discussion that water moves within the hydrologic cycle along many complex pathways over a wide variety of time scales. The challenge for humans is to monitor the hydrologic cycle for some geographic feature of interest, such as a watershed, a reservoir, or an aquifer. Such a feature will be referred to as an accounting unit. A water budget states that the difference between the rates of water flowing into and out of an accounting unit is balanced by a change in water storage:

Flow In – Flow Out = Change In Storage.

Simple, yet universal, the water-budget equation is applicable over all space and time scales, from studies of rapid infiltration in a laboratory soil column to investigations of continental-scale droughts over periods of decades or centuries. A 1-m^2 soil column in the middle of an agricultural field, the entire field itself, or the watershed in which the field lies—these are all examples of water-budget accounting units.

*"And you, vast sea,
sleepless mother,
Who alone are peace and
freedom to the river
and the stream,
Only another winding will
this stream make,
only another murmur
in this glade,
And then shall I come to
you, a boundless drop to a
boundless ocean."*

Kahlil Gibran (1923)

A   The Water-Budget Equation

The water-budget equation is simple, universal, and adaptable because it relies on few assumptions on mechanisms of water movement and storage. A basic water budget for a small watershed can be expressed as:

$$P + Q_{in} = ET + \Delta S + Q_{out} \qquad (A1)$$

where

P is precipitation,
Q_{in} is water flow into the watershed,
ET is evapotranspiration (the sum of evaporation from soils, surface-water bodies, and plants),
ΔS is change in water storage,

and

Q_{out} is water flow out of the watershed.

The elements in equation A1 and in all other water-budget equations are referred to as components in this report. Water-budget equations can be written in terms of volumes (for a fixed time interval), fluxes (volume per time, such as cubic meters per day or acre-feet per year), or flux densities (volume per unit area of land surface per time, such as millimeters per day). Typically, water budgets are tabulated in spreadsheets or tables such as that shown in table A–1, which contains monthly and yearly data for Seabrook, New Jersey, from Thornthwaite and Mather (1955). With the approach used by those authors, it is assumed that Q_{in} is zero and Q_{out} is equal to runoff.

Equation A1 can be refined and customized depending on the goals and scales of a particular study. Precipitation can be written as the sum of rain, snow, hail, rime, hoarfrost, fog drip, and irrigation. Water flow into or out of the site could be surface or subsurface flow resulting from both natural and human-related causes. Evapotranspiration could be differentiated into evaporation and plant transpiration. Further refinement could be based on the source of the water that is evapotranspired. Evaporation can occur from open water, bare soil, or snowpack (sublimation); plants can extract ground water or water from the unsaturated zone. Such refinements must be balanced with available measurement techniques, which often are not designed, or lack sufficient resolution, to distinguish among sub-components. Most methods for measuring evapotranspiration, for example, quantify the flux of water from the land/vegetation surface to the atmosphere and do not distinguish between different water sources. Fashioning a viable water-budget approach for estimating evapotranspiration or other water-budget components requires analysis of available measurement techniques.

Water storage occurs within all three compartments of the hydrologic cycle. The amount of water stored in the atmosphere is small compared to that on land surface and in the subsurface. Surface water is stored in rivers, ponds, wetlands, reservoirs, icepacks, and snowpacks. Subsurface storage can be categorized into various subaccounting units, such as the root zone, the unsaturated zone as a whole, the saturated zone, or different geologic units. An expanded form, but certainly not an exhaustive refinement, of the water budget appropriate for many hydrologic studies can be written as (Scanlon and others, 2002):

$$P + Q^{sw}_{in} + Q^{gw}_{in} = ET^{sw} + ET^{gw} + ET^{uz} + \Delta S^{sw} + \Delta S^{snow} + \Delta S^{uz} + \Delta S^{gw} + Q^{gw}_{out} + RO + Q^{bf} \qquad (A2)$$
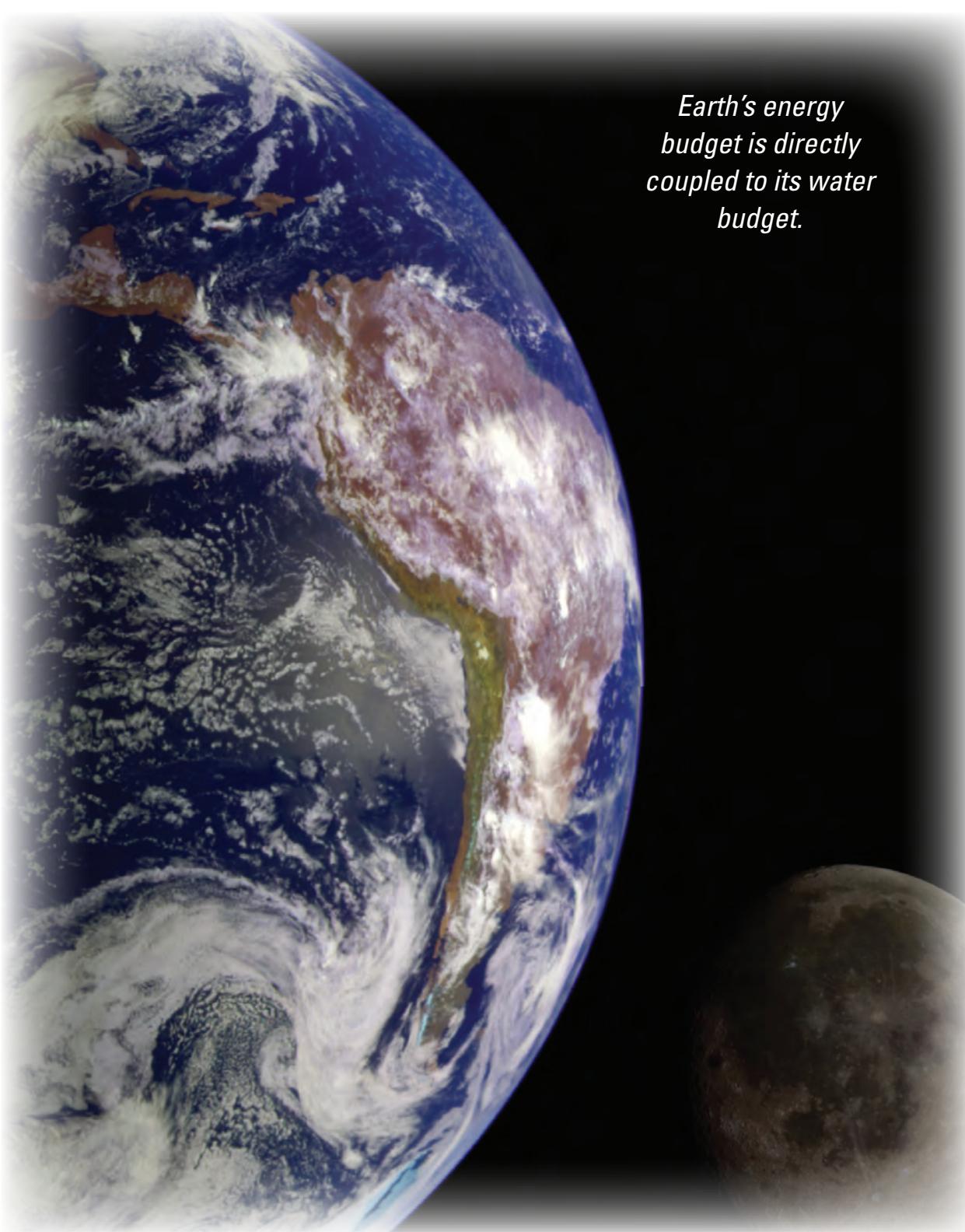
where the superscripts refer to surface water (*sw*), ground water (*gw*), unsaturated zone (*uz*); *RO* is surface runoff; Q^{gw}_{out} refers to both ground-water flow out of the site and any withdrawal by pumping; and Q^{bf} is base flow (ground-water discharge to streams). It is unlikely that all elements in equation A2 will be of importance at any one site; some will be of negligible magnitude and can be ignored. Indeed, when selecting an accounting unit for developing a water budget, judicious selection of boundaries can greatly facilitate the account-ing process. Consider, for example, a small watershed and associated shallow ground-water system. Watershed boundaries are well defined: there is no surface flow in, and surface flow out occurs only in a stream channel, where discharge can be readily measured. If watershed boundaries correspond to ground-water divides, there is also no subsurface inflow. Suppose all ground water that is not lost to *ET* eventually discharges to the stream; an appropriate water budget for the watershed could be stated as:

$$P = ET + \Delta S + RO + Q^{bf} \qquad (A3)$$

If the annual change in storage is small, evapotranspiration can be estimated as the difference between precipitation and stream-flow out of the watershed.

Table A–1.   *Monthly and yearly water budget, in millimeters, for Seabrook, New Jersey (Thornthwaite and Mather, 1955).*

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year total
Precipitation	87	93	102	88	92	91	112	113	82	85	70	93	1,108
Storage change	0	0	0	0	0	−38	−35	−17	−10	32	51	17	0
Evapotranspiration	1	2	16	46	92	129	147	130	92	53	19	3	730
Runoff	61	76	81	61	31	15	8	4	2	1	1	37	378
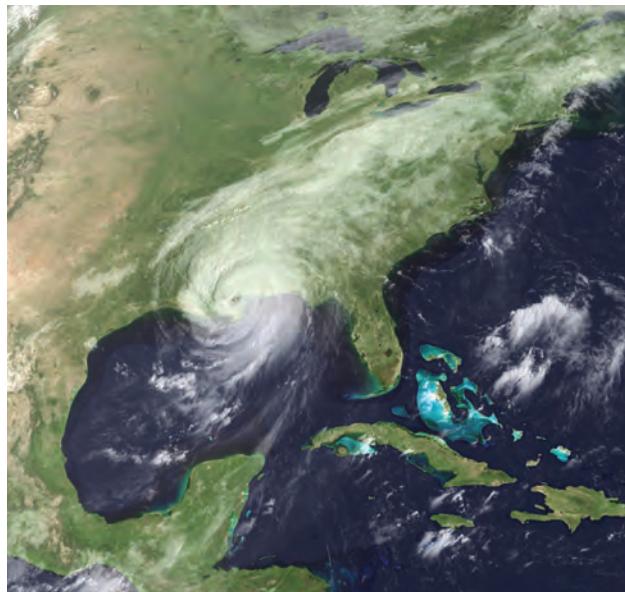
Earth's energy budget is directly coupled to its water budget.

Storage and Movement of Water Within the Principal Compartments of the Hydrologic Cycle

The atmosphere, the land surface, and the subsurface are the three compartments that hold the Earth's water. Each compartment acts as a storage reservoir within which water moves from its point of entry to the compartment to its point of outflow. Water also moves between compartments. A water-budget accounting unit may consist of a single part of one compartment, such as a lake, or an accounting unit may comprise parts of all three compartments, such as a watershed. This section discusses storage and movement of water within individual compartments. The following section discusses exchange of water between compartments.

Water in the Atmosphere

At any one time, the atmosphere holds only a small fraction of the Earth's water (table 1, fig. 2), the equivalent of a layer about 25 mm thick over all of the Earth's surface. Yet this compartment is a vital part of the hydrologic cycle in terms of water storage and transport. Water flows to the atmosphere in a gaseous form as it evaporates from water, plant, and soil surfaces. This water will eventually condense, and possibly freeze, and be returned to the Earth's surface as precipitation. Between the times of entry and departure from

National Aeronautics and Space Administration
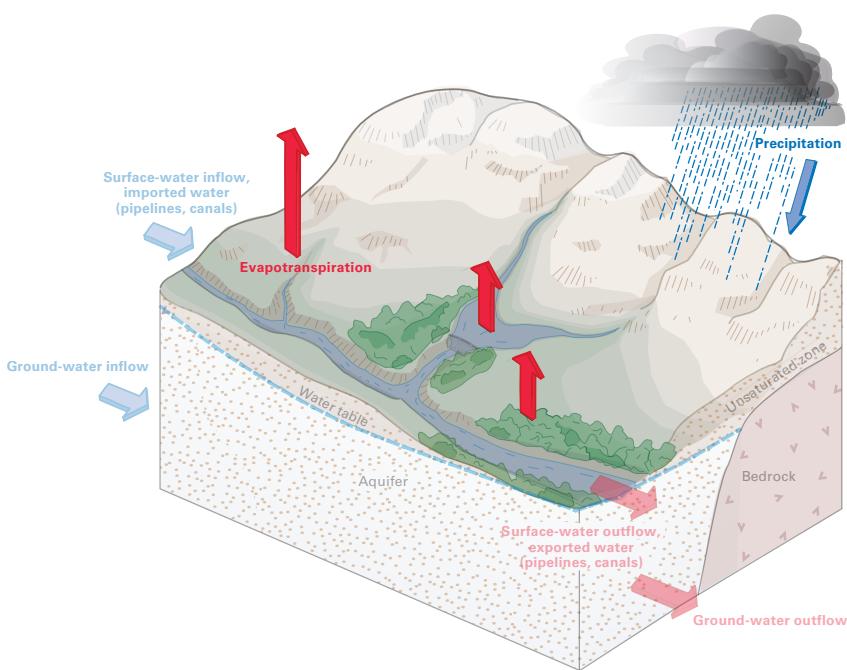
Hurricane Katrina.

the atmosphere, a water molecule can be transported rapidly over long distances. The atmosphere is part of an amazing water-distribution system, carrying water from where it is plentiful (primarily oceans) and depositing it in regions where it is less plentiful (land surfaces).

Water in the atmosphere is also important in Earth's energy balance and climate. Evaporation and subsequent condensation of water require transfers of energy. As water moves from the liquid to gaseous state, it absorbs energy; as it condenses, that energy is released. Thus, the transport of water in the atmosphere is accompanied by a large transport of energy, effectively distributing energy across the Earth. Atmospheric water also affects radiation transfer at land surface. The formation of clouds limits the amount of solar radiation that reaches land surface. Long-wave radiation emitted by the Earth is absorbed and reflected back by gases, including water vapor, in the atmosphere (the greenhouse effect). Global climate and water storage in the atmosphere are linked. As the Earth's temperature changes, so does the ability of the atmosphere to store water. In cyclic fashion, changes in the amount of water stored in the atmosphere can alter the Earth's energy balance and thus affect surface temperatures.

Movement of water within the atmosphere occurs over a range of space and time scales. Movement occurs both by convection (water-vapor transport by moving air masses) and molecular diffusion (the natural

Figure 2.   The atmosphere within the hydrologic cycle.

tendency of water vapor to move from areas of high concentration to areas of low concentration). The lower part of the atmosphere, called the atmospheric boundary layer, is the part of the atmosphere that is most influenced by the Earth's surface. The layer varies in height between about 500 and 2,000 m and typically holds about one-half of all atmospheric water. It is characterized by turbulent mixing generated as warm, moist air pockets move up from the heated surface and by frictional drag as the atmosphere moves over the Earth's surface. Horizontal transport rates of water vapor within the atmospheric boundary layer can be as high as 50 to 100 km/day (Oke, 1978).
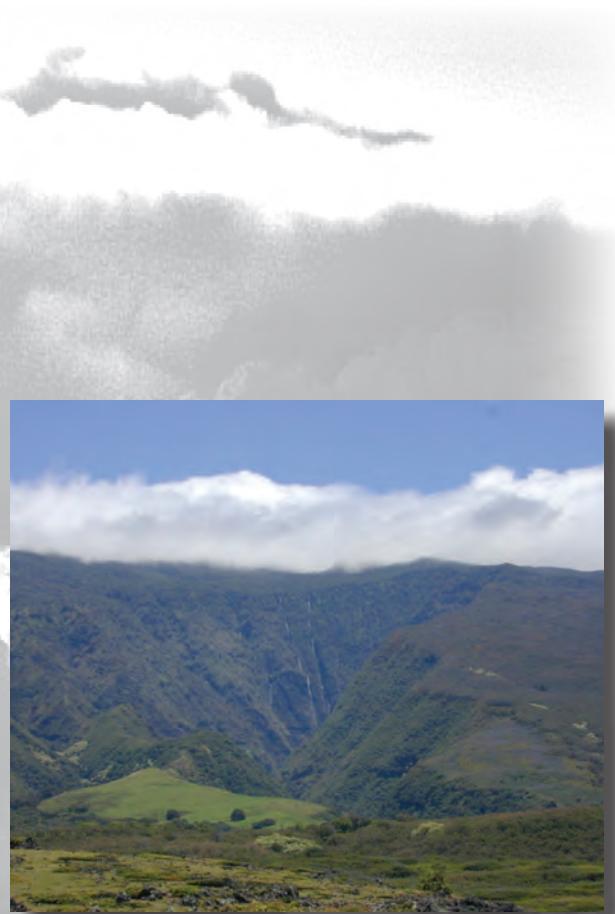
Atmospheric transport of water is driven by gradients in pressure, temperature, and humidity. Predictions of moisture storage and movement are integral parts of weather forecasts. These forecasts are based on large-scale computer models that rely on data collected at National Weather Service surface monitoring sites across the United States. These surface sites provide point measurements of temperature, pressure, and humidity. Radar and satellite imagery provide additional data that are integrated over large areas.

Cumulus clouds.

U.S. Department of Energy Atmospheric Radiation Measurement Program

Clouds in Hawaii.

Anvil cloud.

U.S. Department of Energy Atmospheric Radiation Measurement Program

B   Water Budgets are Intimately Linked to Energy and Chemical Budgets
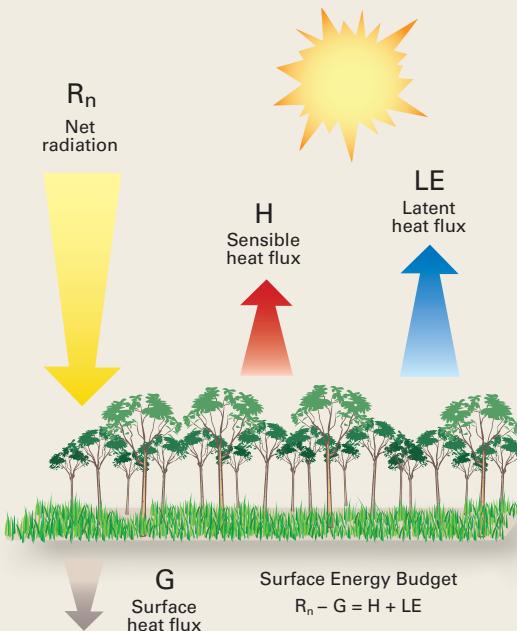
Energy Budget

The global water budget is intrinsically linked to the global energy budget. When water changes among its different phases (solid, liquid, and gas) energy is absorbed or released, thus affecting the energy budget. A simple energy budget for the Earth is (Sellers, 1965):

$$Rn = G + LE + H \qquad \text{(B1)}$$

where Rn is net radiation (the sum of incoming solar and long-wave radiation minus reflected solar and emitted longwave radiation); G is surface-heat flux (that is, the energy used to warm soil, or water in the case of a surface-water body); LE is latent heat flux (that is, the energy used to evaporate water); and H is sensible heat flux, or the energy used to warm air. The equation states that available energy at the Earth's surface goes to heating the surface, warming the air, and evaporating water (fig. *B–1*) . Latent heat flux is the product of latent heat of vaporization (λ) and evapotranspiration rate (ET); that is, $LE = \lambda ET$. Evapotranspiration provides a direct link between the energy-budget and the water-budget equations because it appears in both equations. These equations form the basis of general circulation computer models that are used to predict climate trends. Estimation of ET rates can be addressed from both energy-budget and water-budget perspectives.

The movement of heat in ground and surface waters may be materially affected by the movement of water. An important component of energy transport is convection, or the movement of heat by the movement of water. The transport of energy by surface water is important in studies of powerplant or dam discharges in rivers where the health of natural fish populations is affected by heat loads or changing temperatures. Ground-water flow has been shown to be an important controlling factor on the occurrence and severity of volcanic eruptions (Matsin, 1991). The interdependence of water and energy movement has proved useful for estimating rates of exchange between ground and surface waters (Lapham, 1989; Stonestrom and Constantz, 2003).

Figure B–1.   *Schematic of the energy budget at the Earth's surface.*

Geothermal plant in California.

Geothermal Education Office, Tiburon, California

Water vapor rises from hot springs in Yellowstone National Park.

Chemical Budget

Chemical fluxes are important to our environment. For example, fluxes and storage of carbon in the ocean, on land, within inland waters, and in the atmosphere have vital implications for ecosystems and climate. Water movement within and among the atmosphere, surface, and subsurface is an important mechanism for transport of chemicals through the environment. The water budget provides a foundation for understanding chemical fluxes and balances. As water contacts rocks, sediment, and organic materials, its chemistry is altered by reactions such as dissolution, precipitation, ion exchange, and oxidation/reduction. Ground- and surface-water flows sustain many wetlands, lakes, and ponds. In addition to supplying water, these inflows also provide nutrients and chemicals that support biogeochemical processes within these bodies.

Chemicals are transported to the atmosphere naturally (by diffusion and wind advection, and through plants, fires, and volcanic activity) and as a result of human activities (combustion of fossil fuels, application of agricultural chemicals, and production of chemical compounds). Some chemicals become dissolved in atmospheric water and fall back to Earth in precipitation. Sulfate-bearing precipitation has been implicated as a major cause for acidification of some lakes in the Adirondack Mountains of New York (Driscoll and others, 2003).

Surface waters are reservoirs and conveyance mechanisms for chemicals and sediment. Sediment and contaminants can be washed off of streets and fields during rainfalls and be carried through storm drains to streams. It is estimated that, in one year, the Mississippi River discharged 900,000 tons of nitrate and 35,000 tons of orthophosphate to the Gulf of Mexico (Antweiler and others, 1995). Severe rainfalls can lead to flooding, which can greatly enhance the transport capabilities of surface water. Floods are capable of transporting not only sediment and chemicals but also pathogens, animals, cars, and even houses.

National Aeronautics and Space Administration

The Mississippi discharging water and sediment to the Gulf of Mexico.

Floodwaters can transport debris.

Water moves more slowly through the subsurface than it does through surface-water bodies or the atmosphere. Hence, removal of subsurface contaminant plumes may take much longer than cleanup of surface plumes. Long residence times in the subsurface allow more time for reactions to occur and, in some instances, may promote natural remediation of contaminants by indigenous microbes (Lahvis and others, 1999).

National Aeronautics and Space Administration

*Ice fields in Antarctica, such as the Ross Ice Shelf, store
about 70 percent of Earth's freshwater.*

Water on Land Surface

Freshwater is present on the Earth's land surface in solid and liquid forms. Solid forms include snow and ice; liquid water is stored in lakes, surface-water reservoirs, some wetlands, and streams.

Snow and Ice

The largest amount of freshwater on Earth (29.2 million km³) is stored in glaciers and polar ice (Nace, 1967). Most of this ice is present in Antarctica and Greenland and is largely inaccessible to humans. Solid water present as glaciers and snow in more temperate regions may be available for humans (fig. 3). Here, snow and ice serve as seasonal storage receptacles that contribute to water supplies upon melting. Melt from the annual snowpack, especially that captured in reservoirs, is the primary source of water for humans and aquatic ecosystems in many parts of the world. Glaciers represent a more permanent form of water storage. Residence time of water stored in glaciers can be decades to centuries. Meltwater from glaciers can sustain streamflows throughout the year.

*Snowpits are dug to determine water content and chemistry
of snowpacks.*
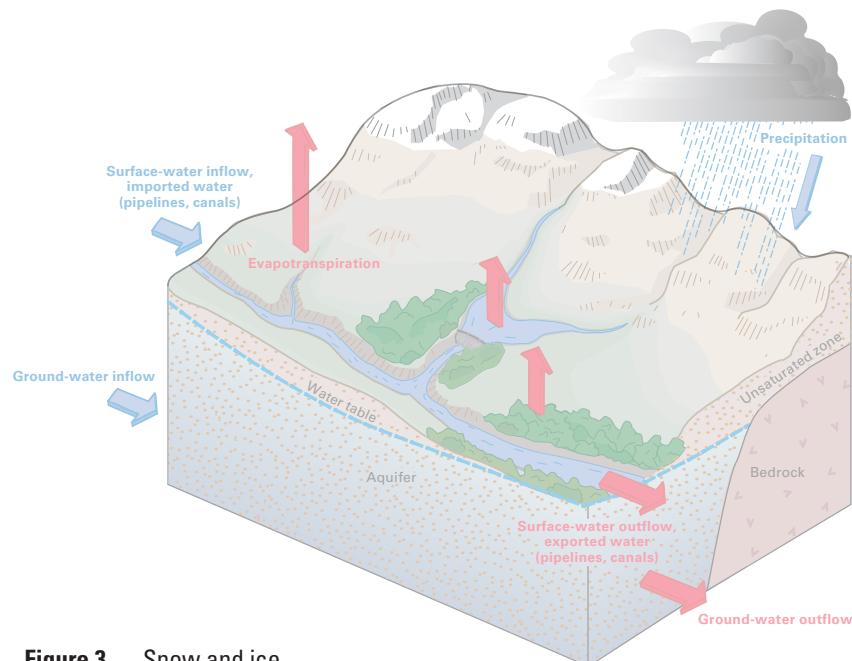
Toboggan Glacier, Alaska, photographed by S. Paige on June 29, 1909 (left), and by Bruce F. Molnia on September 4, 2000 (right).

Atmospheric water, primarily in the form of snow, is the source of water to glaciers and snowfields. Water moves from these bodies to the atmosphere (as ablation), to the subsurface (as infiltration), and to streams. Measurement of water storage in seasonal snowpacks generally is done by conducting snow surveys, where snow depth and the water content of snow are determined in designated areas or along snow courses that transect an area. Measurement of changes in water stored in glaciers has historically been difficult because high mountain terrain is often inaccessible. Storage changes were determined by repeated detailed surveys of the ice surface topography. In recent years, remote sensing from aircraft or satellite, used in conjunction with high-resolution digital-elevation models, has greatly enhanced the accuracy of these measurements.

Accurate determinations of water budgets of glaciers are rare. Only a few studies of glaciers have resulted in detailed, long-term monitoring of their water budgets (Mayo and others, 2004). However, in a general way, comparative photographs (a form of remote sensing) of glaciers show that many glaciers have been shrinking over the last few decades.

Figure 3.   Snow and ice within the hydrologic cycle.

C   Chemical, Isotopic, and Energy Tracers Provide Insight into Hydrologic Processes

Direct physical measurements of water-budget components may at times be inconvenient, problematic, or impractical. In such cases, indirect methods may provide estimates of water-budget components or act to reduce the uncertainty associated with those estimates. Chemical, isotopic, and energy (heat) tracers are commonly used to provide insight into processes such as ground-water recharge, ground-water discharge to lakes and wetlands, and base flow. A tracer is simply a chemical or isotope (or property, in the case of heat) that is transported by water. Analysis of spatial or temporal patterns of tracer concentrations can be used to identify trends in water movement and therefore can provide insight for shaping conceptual models of water budgets. The ideal hydrologic tracer is one that moves with water, is conservative (that is, not altered by reactions or other processes in water, porous media, or atmosphere), and is easily and accurately detected. Tracers can be categorized as environmental, historical, and applied. Environmental tracers are those that occur naturally in the environment. Isotopes of oxygen and hydrogen have been used for decades to distinguish sources of water and to examine water balances (Gat and Gonfiantini, 1981). These isotopes are well suited as tracers because they are part of the water molecule itself. Carbon isotopes, chloride, sulfate, and nitrate are other useful environmental tracers. Historical tracers are those that were released to the environment continuously or or at specific times during the past. Radionuclides (including tritium, ^{3}H,

Tracers are used to determine the age of subsurface water (that is, the time since that water last had contact with the atmosphere), velocities and traveltimes for ground and surface waters, and travel paths of water in the subsurface.

Applying dye tracer to land surface at research site in Minnesota.
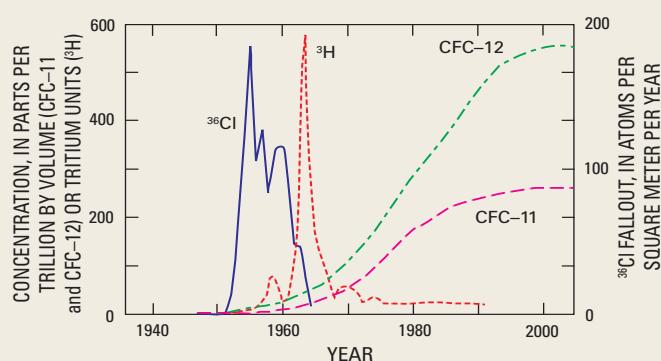
Dye tracer is visible in trench excavated beneath application area.
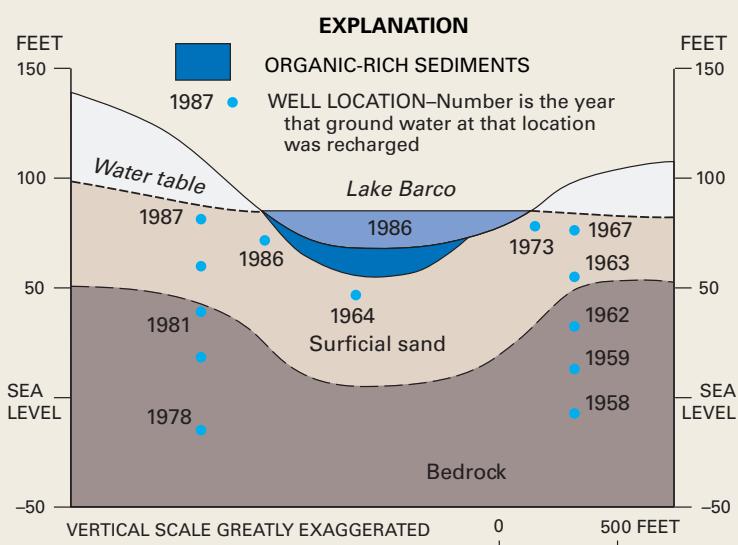
Tracers are used in streams to measure traveltimes.

and chlorine-36, ^{36}Cl) released to the atmosphere from testing of nuclear bombs in the 1950s and 1960s fall into this class (fig. *C–1*). Chlorofluorocarbons (CFCs) and sulfur hexafluoride were released to the atmosphere by industrial processes over the last 50 years and are common hydrologic tracers (*http://water.usgs.gov/lab/*). For example, Katz and others (1995) used concentrations of CFCs to estimate the ages of ground water near Lake Barco in Florida (fig. *C–2*). Applied tracers include those introduced intentionally (for example, chloride, bromide, and dyes) and those inadvertently introduced to the environment, such as through a chemical spill. Applied tracers commonly are used to determine velocities of streamflow and ground-water flow, to identify subsurface flow paths, and to quantify exchange rates between surface and ground waters. Properties and uses of common hydrologic tracers are given in table *C–1*.

Figure C–1. *Atmospheric concentrations for historical tracers, including ^{3}H, ^{36}Cl, CFC–11, and CFC–12 (after Scanlon and others, 2002).*

Figure C–2. *Lake Barco, in northern Florida, is a flow-through lake with respect to ground water. The dates when water in different parts of the ground-water system was recharged indicate how long it takes water to move from the lake or the water table to a given depth (after Katz and others, 1995).*

Table C–1. *Examples of tracers used in water-budget studies.*

Use	Naturally occurring in the environment	Historical — Added to the environment from human activity in the past	Applied — Added to the environment in the present	Example study
Ground-water age — Time since recharge water became isolated from the atmosphere	^{35}S, ^{14}C, ^{3}H/^{3}He, ^{39}Ar, ^{36}Cl, ^{32}Si	^{3}H, ^{36}Cl, ^{85}K, chlorofluorocarbons, herbicides, caffeine, pharmaceuticals		Plummer and others (2001)
Temperature of recharge	N^{2}/Ar solubility			Plummer (1993)
Tracing ground-water flow paths	^{18}O, ^{2}H, ^{13}C, ^{87}Sr	Chlorofluorocarbons, herbicides, caffeine, pharmaceuticals	Cl, Br, dyes	Renken and others (2005)
Exchange of surface water and ground water	^{18}O, ^{2}H, ^{3}H, ^{14}C, ^{222}Rn		Cl, Br, dyes	Katz and others (1997)
Surface-water discharge and traveltime			Cl, Br, dyes	Kimball and others (2004)

Lakes
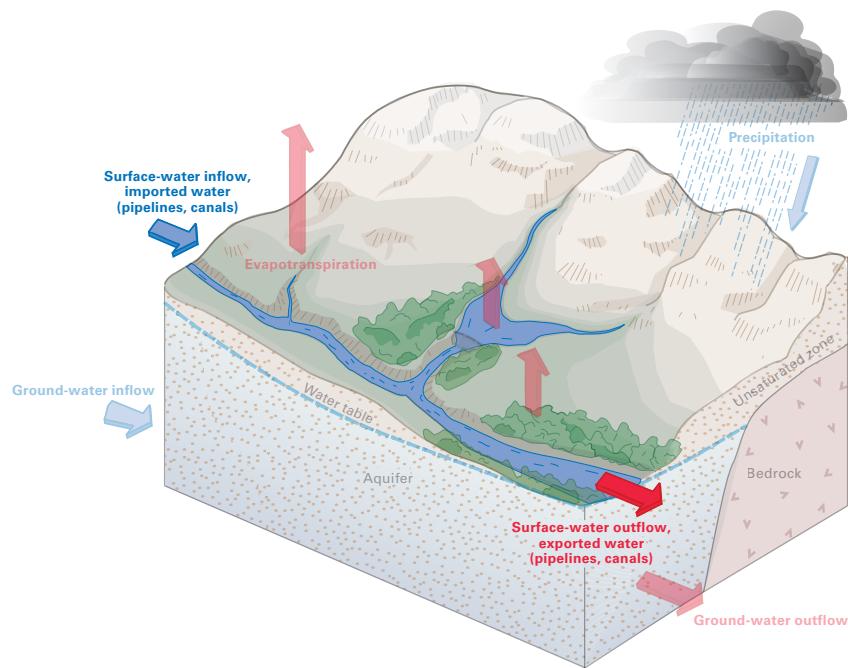
Lakes are the fourth largest reserve of water in the global water budget. The volume of water in natural lakes is estimated to be about 229,000 km^3 (table 1; fig. 4). Of this total volume, 125,000 km^3 are in freshwater lakes and 104,000 km^3 are in saline lakes; the Caspian Sea alone contains about 95 percent of the total volume of water in saline lakes. For this report, surface-water reservoirs are considered to be lakes. Lvovitch (1973) estimated the total volume of water in reservoirs to be about 5,000 km^3. The largest reservoir in the United States, Lake Mead, contains about 38 km^3 of water at full pool elevation; Lake Powell contains about 33 km^3 at full pool elevation.

Lakes interact with the atmosphere, the subsurface, and other surface-water features. They gain water from precipitation, streamflow, and ground water and lose water by evaporation, surface outflow, and seepage to ground water. However, all these interactions do not occur for every lake. Some topographically high lakes have no stream or ground-water inflows, gaining water only from precipitation. At the other topographic extreme, some lakes, called terminal lakes, receive water from precipitation, streams, and ground-water inflow and lose water only by evaporation.

The volume of water in a lake may be determined by preparing a bathymetric map of the lake bottom and by calculating the volume of water present at a given lake stage (lake level). Lake stage is measured by reading a staff gage, or it can be continuously monitored by a recording gage. A stage-volume relation is then established that can be used to determine the volume of water at any given stage. This approach can produce accurate results if the bathymetry is well defined.

Residence times of water in lakes span a wide range. Residence time is calculated by dividing the volume of a lake by the rate of outflow. For very large lakes, like Lake Superior, residence time is nearly 200 years. Lake Powell, much smaller but still a large surface-water reservoir, has a residence time of about 2.3 years. Lakes with no stream outlet, like many in glacial terrain, can have residence times of several years to a decade, and small lakes with outlet streams commonly have residence times of days to weeks (Winter, 2003)

Figure 4.   Lakes, wetlands, and streams within the hydrologic cycle.

Crater Lake, Oregon.

Lake country in northern Wisconsin.

The Great Lakes.

National Aeronautics and Space Administration

"Wetlands are lands where saturation with water is the dominant factor determining the nature of soil development and the types of plant and animal communities living in the soil and on the surface. The single feature that most wetlands share is soil or substrate that is at least periodically saturated with or covered by water."

Cowardin and others (1979, p. 3)

Lower Klamath National Wildlife Refuge, California.

U.S. Fish and Wildlife Service

Wetlands

Wetlands in depressions generally contain standing water and in many respects are much like lakes. Many types of wetlands do not contain standing water, however, or contain it for only brief periods each year. Such wetlands consist mainly of saturated soils. Most wetlands receive surface-water inflow at some time of the year, some are fed by both surface and ground water, and others are supported solely by ground-water flow. Like lakes, some wetlands located high in the landscape gain water only from precipitation; others, low in the landscape, like terminal lakes, lose water only to the atmosphere. A major difference between wetlands and lakes is that wetlands lose water to the atmosphere largely by transpiration from plants, whereas lakes lose water to the atmosphere mostly by evaporation.

Determining the volume of water in a wetland and the change in that volume over time is more difficult than it is for lakes because, other than the open-water portion, water is present in wetland soils. Measurement of water storage in soils is addressed in the section "Unsaturated Zone."

Depressional wetlands, North Dakota.

Red Rock Lakes National Wildlife Refuge, Montana.

"A river seems a magic thing. A magic, moving, living part of the very earth itself—for it is from the soil, both from its depth and from its surface, that a river has its beginning."

Laura Gilpin (1949)

Streams

The volume of water in the Earth's streams at any given time (about 1,250 km^3 according to Nace, 1967) represents only a small part of the total volume stored on land surface. Streams, then, are generally not important in terms of global water storage. Streams function mainly to transport water, conveying it from higher to lower altitudes on the land surface and, in most cases, ultimately to the oceans. Streams also facilitate water exchange between the surface and the subsurface, and to a lesser extent between the surface and the atmosphere.

Sources of water in streams can be surface-water bodies, surface runoff of precipitation (as well as direct precipita-tion on a stream), interflow (shallow subsurface flow usually associated with hillslopes), and base flow (ground-water discharge). Along their course, streams can lose water to other surface-water bodies, to the subsurface, and to the atmosphere (by evaporation). Streams range in size from small rivulets in headwater areas that flow only after precipitation events to large rivers, such as the Mississippi and the Amazon. Magnitudes of velocities in streams are variable; 30 cm/s may be typical, whereas 300 cm/s is quite high. Because they are confined to channels on the Earth's surface, streams are visible and relatively accessible for measurement of discharge and are therefore the part of the hydrologic cycle that can be measured most accurately.
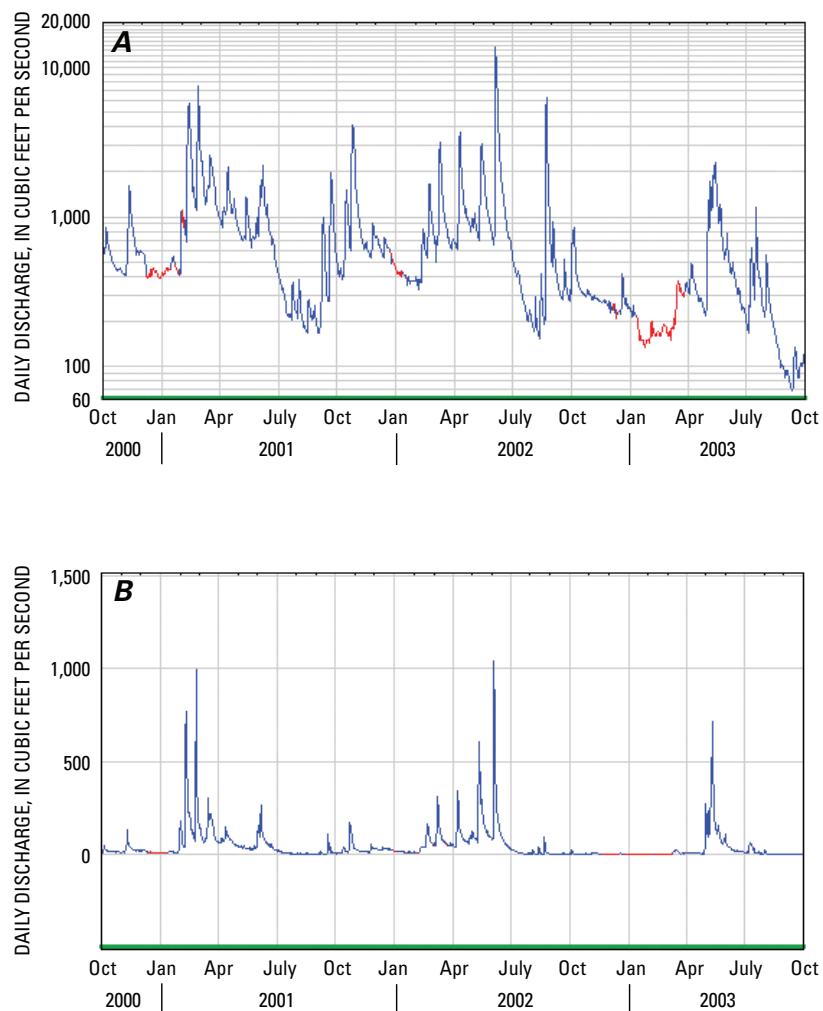
Large streams in the United States tend to show seasonal trends (fig. 5A); highest discharges generally occur in spring, a time when snow melts, soils thaw, and soil moisture contents are high. Small streams are usually more dynamic than large streams and they show rapid rises and falls in response to storms (fig. 5B). The source of water in a stream also influences discharge patterns. Streams dominated by snowmelt or base flow follow a more predictable pattern than those dominated by surface runoff.

For major streams, the U.S. Geological Survey maintains a network of thousands of stream gages across the United States (*http://water.usgs.gov*). Stream level (stage) is

American Geological Institute, Michael Collier

Stream network in arid region, Organ Pipe National Monument, Arizona.

Figure 5.   Streamflow hydrographs for two gaging sites: (*A*) Kishwaukee River near Perryville, Illinois (U.S. Geological Survey station number 05440000; drainage area 1,099 square miles) and (*B*) South Branch Kishwaukee River at DeKalb, Illinois (U.S. Geological Survey station number 05439000; drainage area 77.7 square miles). Red indicates estimated values.

monitored continuously at these sites, and a stage/discharge relation is developed using periodic discharge measurements. Discharge is the product of stream velocity and cross-sectional area integrated over that area. Velocity has historically been measured manually at many locations along a cross section by using a current meter. Recently, acoustic velocity meters have reduced the need for manual measurements (Yorke and Oberg, 2002). By establishing good stage/discharge relations, stream

discharge can be determined from measurements of stage. Typical errors in stream discharge measurements are 10 percent (Rantz and others, 1982).

For small streams, more accurate measurements of discharge can be obtained by installing a flume or a weir and a stage recorder in the channel. Flumes and weirs are carefully calibrated in hydraulic laboratories, so measurements of discharge commonly have errors of about 5 percent.

Discharge is determined with measurements of stream depth and velocity.

USGS gaging station.
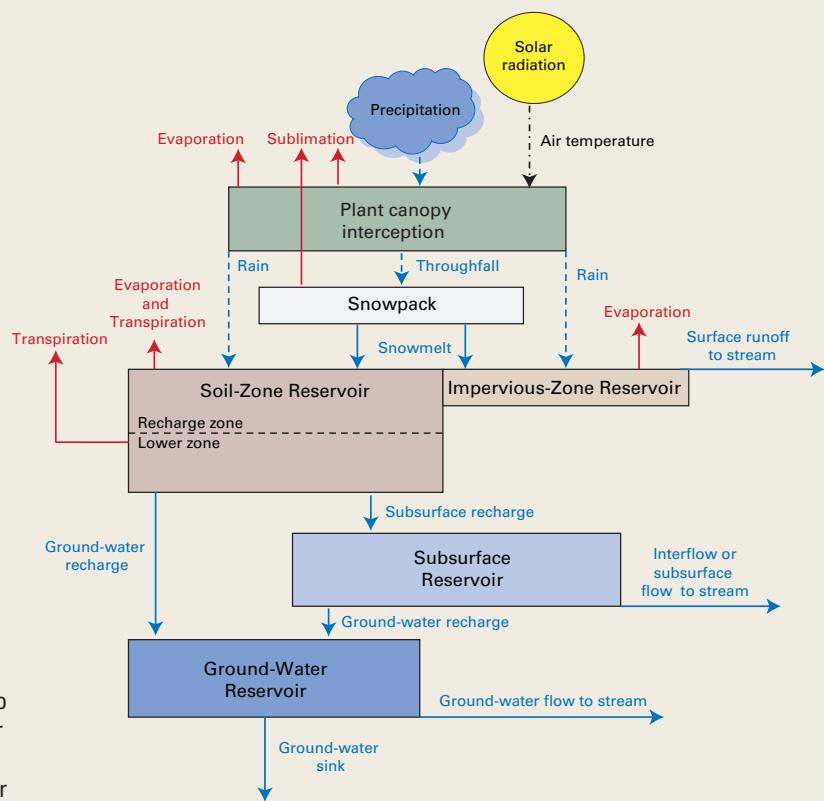
Flumes can be used to measure discharge in small streams.
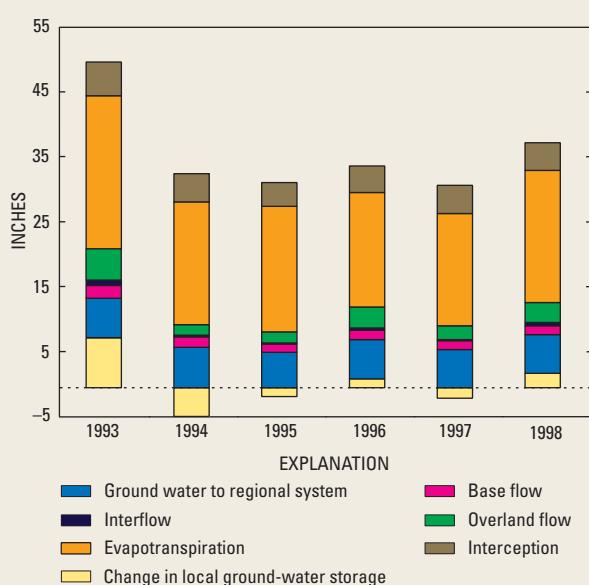
D   Models—Important Tools in Water-Budget Studies

Hydrologic computer-simulation models contribute substantially to our understanding of the hydrology of watersheds, rivers, and aquifers. They are integral tools for managing water resources in many areas. Using calculations that are too cumbersome to be performed by hand, these models allow detailed investigation of complex hydrologic processes and provide predictions of responses within a specific water-budget accounting unit to external or internal stresses. Most hydrologic computer-simulation models are derived from some variant of equation A2 and thus are truly water-budget models. As water-budget equations vary greatly in complexity, so do the models that are based on them. A simple model may provide a quick view of the water budget for an accounting unit but is unlikely to provide insight into the processes that drive water movement within that unit. A more complex model may provide that insight but at substantially greater expense.
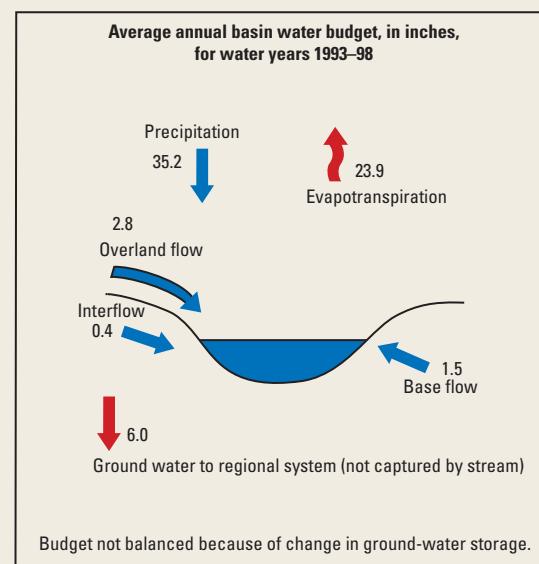
Watershed models are perhaps the most complete form of a water-budget model. They predict stream discharge within a basin in response to precipitation and snowmelt, usually accounting for processes such as evapotranspiration, ground-water/surface-water exchange, and surface-water routing (fig. *D–1*). Watershed models are widely used for watershed management and planning. For example, they can be used to predict the effects of land-use changes (such as urban development) on streamflow (fig. *D–2*).
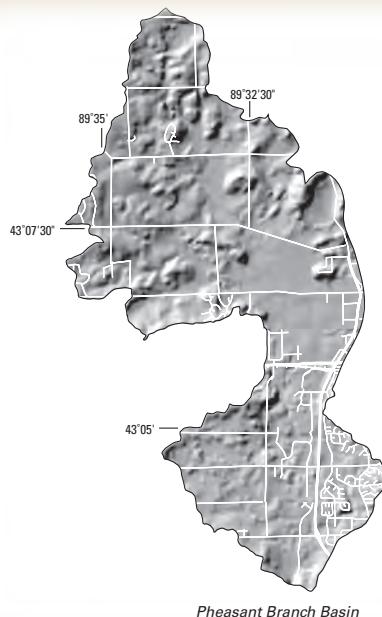
Figure D–1.   Shematic diagram showing various reservoirs and processes that are considered in a watershed model (R.S. Regan, written commun., 2007).

Figure D–2.   Steuer and Hunt (2001) used a watershed model to simulate water fluxes in the Pheasant Branch Creek watershed near Middleton, Wisconsin, for the period 1993 to 1998. The model was subsequently used to predict the effects of urban development in the watershed.

Pheasant Branch Basin

Figure D–2. *Steuer and Hunt (2001) used a watershed model to simulate water fluxes in the Pheasant Branch Creek watershed near Middleton, Wisconsin, for the period 1993 to 1998. The model was subsequently used to predict the effects of urban development in the watershed.—Continued*

Ground-water-flow models predict how water levels in an aquifer will be affected by changes in withdrawals or in recharge rates (fig. *D–3*). They are used in studies of ground-water supply and ground-water contaminant transport. Most of these models simulate flow only in the saturated zone (that is, the region beneath the water table). Other more complex models simulate water movement within both the unsaturated and saturated zones.
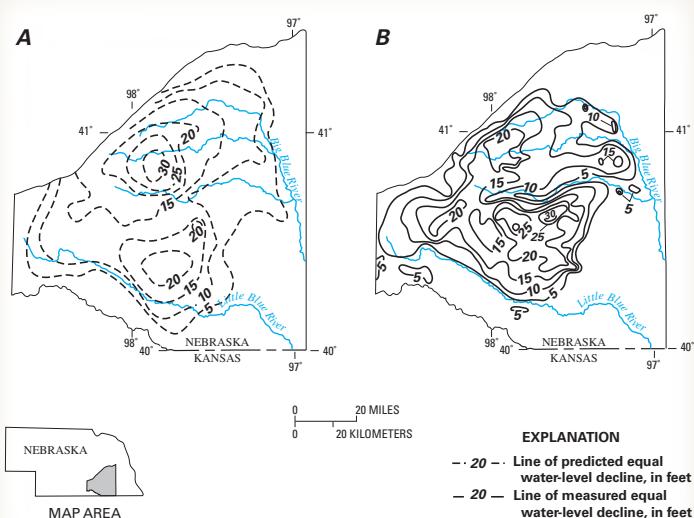
Streamflow routing models predict stream discharge and velocity. Managers use these models to estimate where, when, and at what stage flood waves will crest, allowing them to adjust release rates from reservoirs to mitigate adverse effects of flooding.

General circulation models forecast weather and climate trends at the continental scale over periods of days to centuries.

Soil–vegetation–atmospheric transport models are used to study the movement of water from the atmosphere to the soil through plants and back into the atmosphere.

Coupled models combine water-budget models with mass or energy transport models and are useful for simulating contaminant transport in surface or ground water.

Statistical techniques (such as regression, nonparametric statistics, and geostatistics), while not water-budget models, are important in many water-budget studies. They can be used for quantifying uncertainty in simulation results, determining which types of data can improve simulation results, and interpolating and integrating point measurements (from a rain gage, for example) over entire watersheds or basins.

Figure D–3. *Contours of model predicted (A) and measured (B) ground-water levels for the Blue River Basin, Nebraska (Alley and Emery, 1986).*

Water Budgets: Foundations for Effective Water-Resources and Environmental Management

Heterogeneity of subsurface sediments complicates study of water movement in the subsurface.

Water in the Subsurface

The Earth's subsurface consists of solid rock, mineral grains, organic matter, and varying amounts of water and other liquids and gases that occupy open spaces or voids. The subsurface serves as the major reservoir of extractable fresh-water, accounting for more than 95 percent of worldwide storage. On the annual global scale, change in storage of water in the subsurface is negligible. At smaller scales, changes in subsurface storage can be substantial and significant. Ground-water levels in the San Joaquin Valley of California declined as much as 100 m between 1920 and 1970 as a result of pumping for irrigation. In addition to a reduction in the amount of water stored in the subsurface, the declining water levels resulted in land-surface subsidence of more than 9 m in some areas (Galloway and others, 1999).

A principal difficulty in quantifying the movement and storage of water in the subsurface is the natural variability in the physical and hydrologic properties of earth materials at all spatial scales. For convenience, discussion of subsurface hydrology is divided into the unsaturated zone (where open spaces or voids in the earth materials are partly filled with water and partly filled with air) and the saturated zone (where voids in the earth materials are completely filled with water).

Unsaturated Zone

The unsaturated zone, sometimes referred to as the vadose zone or zone of aeration, encompasses the earth materials that lie between the land surface and the water table

(fig. 6). The thickness of this zone varies spatially and temporally and may range from 0 to more than 1,000 m. In general, thicker unsaturated zones are found in more arid regions. No known estimates exist for the amount of water stored in unsaturated zones at the global or continental scales. The importance of the unsaturated zone as a storage reservoir is often overlooked because the water held there generally is not extractable for human use. The unsaturated zone, however, is the primary source of water for vegetation and therefore plays a critical role in the hydrologic cycle. An estimated 76 percent of precipitation infiltrates the subsurface (table 2). Because water moves through the unsaturated zone at a relatively slow rate, plants are able to extract that water over extended periods of time. About 85 percent of the water that infiltrates the soil surface returns to the atmosphere either by evaporation from soil or by plant transpiration.

Water storage within the unsaturated zone is determined by measuring moisture content at different depths between the land surface and the water table. Repeated measurements over time can be used to infer rates of storage change. Moisture content can be measured directly by collecting samples in the field and weighing the sample before and after oven drying. Indirect techniques, which are more conducive to automatic recording, take advantage of electrical or physical properties of the sediment-water continuum (for example, time domain reflectometry and neutron moderation).

Infiltrated water moves predominantly in a downward direction through the unsaturated zone toward the water table. Water also can move upward (in response to evaporative demand) or laterally (in the case of impeding layers of soil). Rates of water movement are notoriously difficult to measure
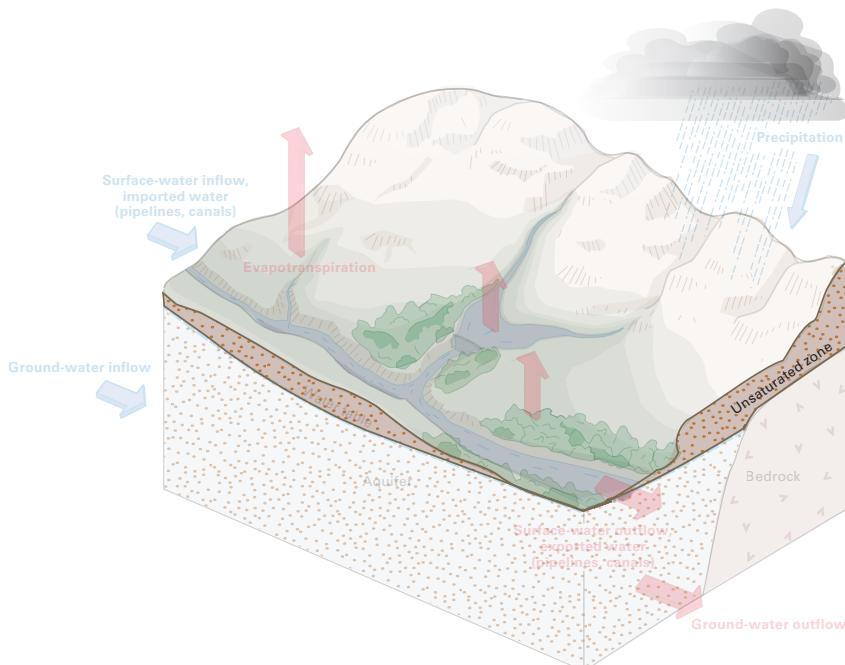
Figure 6 .   The unsaturated zone within the hydrologic cycle.

Water in the unsaturated zone sustains most vegetation.

Installing moisture content sensors through the wall of a trench; the trench was later backfilled.

directly because of problematic measurement techniques and the variable nature of the fluxes. Lysimeters (see Box E—Lysimeters: Water-Budget Meters) can provide accurate, albeit expensive, measurements of these fluxes. More commonly, flux rates are inferred by using indirect approaches such as the Darcy approach or unsaturated-zone water-budget methods. The Darcy approach requires measuring depth profiles of pressure head (sometimes referred to as matric potential or soil-water tension, measured with tensiometers, heat-dissipation or electrical conductivity probes, or thermocouple psychrometers) and unsaturated hydraulic conductivity. Unsaturated-zone water-budget methods are based on measurement of changes in water storage in the unsaturated zone over time (for example, the zero-flux plane method) or analysis of fluctuations in water-table elevations (Scanlon and others, 2002).

Moisture content profiles within the unsaturated zone typically display seasonal trends (fig. 7). Largest fluctuations occur near land surface; the magnitude of the annual fluctua-

tions decreases with depth. At some depth, moisture contents may show no measurable change throughout the year. This does not mean that there is no flow occurring at these sites; rather, this implies a constant flux of water (usually small in magnitude).

Residence times of water within the unsaturated zone depend upon factors such as climate, geology and soils, depth to water table, and vegetation. In most areas, the residence time of water in the root zone ranges from days to months (although some water is maintained in small pores over much longer periods; this is referred to as immobile water, and its presence has been identified through tracer tests). For the region below the root zone, residence times can be estimated as the amount of water stored there divided by the estimated flux through that region. In humid areas with thin unsaturated zones, residence times are usually a year or less. In arid regions, residence times may be millennia.

Figure 7. Hypothetical moisture-content profiles at four different times of the year. As depth increases, the variation in moisture content decreases.

The unsaturated zone at Yucca Mountain, Nevada is as thick as 500 meters.

E   Lysimeters—Water-Budget Meters

Lysimeters are instruments specifically designed for measuring one or more components of the water budget, such as evapotranspiration or ground-water recharge. Most lysimeters consist of containers filled with soil, hydrologically isolated from the surrounding undisturbed environment but intended to mimic the hydrologic behavior of that environment. Lysimeters vary in design from simple collection vessels with a surface area on the order of 100 cm^2 to units constructed on sensitive weighing balances with surface areas of several square meters (Young and others, 1996). Some instruments are capable of resolving fluxes of less than 1 mm/d. When properly constructed and maintained, lysimeters provide perhaps the most sophisticated approach for studying water budgets at a small scale.

Assuming that there is no surface or subsurface flow to it, the water budget for a lysimeter is:

$$\Delta S = P - ET - RO - D \tag{E1}$$

where

ΔS	is change in storage within the lysimeter and is determined on a weight basis,
P	is precipitation and irrigation,
ET	is evapotranspiration,
RO	is runoff,

and

D is drainage out the bottom of the lysimeter.

Installations with weighing lysimeters typically are also equipped with precipitation gages and runoff collectors. In addition, most lysimeters permit measurement and collection of drainage, either by having a free-draining base or by having a porous plate base across which a tension can be imposed by means of a vacuum or wick system. With independent measurements of P, RO, and D, the lysimeter provides a direct measurement of ET:
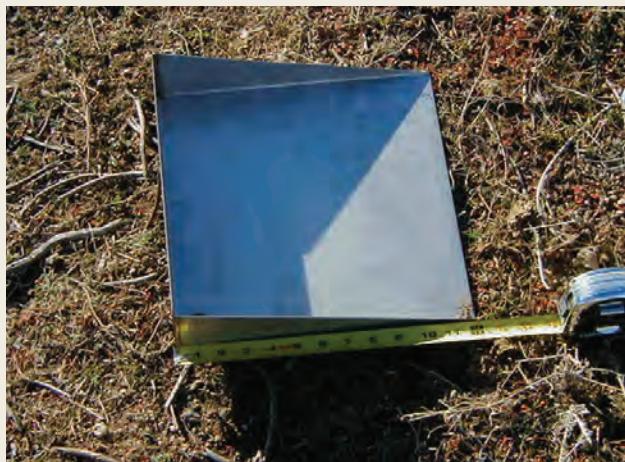
$$ET = P - \Delta S - RO - D \tag{E2}$$

During periods when precipitation, runoff, and drainage are all zero, changes in weight of the lysimeter are due solely to evapotranspiration.

The GSF National Research Center for Environment and Health of the Institute for Soil Ecology, Neuherberg, Germany, operates 32 lysimeters for studies of water budgets, ground-water recharge, and nutrient uptake by plants. (http://www0.gsf.de/eus/index_e.html, accessed on February 26, 2007)

GSF – National Research Center for Environment and Health

Pan lysimeter.

Installing pan lysimeter through trench wall with a hydraulic jack.

Large drainage lysimeters are expensive to construct and often problematic to maintain. As such, they are rarely used in hydrologic studies. Figure *E–1* shows measurements of rainfall and drainage at one such lysimeter at Fleam Dyke, England (Kitching and Shearer, 1982). Small, simple lysimeters are easier to install and maintain and are practical for evaluation of spatial variability of evaporation and irrigation, for example. With any lysimeter, careful design and installation are required to avoid altering the natural hydrologic conditions of the system under study.
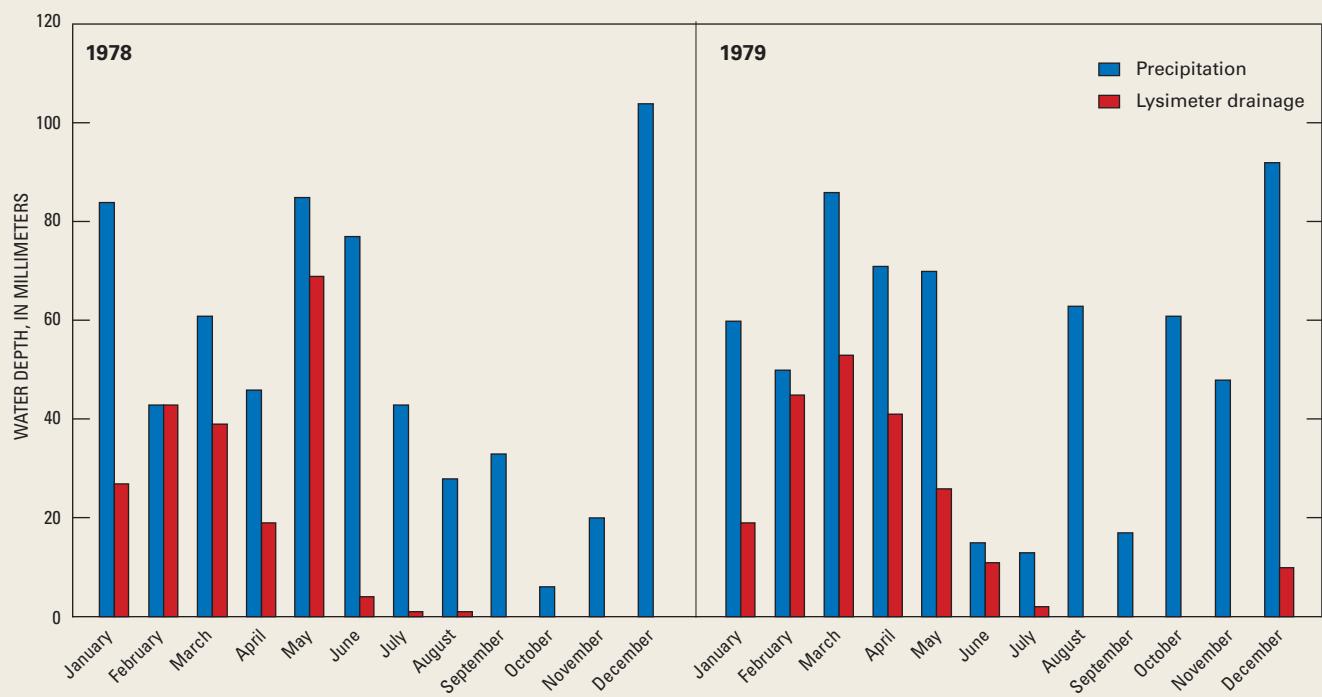
Figure E–1.   *Monthly rainfall and drainage from lysimeter at Fleam Dyke (Kitching and Shearer, 1982).*
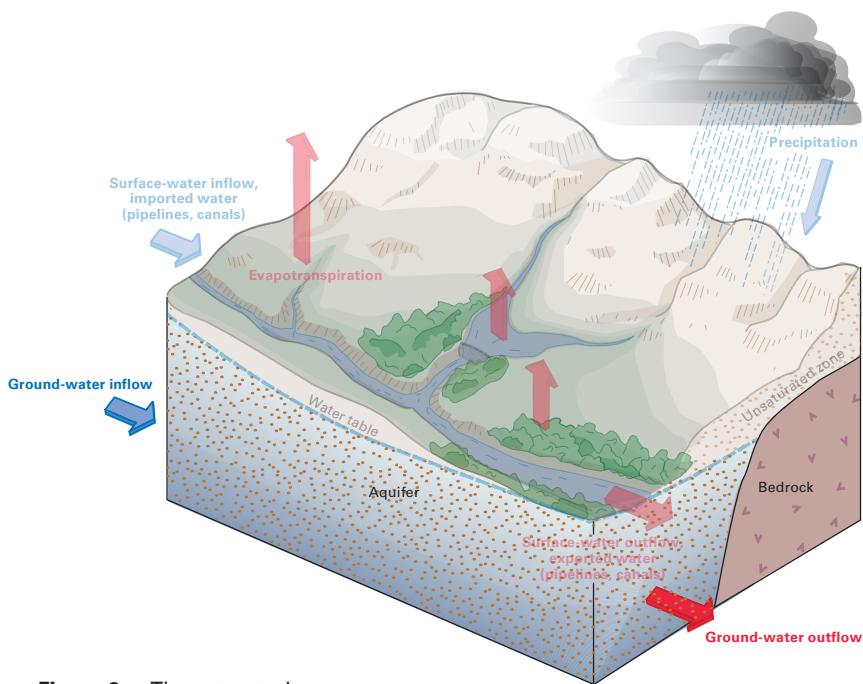
Saturated Zone

Ground water, water stored within the saturated zone, constitutes the largest reservoir of extractable freshwater on Earth (table 1, fig. 8). More than 1.5 billion people worldwide, including about 50 percent of the population of the United States, rely on ground water for their drinking water. The importance of ground water is sometimes overlooked simply because the subsurface is hidden from our view. There are no windows through which we can view the vastness and complexities of the saturated zone.

The saturated zone is bounded above by the water table or by the fixed interfaces at the bottom of surface-water bodies. The lower boundary of the saturated zone is difficult to define. There is a tendency for pores in earth materials to become smaller and fewer with depth, thus limiting the availability of the stored water to humans. Saline ground water underlies fresh ground water in most areas.

Figure 8. The saturated zone within the hydrologic cycle.

Artesian well in Boyd County, Nebraska, circa 1900.

Inflow to the saturated zone, often referred to as ground-water recharge, occurs when water from precipitation (and perhaps irrigation) percolates downward through the unsaturated zone or when water moves from surface-water bodies to the water table (see Box F—Ground-Water Recharge). Outflow from the saturated zone occurs naturally to surface-water bodies (for example, through seeps or springs) and to the atmosphere by evapotranspiration. In humid regions, ground-water discharge to streams is typically the dominant outflow mechanism and can account for more than 90 percent of annual flow in some streams. In arid regions, there may be essentially no ground-water discharge to streams but high rates of ground-water evapotranspiration. In some regions, human extraction of ground water for domestic, agricultural, and industrial uses constitutes the major portion of outflow.

The subsurface is composed of geologic materials of varying chemical and physical properties. Conceptualization of the geologic features and how they affect ground-water flow (fig. 9) is a difficult but fundamental part of ground-water investigations. Insight on boundaries of ground-water flow systems, rates of water movement, amounts of water in storage, and rates and locations of recharge and discharge must be inferred from sources such as geologic maps, geophysical tests, ground-water levels, physical and chemical properties of water and rock, spring and streamflow records, and ground-water-flow models.

An aquifer is a body of earth material that contains sufficient permeable material to yield significant quantities of water to wells. "Significant quantities" is a relative term: pumping rates of 2 m^3/min or greater are considered large rates; 0.04 m^3/min is considered a small rate even though it is more than sufficient to supply the needs of most households.

EXPLANATION

High hydraulic conductivity
Moderate hydraulic conductivity
Low hydraulic conductivity
Direction of ground-water flow
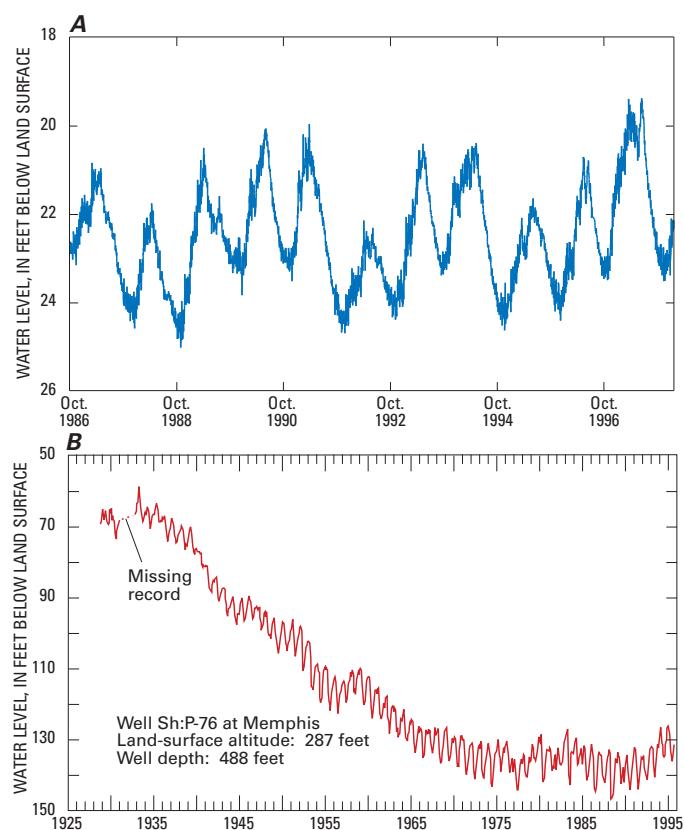
Figure 9.   Ground-water flow systems in complex geological terrain. Ground water in the uppermost part of the ground-water system flows from surface-water body to surface-water body in the high hydraulic conductivity zone (characteristic of glacial outwash) because the water table slopes uniformly from one to the other (**1**). However, where the hydraulic conductivity of the deposits is lower (characteristic of glacial till), water-table mounds beneath the land-surface highs cause the local flow systems to discharge to contiguous surface-water bodies, such that there is no flow from one surface-water body to the next lower surface-water body (**2**). In both settings, intermediate-scale ground-water flow systems pass at depth beneath the local flow systems (**3**), and a regional ground-water flow system passes at depth beneath both local and intermediate flow systems (**4**).

Aquifers in direct connection to the atmosphere are unconfined or water-table aquifers. Confined aquifers are separated from the atmosphere by a confining unit, which consists of materials with a hydraulic conductivity much lower than that of the aquifer. Hydraulic conductivity is a measure of a geologic material's ability to transmit water (see Box G—Estimating Aquifer Hydraulic Conductivity).

Wells are important in ground-water studies. They provide direct access to the subsurface environment and make it possible to measure ground-water levels, to obtain water samples for chemical analysis, to conduct aquifer tests to estimate aquifer properties, and to apply geophysical techniques to estimate physical and chemical properties of earth materials. Long-term measurements of ground-water levels provide data for evaluating trends over time, calibrating ground-water-flow models, and assessing resource-management schemes. Systematic water-level measurements in networks of monitoring wells across the country are conducted by a wide array of organizations. Measurements are made electronically or manually at frequencies ranging from hourly to annually.
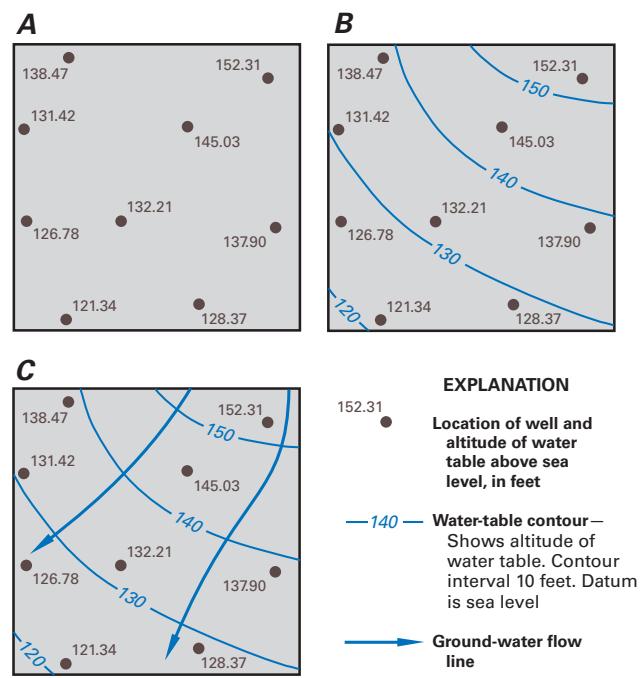
Water levels in many aquifers fluctuate seasonally in response to recharge and discharge patterns. Levels tend to rise from fall through early spring, when precipitation rates usually exceed evapotranspiration rates, and decline in late spring and summer when evapotranspiration rates are high (fig. 10A). The magnitude of fluctuations varies from aquifer to aquifer and year to year depending on geology, water use, and climate. In addition to seasonal patterns, long-term trends occur in some aquifers as a result of changes in climate (for example, droughts) or stresses imposed by humans. Figure 10B shows water levels for a period of almost 70 years from a well located in Memphis, Tennessee. Water levels declined by about 70 ft between 1928 and 1975 as the rate of pumping from the aquifer increased. After 1975, pumping rates stabilized and the long-term decline abated (Taylor and Alley, 2001).

Figure 10.   (A) Hydrograph of daily water-level measurements over a 10-year period for a well in Vanderburgh County, Indiana; and (B) water-level trends in a long-term observation well in Memphis, Tennessee (Taylor and Alley, 2001).

Water-level contour maps of aquifers provide a means by which ground-water movement and storage can be assessed. These maps can be constructed from water-level measurements obtained from multiple wells at about the same time (fig. 11). Alternatively, maps can be based on water levels generated by ground-water-flow models. In general, ground water moves from areas of higher water-level altitudes to areas of lower water-level altitudes. As shown in figure 11*C*, lines drawn perpendicular to water-level contours indicate direction of ground-water flow.

The amount of water stored in the subsurface changes as pores (voids between soil grains) drain or fill and as water and the geologic material compress or expand. Change in aquifer storage between any two points in time is calculated as the product of the difference in water levels at the two times and a storage coefficient, S_c. For unconfined aquifers, gravity drainage and filling of pores is the dominant mechanism for storage change, and the storage coefficient, called specific yield (S_y), has values that range from about 0.02 for fine-grained sediments to 0.35 for very coarse grained sediments. Storage
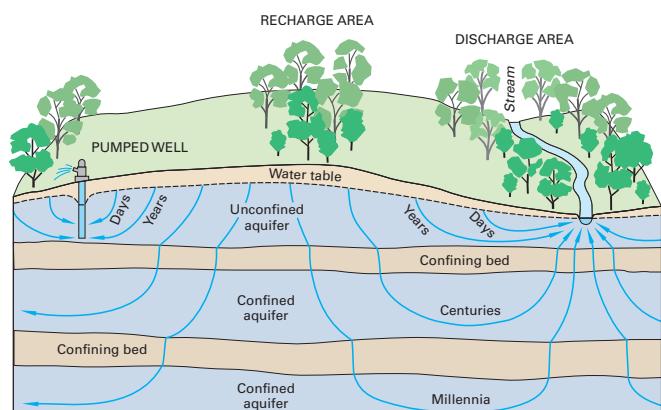
Figure 11. By using known altitudes of the water table at individual wells (*A*), contour maps of the water-table surface can be drawn (*B*), and directions of ground-water flow along the water table can be determined (*C*) because flow usually is approximately perpendicular to the contours (Winter and others, 1998).

Well installation requires special equipment.

changes for confined aquifers are dominated by water and sediment compression and expansion, and storage coefficient values are much less, typically in the range of 10^{-5} to 10^{-4}. Values of storage coefficient are best determined with aquifer tests that integrate over a fairly large area. Laboratory and empirical methods for determining specific yield (Healy and Cook, 2002) are easier to apply but test only a very small sample of an aquifer.

Ground-water flow simulations and chemical, isotopic, and energy tracers are useful tools for identifying ground-water flow paths and estimating traveltimes (fig. 12). Magnitudes of ground-water velocities vary widely. A value of 1 meter per day or greater is considered high. A value of 1 meter per decade is considered low but not unusual for a confining unit. Thus, even for water-table aquifers, the time needed for small parcels of ground water to traverse the aquifer along the longest flow paths from point of recharge at the water table to point of discharge can be decades or longer.

Figure 12.   Ground-water flow paths vary greatly in length, depth, and traveltime from points of recharge to points of discharge in the ground-water system.

Measuring water levels in an observation well in Colorado.

F   Ground-Water Recharge

Ground-water recharge is an important component in aquifer water budgets. Information on recharge rates is useful in assessing the sustainable yield of aquifers, but recharge rates are difficult to quantify accurately because they vary widely in space and time. How, where, and when ground-water recharge occurs depends on factors such as climate, geology, soils, land-use practices, and depth to the water table. Annual rates of recharge in Minnesota (fig. *F–1*) tend to increase from west to east, similar to the trend in annual precipitation. In humid areas, recharge generally occurs as the widespread movement of water from land surface to the water table as a result of precipitation infiltrating and percolating through the unsaturated zone. This type of recharge generally occurs in winter and spring when evapotranspiration rates are low (fig. *F–2A*). In arid regions, focused recharge, or water that percolates down to the water
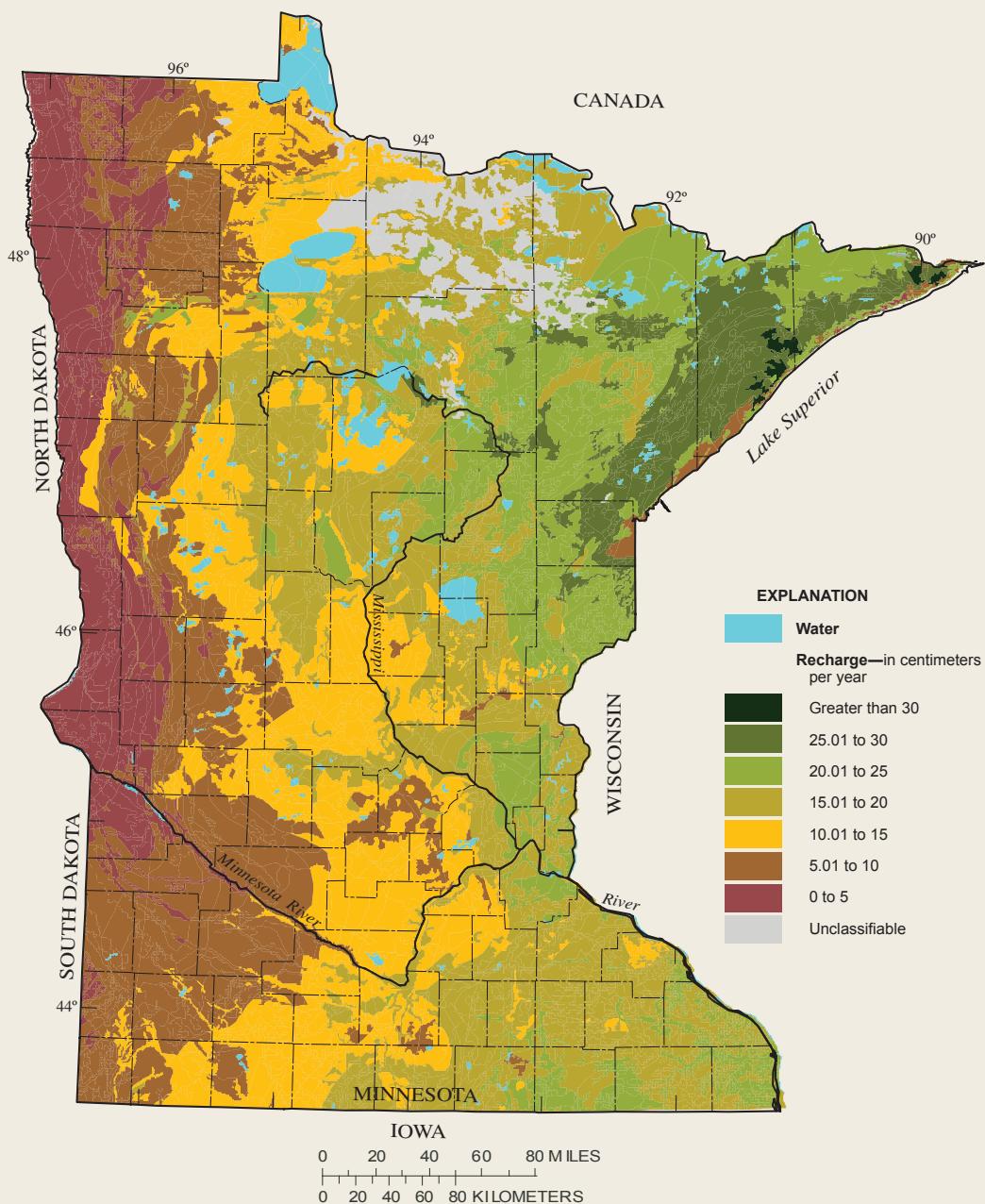
Figure F–1.   *Estimated annual rates of recharge in Minnesota (Lorenz and Delin, 2007).*

table from streams, is usually the dominant mechanism of natural recharge. If those streams drain mountainous areas, then their flow may be mostly snowmelt, and recharge would be expected in spring. Irrigation also can be a source of ground-water recharge during the growing season (fig. *F–2B*).

The water budget of an aquifer provides some insight into what becomes of water that enters an aquifer as recharge (*R*):

$$R = \Delta S^{gw} + Q^{bf} + ET^{gw} + \Delta Q^{gw} \qquad \text{(F1)}$$
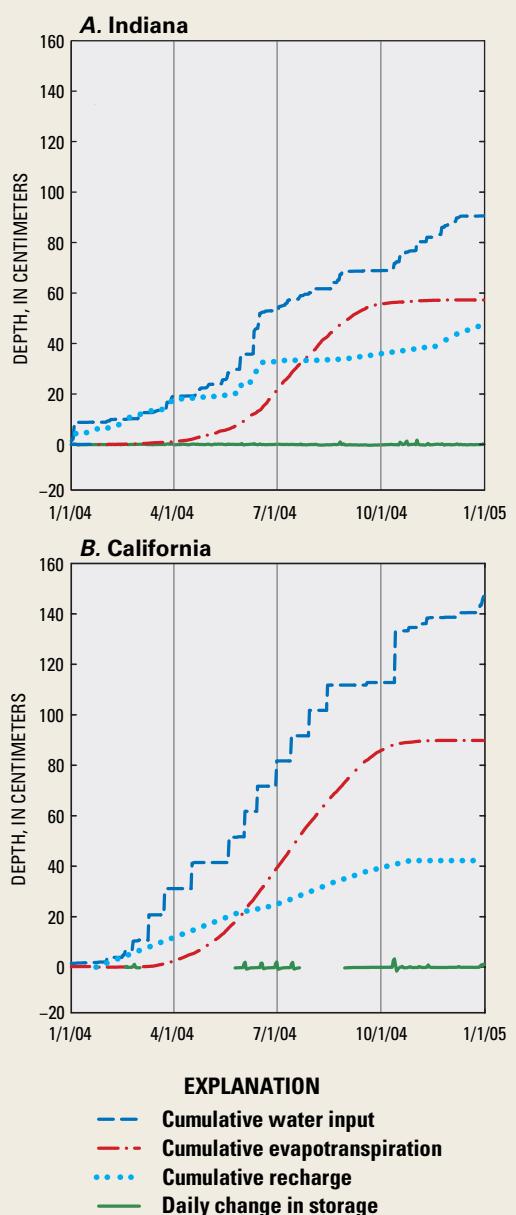
Recharge arriving at the water table augments ground-water storage (ΔS^{gw}), discharges to the surface as base flow (Q^{bf}), is extracted by plant transpiration (ET^{gw}), or moves out of the accounting unit as ground-water flow (ΔQ^{gw}). Often, one of these processes dominates the others. So, for example, measurements of base flow or changes in storage are sometimes used to infer recharge rates. Other methods for estimating recharge are based on physical or chemical data on ground water, water in the unsaturated zone, or surface water (Scanlon and others, 2002).

Local depressions in land surface may be areas of enhanced recharge.

Central Arizona Project
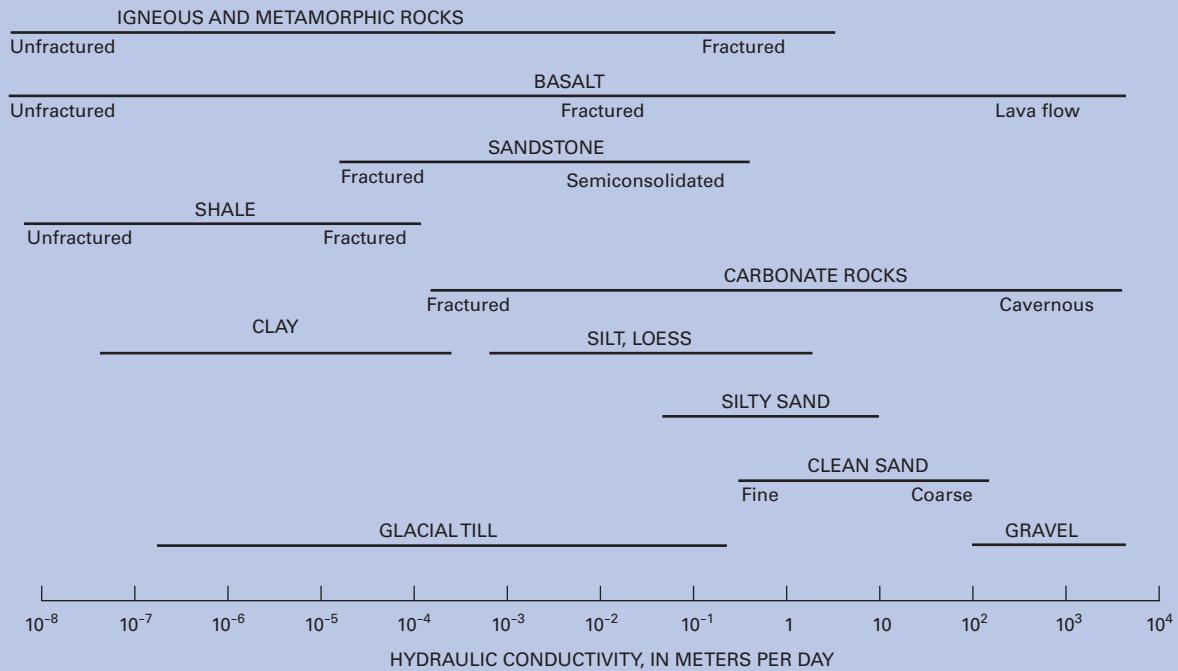
Natural recharge to aquifers can be augmented by induced recharge through spreading basins such as these in the Avra Valley of Arizona.

EXPLANATION

- – – Cumulative water input
- –·– Cumulative evapotranspiration
- ···· Cumulative recharge
- —— Daily change in storage

Figure F–2.   *Water budgets for (A) an unirrigated agricultural field in central Indiana, where most recharge occurs in winter and spring, and (B) an irrigated agricultural field in central California, where most recharge occurs during the growing season (after Fisher and Healy, 2007). Water input refers to the sum of precipitation and irrigation.*

G   Estimating Aquifer Hydraulic Conductivity

Estimation of hydraulic conductivity is an essential, yet problematic, activity in ground-water hydrology. Values of hydraulic conductivity vary over more than 10 orders of magnitude for common earth materials (fig. *G–1*). Very permeable material such as karst limestone, fractured basalts, and coarse gravel can have hydraulic conductivities as large as 1,000 meters per day. Shale, marine clay, and glacial tills, on the other hand, may have conductivities on the order of 10^{-6} meters per day. Difficulties in estimating hydraulic conductivity arise from the highly variable manner in which geologic material was formed and deposited, the limited accuracy with which this parameter can be measured, and the small spatial scale over which measurements are made.

Figure G–1. Approximate ranges in hydraulic conductivity for selected earth materials. A total range of 13 orders of magnitude is shown, which is indicative of the range for more common earth materials. In general, the average hydraulic conductivity of earth material in the same hydrogeologic terrain can vary by orders of magnitude (after Heath, 1983).

Roadcuts and outcrops provide insight to the geologic complexities of the subsurface.
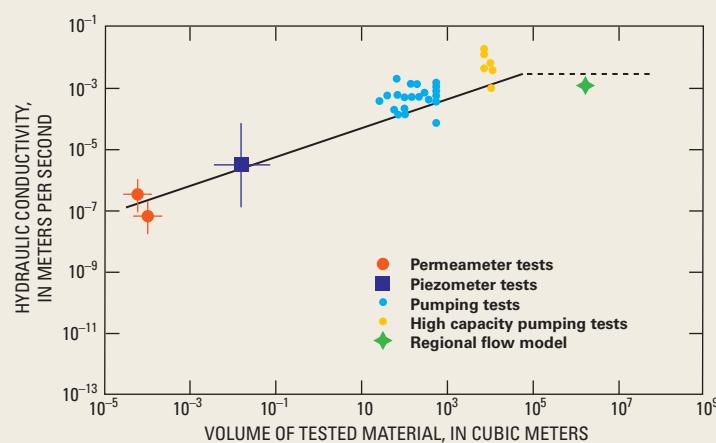
Methods for estimating hydraulic conductivity fall into two categories: direct and indirect. Direct methods are based on hydraulic tests done in the laboratory or the field. Laboratory tests are conducted on sediment cores obtained during drilling of boreholes. The cores are generally several centimeters in diameter and may be a few centimeters to more than a meter in length. Exacting laboratory procedures can produce very accurate measurements of hydraulic conductivity. However, because of the small size of the sample, the representativeness of the measured values to the aquifer as a whole is largely unknown. Field methods, using single or multiple wells, provide estimates that are integrated over larger volumes than those of laboratory tests. The most common single well test is the slug test, whereby a known volume of water is instantaneously withdrawn from (or injected into) the well, and the resulting change in water level in the well is monitored over time. This test samples the aquifer material within perhaps a meter of the well screen. Multiple well tests, sometimes referred to as aquifer or pumping tests, are labor intensive and expensive. One well is pumped and water-level changes are monitored in observation wells at various distances from the pumped well. These tests can run for days or weeks and produce values of hydraulic conductivity integrated over the distances over which drawdown is measured, typically tens to hundreds of meters. Field data are processed by using analytical or numerical mathematical models, models that have inherent assumptions on aquifer boundaries and uniformity, to produce estimates of hydraulic conductivity.

Indirect methods for estimating hydraulic conductivity are generally less expensive than direct methods, although they may not provide the same level of accuracy. The simplest such approach is to consult the literature; many reports contain tables of hydraulic conductivity for various consolidated and unconsolidated earth materials, such as shown in figure *G–1*. Geophysical measurements, made within boreholes, on land surface, or from aircraft or satellites, provide information that can be used to infer values of hydraulic conductivity on the basis of correlations developed at specific sites. Inverse ground-water-flow modeling uses best-fit algorithms to determine parameter values (including hydraulic conductivity) that produce simulated results that most closely match measured water levels and fluxes.

All methods are tied to a distinct spatial scale (fig. *G–2*). Direct measurements, in particular, are made over a relatively small volume. If the aquifer is heterogeneous, many measurements may be required to adequately describe that variability. Indirect approaches may be applied over much larger spatial scales.
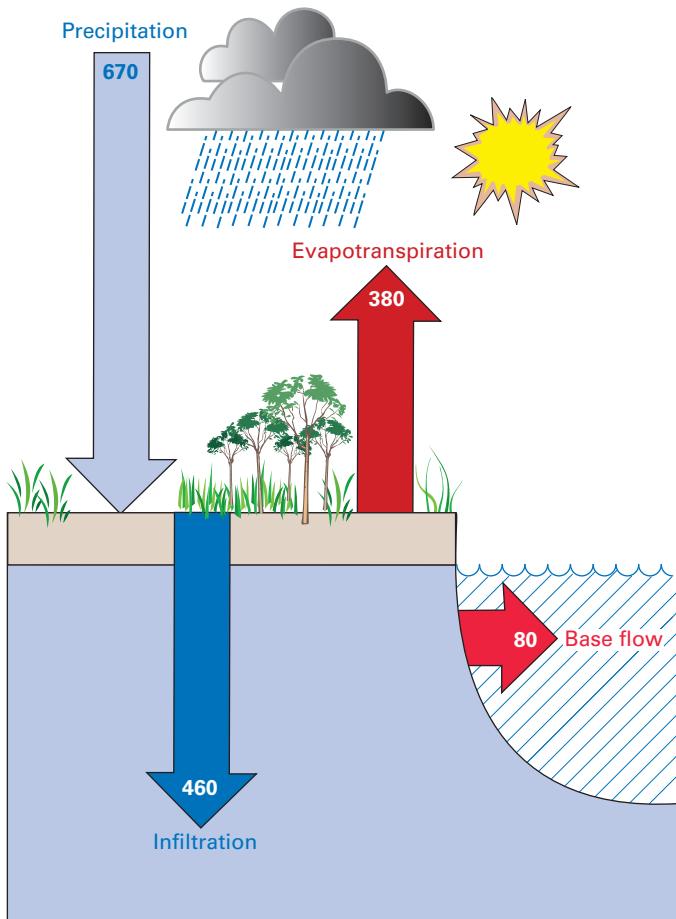
Performing slug tests to determine aquifer hydraulic conductivity.

Figure G–2.   *Hydraulic conductivity as a function of sample size (after Schulze-Makuch and others, 1999).*

Exchange of Water Between Compartments of the Hydrologic Cycle

Water moves from the atmosphere to the surface through the process of precipitation. Water in the subsurface is obtained from land surface either as infiltration of direct precipitation or as seepage from streams or other water bodies. Subsurface water discharges to the surface naturally at springs, streams, lakes, or wetlands and in response to human activities such as a pumping well. Subsurface water also discharges directly to the atmosphere by evapotranspiration. Surface water discharges to the oceans, infiltrates the subsurface, or returns to the atmosphere by evaporation. An understanding of these exchange processes is useful in discussions of water budgets, especially when considering how changes to one process may affect other exchange rates. Figure 13 shows annual rates for North America of water movement from the atmosphere to the land surface (precipitation), from the land surface to the subsurface (infiltration), from the subsurface back to the land surface (base flow), and from the subsurface and land surface back to the atmosphere (evapotranspiration).

Rain storm.

Precipitation

Precipitation, in the form of rain, snow, dew, or fog drip, is the ultimate source of all water on the Earth's landmasses. Average annual precipitation rates vary across the world (table 3). Within the United States, annual rates exceed 10 m in parts of Hawaii and are as low as 50 mm in Death Valley. Figure 14 shows how annual precipitation rates are distributed across the conterminous United States. Precipitation patterns vary with location and season. Although some areas of the United States (such as coastal areas in the Northwest) experience steady, predictable precipitation patterns during some months, precipitation in most of the United States is episodic with no precipitation on most days and rainfall rates up to 100 mm/hr for short periods on other days. Local, convective-cell type thunderstorms can produce several centimeters over an hour in one locale, while no rain may be falling a short distance away. Such variability complicates efforts to determine precipitation rates for any study area.

Figure 13. Average annual water exchange rates for North America (from Lvovitch, 1973) in millimeters of depth per unit area of land surface.

Weighing-bucket precipitation gage.

Table 3. Average annual precipitation rates for various locations (BBC Weather, *http://www.bbc.co.uk/ weather*, accessed on February 12, 2007).
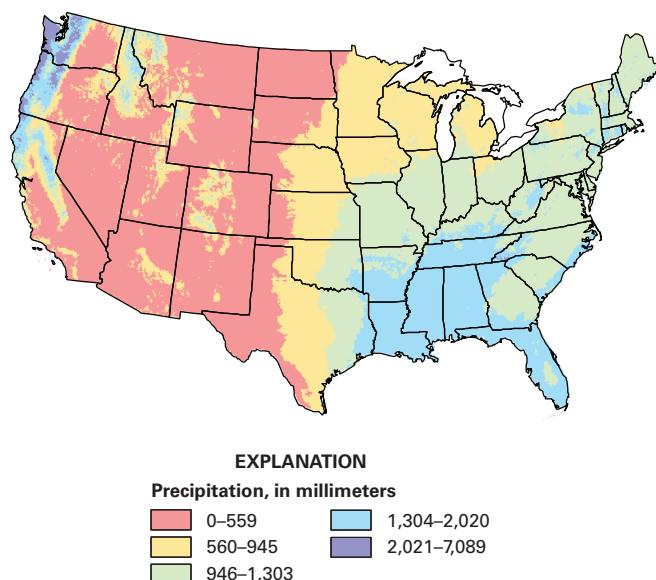
	Precipitation, in millimeters per year
Atlanta, Ga.	1,500
Nashville, Tenn.	1,220
Cleveland, Ohio	1,040
Denver, Colo.	400
Death Valley, Calif.	50
Hilo, Hawaii	3,200
Fairbanks, Alaska	250
Tokyo, Japan	1,570
Singapore, Singapore	2,400
London, UK	580
Cairo, Egypt	25
Barcelona, Spain	580
Sydney, Australia	1,200
Buenos Aires, Argentina	940

The largest source of precipitation data in the United States is the National Climatic Data Center (*http://www.ncdc. noaa.gov*). It holds daily precipitation data for thousands of sites across the country. Precipitation at these sites is measured with a weighing-bucket gage, a cylindrical container with an opening at the top that is 20.32 cm in diameter. Accumulated water in the gage is measured either manually or with an automated sensor that monitors the weight of the container. A standard, manually read gage measures the total precipitation between readings. Another widely used gage is the tipping-bucket gage.

Snowfall is difficult to measure directly. Instead, snow accumulation, or snow depth, is measured. At fixed reporting stations, such as those operated by the National Weather Service or the Natural Resources Conservation Service (NRCS), snow depth is determined by manual observation or by sonic sensors. Some stations are equipped with snow pillows. These are electronic balances that determine the weight of accumulated snow; they can provide hourly information on equivalent water content of the snow. SNOTEL is a network of remote snowpack stations maintained by NRCS in the western United States (*http://www.nrcs.usda.gov*). Using satellite communications, near real-time data on snow accumulation is available for more than 600 sites. A limitation to measuring snow depths at a fixed location is that wind may move the snow around after it has fallen. To compensate for this, agencies such as NRCS make repeated manual measurements of snow depth at multiple points along set courses throughout the snow season.

Uncertainty in precipitation estimates arises from inaccuracies in gage measurements and a limited number of gages. High winds or heavy rainfalls can lead to underestimation of rainfall rates. Gages can get clogged with debris or freeze over. Proper placement of precipitation gages is critical. Nearby objects, such as trees and buildings, may produce a shadow effect, essentially blocking rainfall from the gage.

EXPLANATION

Precipitation, in millimeters

0–559	1,304–2,020
560–945	2,021–7,089
946–1,303	

Figure 14. Average annual precipitation in the conterminous United States for the period 1980–1997 as determined by the DAYMET model (*http://www.daymet.org*).

Snow gage in Idaho.

Agricultural Research Service
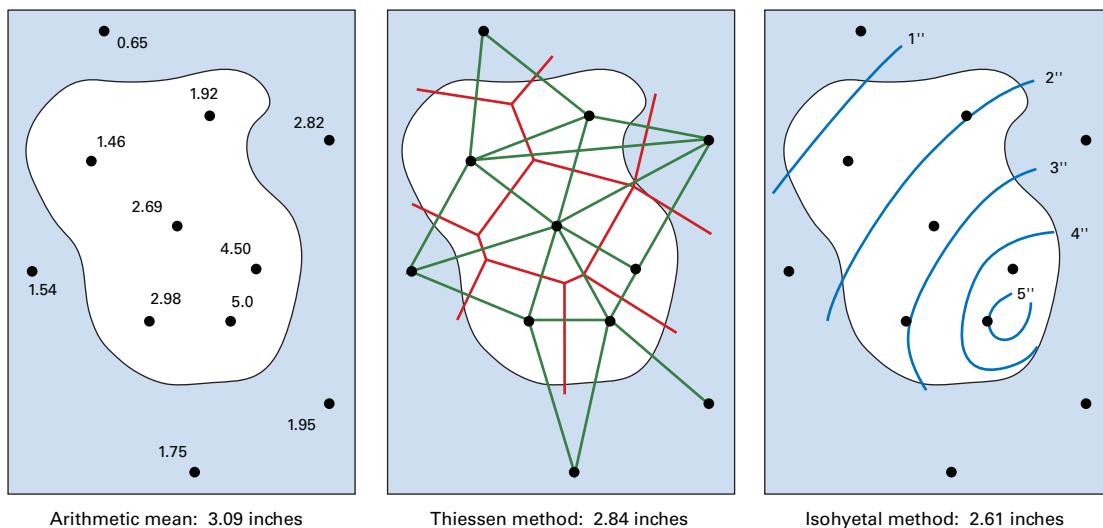
With a collection area of about 324 cm^2, a rain gage samples only a very small part of a landscape's surface area. To generate an average depth of precipitation for a specific area, data from gages within that area can be combined using one of several methods. Results from using three methods (arithmetic mean, Thiessen method, and isohyetal method; Linsley and others, 1982) for a simple example are within 18 percent of each other (fig. 15). Geostatistical techniques, such as kriging, can also be used to integrate precipitation over an area (Seo, 1998). Hevesi and others (1992) made use of the correlation between elevation and precipitation rates to improve estimates of rainfall rates for an area in southern Nevada.

The density of rain gages within a watershed affects the accuracy with which the average rainfall for the entire watershed can be estimated. Increasing the number of gages should increase the accuracy of that estimate, especially for short periods of time. At a site in Illinois, Huff and Schickedanz (1972) found that a gage density of 50 mi^2/gage resulted in a 17-percent sampling error for a 3-hour sampling period and a 6-percent sampling error for a monthly period. National Weather Service weather stations have an average density of 250 mi^2/gage. Studies in mountainous terrain in Idaho indicate that a gage density of about 2 mi^2 per gage is needed to obtain reasonably accurate estimates of precipitation (Molnau and others, 1980).

Fog drip collector in Hawaii. In some areas, fog drip is the primary form of precipitation.

Arithmetic mean: 3.09 inches   Thiessen method: 2.84 inches   Isohyetal method: 2.61 inches
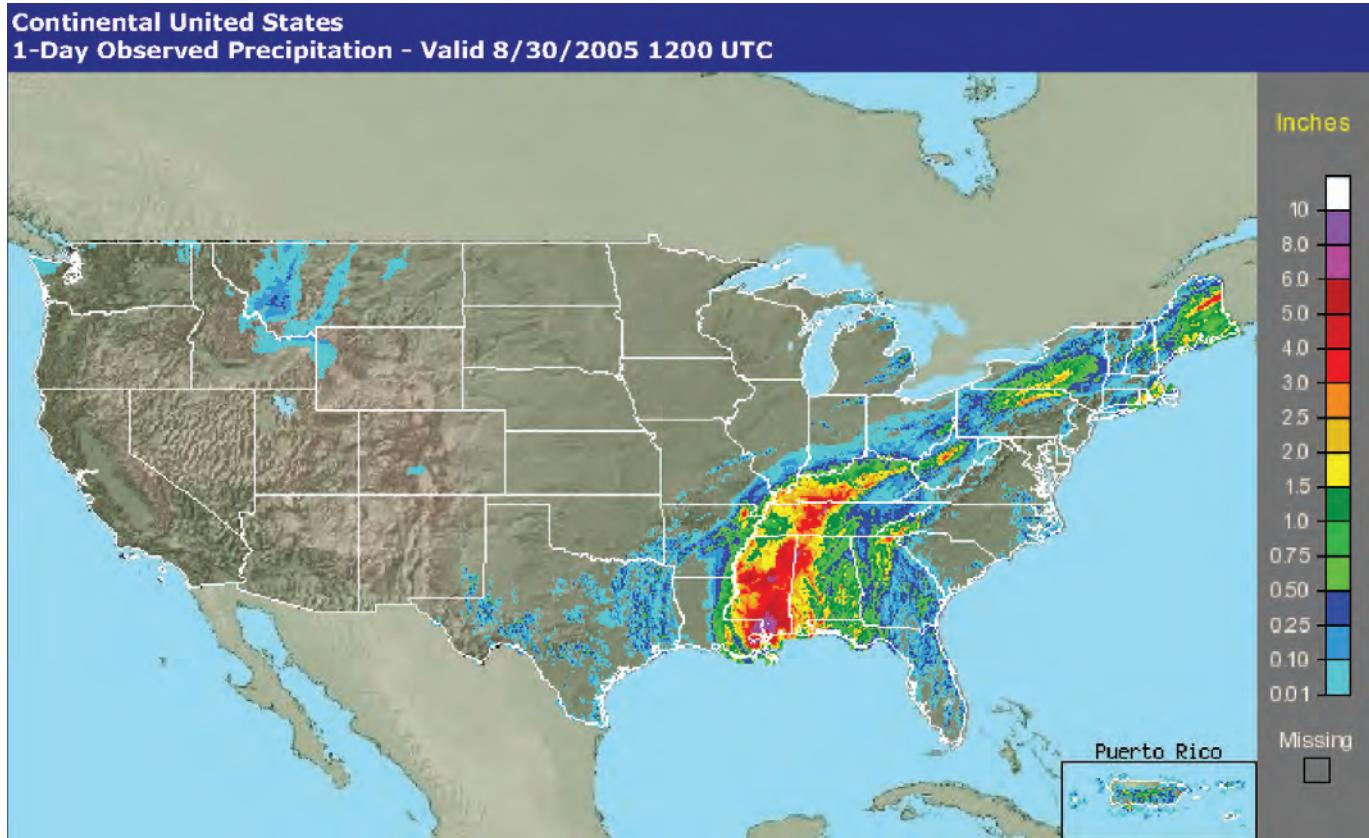
Figure 15.   Estimating average precipitation for an area from precipitation gage records using three methods (after Linsley and others, 1982).

National Weather Service

Doppler radar station.

An alternative method for measuring precipitation, one whose usefulness has yet to be fully integrated into water-budget studies, is Doppler radar. A single radar installation can provide virtually instantaneous estimates of rainfall over areas of more than 70,000 km². The National Weather Service has a radar network that permits estimation of rainfall on a 4-km grid over most areas of the United States, but some biases may impair estimates. A recently developed program called MPE uses a network of real-time standard rain gages and satellite imagery to remove these biases (Seo, 1998; Seo and others, 1999). The program produces hourly to daily total precipitation estimates on a 4-km grid across the conterminous United States (fig. 16). Used with various hydrologic models, these estimates have greatly improved our ability to predict runoff and streamflow, to provide warnings of severe weather and floods, and to manage reservoir systems. These estimates are particularly useful for large river basins and for areas that have few or no standard gages. However, the 4-km grid may not provide sufficient detail for studies conducted on areas of less than a few square kilometers.

Figure 16. Daily rainfall for August 30, 2005 on a 4-kilometer spatial grid as determined by the National Weather Service MPE program. (National Weather Service, *http://www.srh.noaa.gov/rfcshare/precip_analysis_new.php* accessed on Feb. 13, 2007)

Infiltration and Runoff
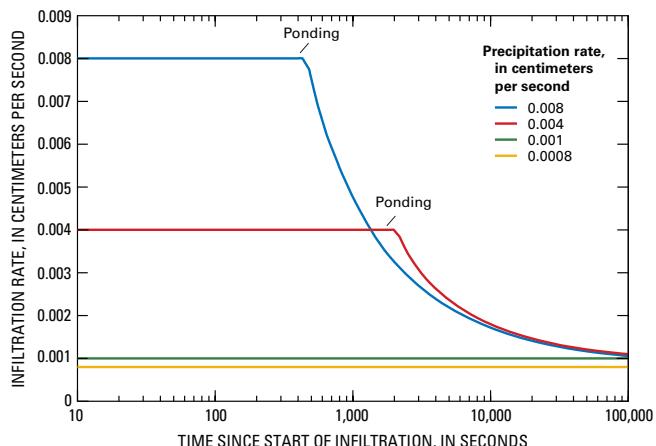
Precipitation falling on land surface can evaporate, be stored on the surface, run off to another point on the surface, or infiltrate the subsurface. Surface storage is mainly in the form of snow. Precipitation falling directly on surface-water bodies or on small surface depressions and precipitation intercepted by vegetation also constitute surface storage. Surface storage is relatively short lived, with the exception of glaciers and ice fields. During periods of rainfall, rates of evaporation and storage change are usually much less than those of infiltration and runoff. If these two processes can be ignored, the sum of infiltration and surface runoff is equal to precipitation. A common practice in many hydrologic studies is to measure precipitation and either infiltration or surface runoff and to calculate the third value by difference. Factors such as soil properties, vegetation, land use, slope, climate (especially precipitation rate and temperature), and water-table depth can affect infiltration and runoff rates.

Figure 17. Infiltration rates generated for a one-dimensional uniform soil column with a variably saturated flow model as a function of time for four precipitation rates. The saturated hydraulic conductivity of the soil is 0.001 centimeter per second.

Infiltrometer.

If the rate of precipitation on bare soil is less than the rate at which the soil can absorb water, then all precipitation will infiltrate. Runoff is initiated once the rate of precipitation exceeds the rate at which the soil can absorb water. The time at which this occurs, the time of ponding, is an important parameter in many hydrologic models. Although there are no theoretical means for determining this time beforehand, empirical equations have been developed for this purpose. Figure 17 shows hypothetical rates of infiltration as a function of precipitation rate and time for a homogeneous soil. When the precipitation rate is greater than the saturated hydraulic conductivity of the soil, the infiltration rate is steady until the time of ponding. It then decreases over time and approaches a magnitude equivalent to the hydraulic conductivity.

Direct measurements of natural infiltration rates are not common but can be made with certain kinds of lysimeters. Infiltrometers are used to determine steady-state infiltration rates after the soil has been saturated (these rates should be similar to values for saturated hydraulic conductivity), but

those rates may be substantially less than early time rates of infiltration into initially dry soils. Indirect methods for estimating infiltration may be based on measurement of pressure head and moisture content at different depths below land surface or on measurements of precipitation and runoff. Empirical methods for estimating infiltration also exist (Chow and others, 1988).

Measurement of surface runoff, sometimes referred to as overland or Hortonian flow, is difficult but has been accomplished in some studies by building berms around a small area to funnel runoff to a central collector or measurement device. Surface runoff flows overland, eventually entering an established stream channel. As discussed previously in the "Streams" section, runoff is only one component of streamflow. It is sometimes possible to analyze a streamflow hydrograph to estimate rates of runoff and base flow (see, for example, Rutledge, 1998). Empirical equations for estimating runoff, most notably the NRCS Curve Number method, are also in widespread use (Chow and others, 1988).

Surface runoff collection apparatus.

Evapotranspiration

Evaporation is the conversion of liquid or solid water into a vapor. It is the process by which water is transferred from a surface-water body or land surface to the atmosphere. Evaporation that occurs through the stoma of plants is called transpiration. In a typical terrestrial setting, it is difficult to measure plant transpiration separately from evaporation from bare soil or water bodies. Therefore, it is common for these two processes to be lumped into a single term—evapotranspiration.

Evapotranspiration, when averaged over one-year periods, is usually second in magnitude among water-budget components to precipitation, representing about 65 percent of precipitation that falls on global landmasses (about 540 mm/yr, table 2). Evapotranspiration may not vary spatially as much as precipitation, but accurate estimates of evapotranspiration are generally more difficult to obtain. There is no national network of evapotranspiration monitoring sites within the United States such as exist for precipitation and streamflow. However, at select National Weather Service sites, pan evaporation (evaporation from 1.2-m-diameter Class A pan) is measured daily.
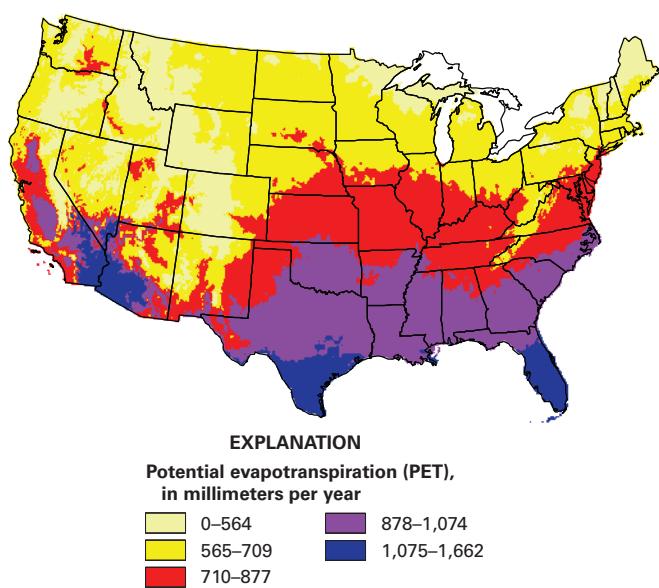
As the common link between the water and energy budgets, evapotranspiration is dependent upon the availability of both water and energy. In arid regions, water availability is the major limitation on evapotranspiration rates. In humid regions, there is generally an excess of water relative to available energy, so rates are energy limited. Evapotranspiration rates generally follow a trend similar to that of net radiation: highest in the summer and lowest in winter. The importance of energy on evapotranspiration rates is also apparent on a daily time scale. Rates are essentially 0 during night hours when no solar radiation is arriving at the site and highest during the daylight hours of peak net radiation.
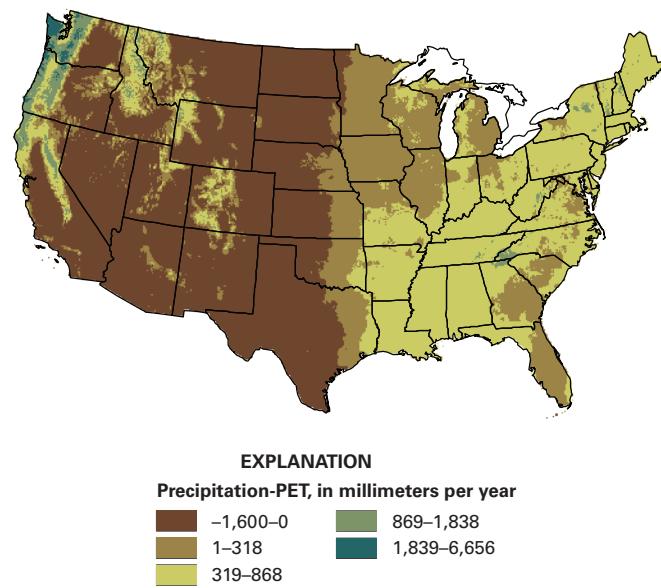
Potential evapotranspiration is the evapotranspiration that would occur if water were plentiful. Figure 18 shows estimates of annual potential evapotranspiration rates for the conterminous United States as calculated with the Hamon method using average temperatures from 1961 to 1990 generated by the PRISM model (Daly and others, 1994) . Estimates of potential evapotranspiration are used in the planning and management of irrigation systems. The difference between precipitation and potential evapotranspiration (fig. 19) provides an index of areas where evapotranspiration is water limited (differences less than 0) and energy limited (differences greater than 0).

Measurement of evapotranspiration rates at specific locations are made with lysimeters or micrometeorological techniques (Rosenberg and others, 1983). These latter techniques, which include eddy correlation, Bowen ratio/energy budget, and aerodynamic profile methods, measure or estimate the vertical flux of water vapor from land surface to the atmosphere. These methods may provide accurate estimates of evapotranspiration rates, but lysimeters and micrometeorological instrumentation are expensive and delicate and require frequent maintenance.

Climatological methods (Rosenberg and others, 1983) provide estimates of potential evapotranspiration. Although not as sophisticated as the micrometeorological techniques, they are much easier to apply, usually requiring only data that are available from National Weather Service stations (for example, daily temperature, relative humidity, or solar radiation). Included in this class are the Thornthwaite,

EXPLANATION

Potential evapotranspiration (PET), in millimeters per year

- 0–564
- 565–709
- 710–877
- 878–1,074
- 1,075–1,662

Figure 18 .   Annual potential evapotranspiration rates across the conterminous United States as calculated with the Hamon equation.

EXPLANATION

Precipitation-PET, in millimeters per year

- −1,600–0
- 1–318
- 319–868
- 869–1,838
- 1,839–6,656

Figure 19.   Difference between annual precipitation and potential evapotranspiration rates across the conterminous United States.

Jensen-Haise, Hamon, and Penman-Monteith methods. Actual evapotranspiration can be estimated from potential rates by application of a correction factor. In the agricultural literature, this factor is referred to as the crop coefficient. The crop coefficient is a related to crop type and maturity and climate; values for different crops and some native vegetation can be found in Jensen and others (1990). Obtaining estimates of actual evapotranspiration from pan evaporation rates requires application of a second correction factor called a pan coefficient (Doorenbos and Pruitt, 1975) as well as a crop coefficient.

Other techniques for estimating evapotranspiration deserve mention even though they are not yet as widely used. Light detection and ranging (LIDAR) is a ground-based system capable of making accurate measurements of sensible and latent heat fluxes over areas as large as several hectares. The expense of the LIDAR equipment limits its use to select research studies. Satellite and aerial remote sensing offers no direct method of measuring evapotranspiration. However, progress has been made in correlating point micrometeorological measurements with variables that can be mapped from space, such as vegetation type and cover (Liu and others, 2003), soil moisture, and surface temperature (Quattrochi and Luvall, 1999). Models incorporating these variables can be used to generate regional evapotranspiration estimates from the point measurements.

Tower to measure evapotranspiration in a forest.

Instrumentation to measure evaporation from a lake.

Instruments used in eddy correlation measurements of evapotranspiration.
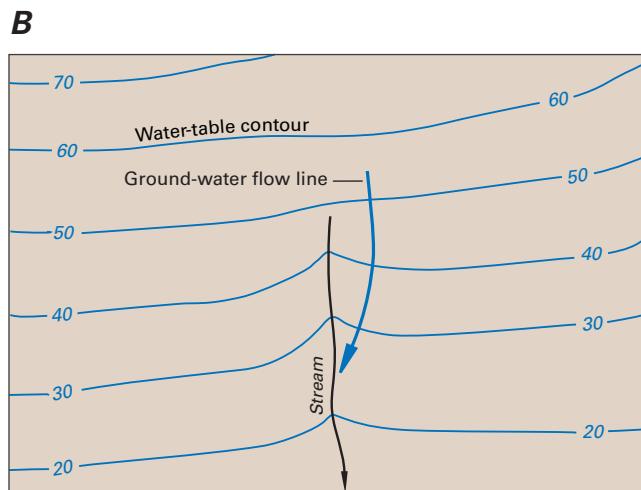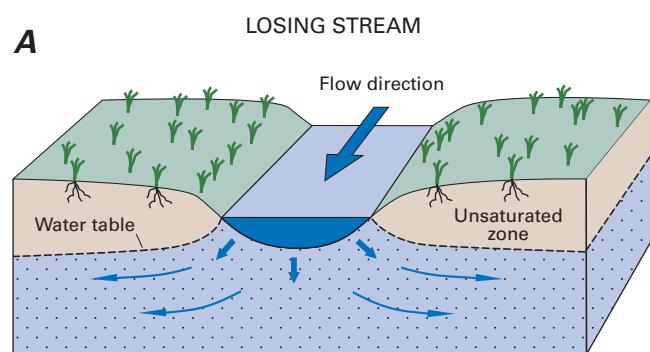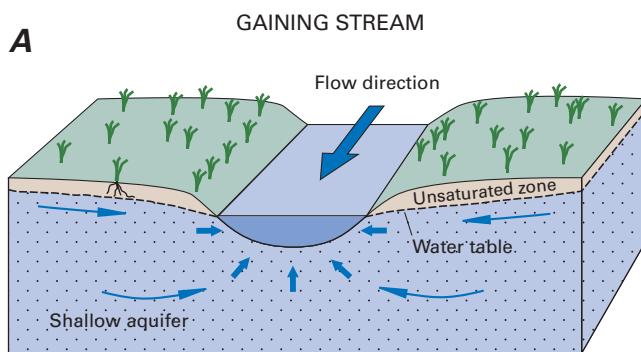
Exchange of Surface Water and Ground Water

Streams and ground-water bodies exchange water in all types of hydrologic settings. Streams gain water from inflow of ground water through the streambed and banks (gaining stream, fig. 20A); they lose water to ground water by outflow through streambed and banks (losing stream, fig. 21A). A stream can be 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 altitude 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 (fig. 20B) and losing streams by pointing in a downstream direction (fig. 21B).
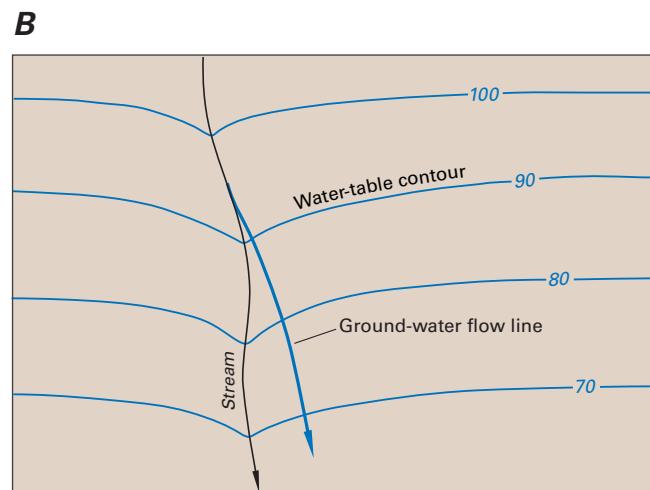
Losing streams can be connected to an aquifer by a continuous saturated zone (fig. 21A) or separated from it by an unsaturated zone (fig. 22). Ephemeral streams flow only in response to snowmelt and storms. Generally, ephemeral streams are not directly connected to an aquifer. However,
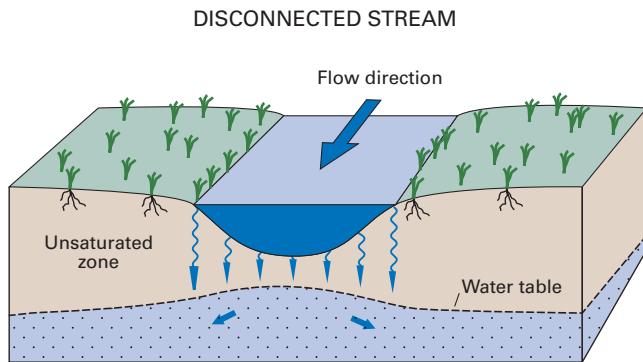
Big Spring, Missouri.

Figure 20.   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 21.   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*).

DISCONNECTED STREAM

Figure 22.   Disconnected streams are separated from the water table by an unsaturated zone.
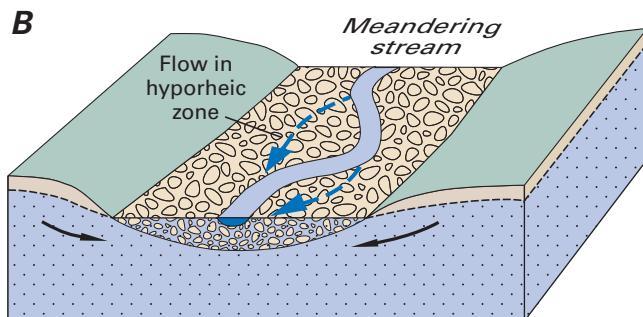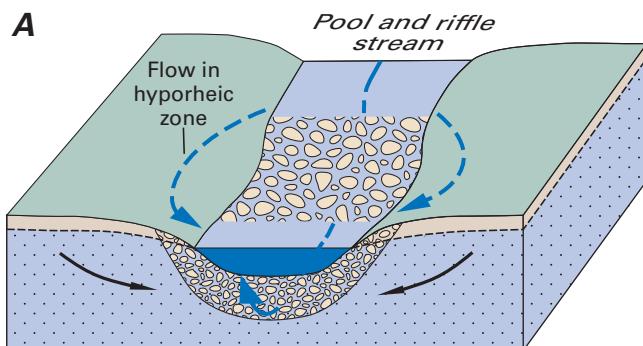
during periods of flow, these losing streams can be important sources of ground-water recharge.

In many stream settings, surface water flows through short segments of the streambed and banks and back into the stream. These segments, which may exist along the entire reach of a stream, constitute the hyporheic zone (figs. 23 and 24). Ground water and surface water mix within the hyporheic zone, providing a unique environment for important biological and chemical reactions. The size and geometry of hyporheic zones surrounding streams vary in time and space. Streams flowing over sand and gravel may have hyporheic zones up to 2 m thick.
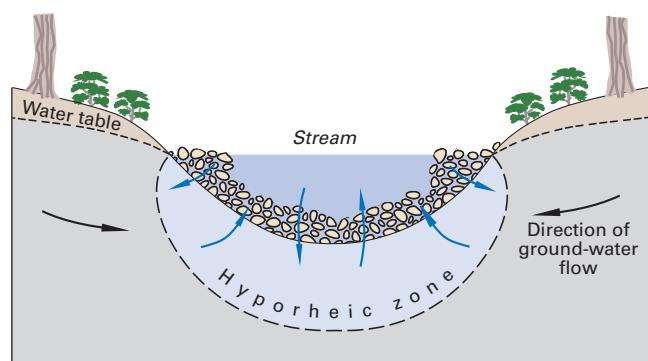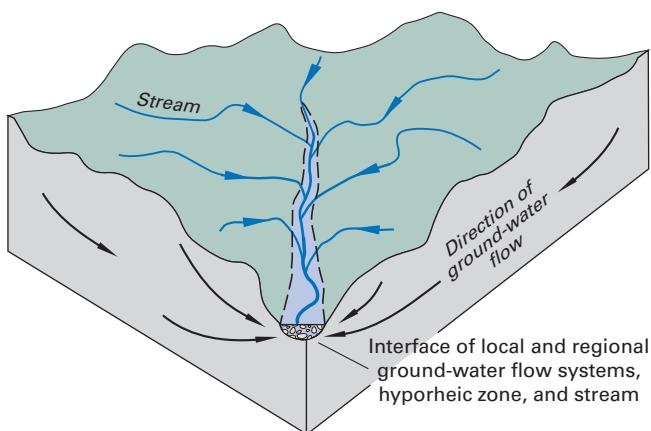
Bank storage, the temporary storage of stream water in the subsurface (fig. 25), occurs when stream stage rises as a result of precipitation, snowmelt, or release of water from a reservoir upstream. As long as the rise in stage does not overtop streambanks, most stored water returns to the stream a few days or weeks after the stage returns to normal level. Bank storage tends to reduce flood peaks and supplement streamflow when stage recedes. 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. 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 even years. Many stream-aquifer systems are in continuous readjustment from exchanges of water related to bank storage and overbank flooding.
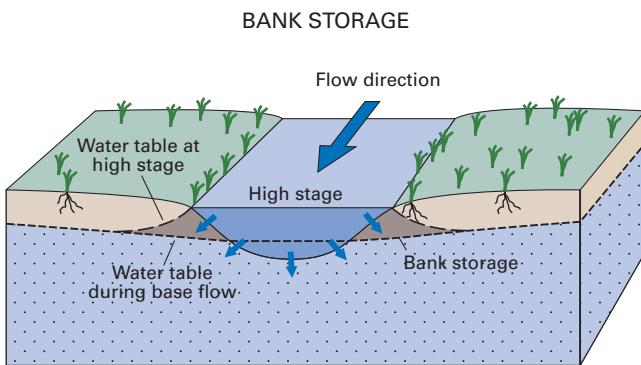
Lakes and wetlands interact with ground water in a manner similar to that of streams. There are some differences, though. Evaporation is generally a larger component of the water budget for a lake or wetland than for a stream. Bank storage is usually of minor importance because water levels do not fluctuate as much as they do in streams. Important exceptions are surface-water reservoirs in arid and semiarid regions.

Figure 23.   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 24.   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.

BANK STORAGE

Figure 25. If stream levels rise higher than adjacent ground-water levels, stream water moves into the streambanks as bank storage.

Reservoir stage can change substantially over the course of a year, rising as spring rain or snowmelt fills the reservoir and falling through summer and fall as water is distributed to users. Wetlands may be present in many different parts of the landscape, whereas lakes and streams occupy local topographic low regions.

Ground water contributes to many lakes, wetlands, and streams. Ground-water discharge can account for more than 90 percent of total annual streamflow (table 4). These contributions can be critical to the maintenance of diverse ecosystems. Wetlands and riparian zones provide wildlife habitat, mitigate floods, and process nutrients and contaminants; the existence of these areas may depend on a steady discharge of ground water. Understanding the water budget for these areas can aid in assessing how changes to one water-budget component will affect other components. For example, an understanding of the water budget could help determine if diversion of ground water to a domestic water-supply well will reduce the rate of ground-water discharge to a wetland and, if so, what effect that would have on plants and wildlife.

Table 4. Base flow as a percentage of total streamflow for selected streams across the United States (Winter and others, 1998).

Stream	State	Percentage of ground-water contribution
Dismal River	Nebraska	94
Forest River	North Dakota	13
Sturgeon River	Michigan	90
Ammonoosuc River	New Hampshire	61
Brushy Creek	Georgia	68
Homochitto River	Mississippi	36
Dry Frio River	Texas	58
Santa Cruz River	Arizona	35
Orestimba Creek	California	23
Duckabush River	Washington	65

Stream disappearing into sinkhole in karst terrain in Texas.

Techniques are available for estimating exchange rates between surface and ground waters over various space and time scales. Seepage meters provide point measurements over an area of about 1 m^2 for periods of seconds to days. Discharge measurements can be made at different locations along a reach of stream; the difference in discharge between any two points will be equal to the net stream loss or gain along that reach. At the watershed scale, hydrograph separation methods and streamflow duration curves can be used to estimate base flow in gaining streams. Solute- and energy-budget approaches have been used over a variety of scales to estimate exchange rates of ground water with lakes and streams.

Seepage meters are used to measure the exchange of water between surface water and the subsurface.

Water-Budget Studies

Space and time scales associated with water-budget studies largely determine the appropriateness of methods for estimating fluxes and changes in storage. Different methods are applicable over different scales. Some techniques provide estimates at a single point in space, such as the use of standard rain gages. Because rainfall rates vary with location, multiple gages may be needed to determine an average rate for a watershed. Other techniques provide estimates that are integrated over large areas (for example, a stream-discharge measurement provides an estimate of runoff for the entire area that drains to the measurement point). Time scales are also important. For estimating ground-water recharge, the water-table fluctuation method provides an estimate for each recharge event, of which there could be many during a year. Ground-water age-dating techniques, on the other hand, provide a single estimate of recharge that is averaged over several years or decades. Prudence dictates that the time and space scales of measurement and estimation methods match the needs of the water-budget study at hand.

Four intensive water-budget studies are presented in this section to illustrate the different approaches and exacting procedures that have been applied in studies of water budgets. Detailed studies such as these are not commonly undertaken, mostly because of monetary and time constraints. The examples convey the level of complexity inherent in conducting such studies and show that results from even the most detailed studies of water budgets in natural hydrologic systems contain some uncertainty. This uncertainty arises from the natural variability in hydrology, geology, climate, and land use and inaccuracies in the techniques used to collect and interpret data.

Water Budget for a Small Watershed: Beaverdam Creek Basin, Maryland

The Beaverdam Creek basin is situated on the Atlantic Coastal Plain in the Delmarva Peninsula of Maryland (fig. 26). The water budget of the basin was studied for 2 years in the early 1950s (Rasmussen and Andreasen, 1959) to determine the apportionment of precipitation among ground-water recharge, subsurface runoff to ponds and streams, ground-water evapotranspiration, and ground-water storage. The 19.5-mi^2 drainage basin ranges in elevation from 10 to 85 ft above sea level and receives on average 43 inches of precipitation annually. In the subsurface, Quaternary-age surficial sands and silts, as much as 70 ft thick, overlie aquifers of Tertiary-age sand. The water table is generally within 12 ft of land surface. The study area was in a natural setting, largely unaffected by human activity. Understanding such a natural hydrologic system is fundamental to evaluating human influences on this and other systems.

The amount and kinds of data collected are unusual for a small watershed. Ground-water levels were measured weekly in 25 observation wells. Stream discharge was monitored
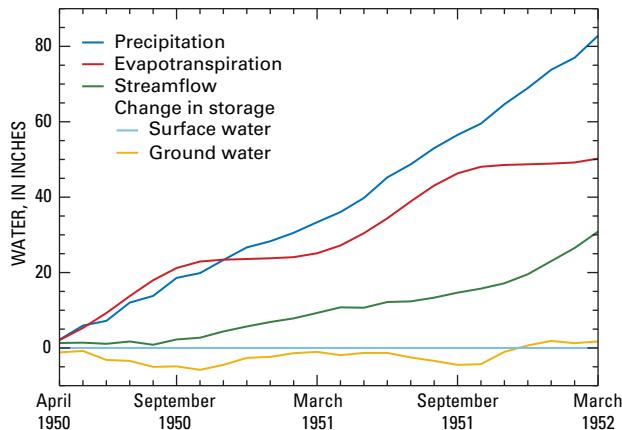
Typical stream in Delmarva Peninsula, Maryland.

by means of a sharp-crested weir at the outlet of the basin. Changes in water storage in the two ponds within the basin were calculated from stage readings and a table (developed by bathymetric survey) that related pond volume to stage. Twelve precipitation gages were deployed across the basin and monitored on a weekly basis. A 4-ft Class A evaporation pan was part of a weather station that also included an anemometer, barometer, wet- and dry-bulb thermometers, and a thermistor for measuring soil temperature. Soil moisture content was determined weekly by electrical resistance (individually calibrated Bouyoucos blocks) at depths of 4, 12, and 39 inches at three locations within the basin.

Major components of the water budget are shown in figure 27. Precipitation for the 2-year period totaled 83 inches. Sixty percent, or 50 inches, was returned to the atmosphere through evaptranspiration; 31 inches (37 percent of precipitation) left the basin as streamflow. Two inches remained in the basin, augmenting surface and subsurface storage. The quasi-linear nature of the precipitation curve in figure 27 indicates relatively uniform precipitation rates throughout the study

Figure 26.   Location map for Beaverdam Creek basin.

Figure 27. Water budget for Beaverdam Creek Basin, April 1950 to March 1952.

Farm pond, Delmarva Peninsula.

period. There is no indication of the seasonality in precipitation that is common in many regions. Evapotranspiration, on the other hand, shows a distinct seasonal trend, with highest values occurring in summer months and negligible rates for most winter months. Evapotranspiration was not measured directly. All other water-budget components were measured or estimated independently. Evapotranspiration was then determined to balance the water-budget equation.

Ground-water recharge, the only source of inflow to the shallow ground-water system, was calculated by application of the water-table fluctuation method, using weekly average water levels from the 25 observation wells. Recharge was then partitioned into its different components:

$$R = \Delta S_{gw} + Q_{bf} + ET_{gw}$$

where:

R is recharge,
ΔS_{gw} is change in ground-water storage,
Q_{bf} is base flow,

and

ET_{gw} is evapotranspiration from ground water.

Change in ground-water storage was calculated from the difference in head between the end and the beginning of the week. Base flow was determined by a stream hydrograph separation method. Evapotranspiration from ground water was then calculated as the residual of the equation. Figure 28 shows plots of calculated recharge components. For the 2-year period, ground-water recharge was 42.6 inches and was partitioned into 21.5 inches of base flow, 1.7 inches of increase in ground-water storage, and 19.5 inches of evapotranspiration of ground water. In summer months, evapotranspiration is the largest draw on ground water. For the rest of the year, base flow is the predominant mechanism of ground-water discharge.

Few studies before the Beaverdam Creek watershed study devoted as much effort to the comprehensive examination of the water budget of a small watershed. Measurements of water-budget components can be made more easily and more accurately with the improved instrumentation that has been developed over the decades since the original study was conducted. Even so, conducting a similar study today would require a substantial commitment of funding and manpower.

Agriculture is the main industry in the Delmarva Peninsula.

Figure 28. Ground-water budget for Beaverdam Creek Basin, April 1950 to March 1952.

Soil-Water Budgets for Prairie and Farmed Systems in Wisconsin

Water budgets of soil zones are important for management of agricultural fields. They also are used to estimate rates of evapotranspiration and ground-water recharge. Agricultural practices can greatly affect soil-water budgets. Irrigation and cultivation techniques (conventional, minimum, or no tillage) can influence surface runoff, erosion, infiltration, and transport of applied agricultural chemicals. Brye and others (2000) determined water budgets for a 132-week period in 1995–98 at three sites in Columbia County in southern Wisconsin: a restored natural prairie, maize under no-tillage, and maize with chisel-plow tillage (fig. 29). The objectives of the study were to evaluate the usefulness of newly designed instrumentation and to assess the effects of agricultural practices on drainage beneath the root zone.

Columbia County, Wisconsin

Figure 29.   Location map of Columbia County, Wisconsin.

K.R. Brye

Agricultural site.

Precipitation was measured with rain and snow gages. Soil-moisture content profiles were measured weekly during the growing season and at 3-week intervals during winter with a neutron moisture meter. Readings were taken to a depth of 1.4 m in four access holes in each field. Water movement through the unsaturated zone is very difficult to measure directly. In this study, drainage beneath the root zone was measured with equilibrium tension lysimeters (ETLs) (Brye and others, 1999). These novel devices have a porous stainless steel surface that allows collection and measurement of drainage. Soil-matric potential sensors and a vacuum system allowed the tension within the ETL to be set slightly greater than that recorded in the bulk soil surrounding the lysimeter. Thus, flow into the ETLs should be similar to natural drainage rates. The ETLs were installed through a 2-m-deep trench; the 75-cm by 25-cm top surface was set at a depth of 1.4 m. Evapotranspiration (*ET*) was equal to the differences in inputs and outputs and storage changes:

$$ET = P - RO - D - \Delta S$$

where:

P	is precipitation,
RO	is surface runoff,
D	is drainage below the root zone,

and
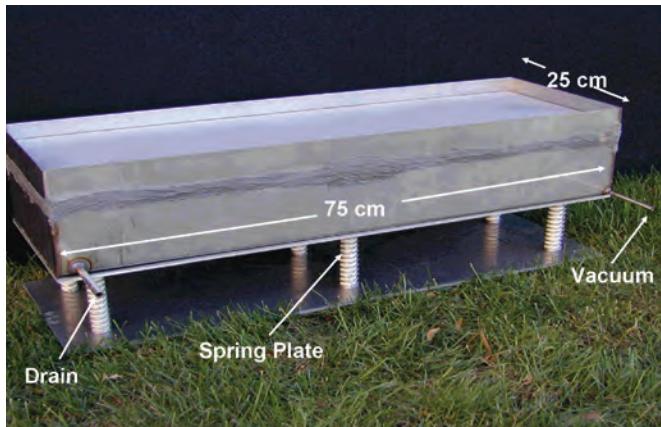
ΔS	is change in storage (soil and surface).

Runoff only occurred immediately after large precipitation events or snowmelt and was estimated by a procedure described by Brye and others (2000). Calculations were made on a weekly basis.

K.R. Brye

Prairie site.

Equilibrium tension lysimeter.
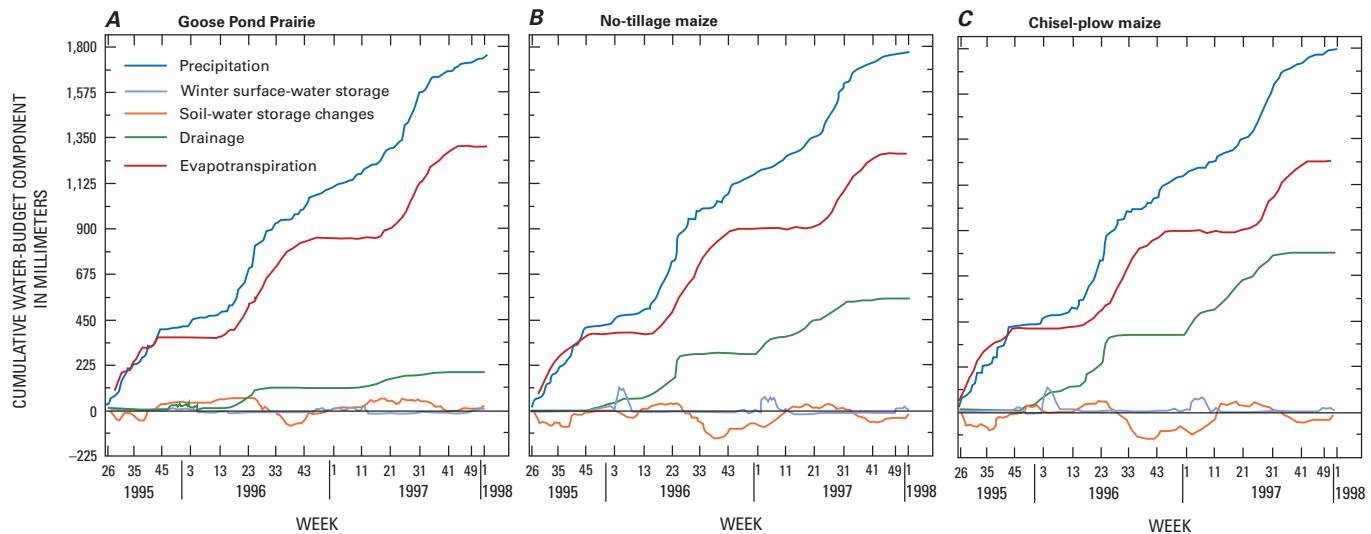
The lysimeter requires careful installation.

Water budgets for the three sites are shown in figure 30. Precipitation rates were similar for the three sites. The prairie site had greater soil-moisture contents, more evapotranspiration, and less drainage compared to the maize fields at the other two sites. Evapotranspiration rates for the prairie site were slightly greater than those from the no-tillage site, which were slightly greater than those from the chisel-plow site. Drainage occurred from late January to mid-June at all sites. However, drainage totals for the 132-week period were substantially different: 199 mm for the prairie, 563 mm for no-tillage, and 793 for chisel-plow. It appears that infiltration

rates increased with increasing disturbance of the land surface. Runoff, because of its episodic nature, is not included in figure 30. For the study period, runoff totaled 197 mm at the Prairie site, 182 mm at the no-tillage site, and –5 mm (due to drifting snow) at the chisel-plow site.

Rates of infiltration, drainage, and evapotranspiration are important in terms of plant health and ground-water quantity and quality. It is difficult to accurately determine the effects of different agricultural practices on local water budgets, but new and improved instrumentation, such as that used in this study, can provide valuable insight into complex processes.

Figure 30.   Water budget for (*A*) prairie site; (*B*) no-tillage maize site; and (*C*) chisel-plow maize site in central Wisconsin (Brye and others, 2000). Imbalance in the water budgets is attributed to runoff for the prairie and no-tillage site and to runon from melting snow at the chisel-plow site.

Water Budget of Mirror Lake, New Hampshire

The hydrology and chemistry of Mirror Lake (fig. 31) have been studied since the late 1970s to develop an understanding of the hydrological processes associated with the lake and to examine uncertainties in estimates of water and chemical budgets for the lake (Winter, 1984). To accomplish this, all components of the water budget were determined independently, either by direct measurement or by calculation from measurements of environmental variables. Precipitation was measured at two gages within 400 m of the east and west shores of the lake. Evaporation was calculated by using an energy-budget method with data from meteorological instruments located on a raft in the lake and a land station near the lake. Stream inflow and outflow were measured by using Parshall flumes and weirs equipped with stage recorders. Ground-water inflow and outflow were calculated by using Darcy's equation with water levels in the lake and in numerous wells near the lake together with measured hydraulic conductivity. Water storage in the lake was estimated from measured lake stage and a stage-volume relation.
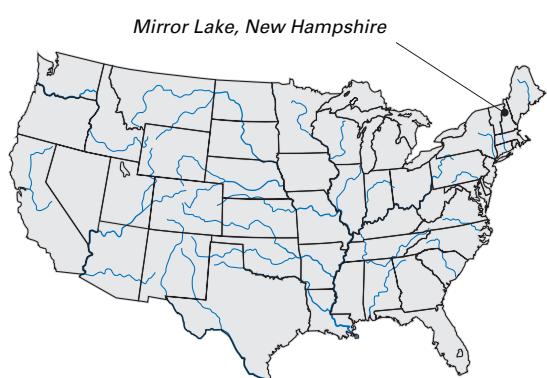
Monthly and annual water and chemical budgets were determined for Mirror Lake for the 20-year period from 1981 to 2000. Streams provided the largest inflow of water to the lake during this period; seepage to ground water was the largest loss of water from the lake (table 5). The largest uncertainty associated with the water budget is in the ground-water

Mirror Lake, New Hampshire.

fluxes and is related to the complexity of the geologic deposits and the associated variability in hydraulic conductivity.

For initial calculations of ground-water fluxes, hydraulic conductivity was determined by single-well aquifer tests, which test only small volumes of the aquifer in close proximity to the wells. The rate of ground-water inflow to the lake was estimated to be 47,000 m^3/yr. A second estimate of ground-water inflow was generated by using data on oxygen isotopes (J.W. LaBaugh, U.S. Geological Survey, oral commun., 2006). The ratio of oxygen-18 to oxygen-16 in water can be used to estimate the rate of ground-water discharge to the lake. The isotopic ratio of lake water must be balanced by the ratios in incoming water and the change in the ratio that takes place as a result of evaporation. The rate of ground-water inflow to the lake can be computed by using measured isotopic

Mirror Lake, New Hampshire

Figure 31. Location map of Mirror Lake, New Hampshire.

Table 5. Initial and final water budgets for Mirror Lake in New Hampshire. Values are in 1,000 cubic meters per year.

	Initial	Final
Inflows		
Precipitation	182	182
Surface-water inflow	417	417
Ground-water inflow	47	113
Outflows		
Evapotranspiration	77	77
Surface-water outflow	257	257
Ground-water outflow	281	347
Lake volume change	16	16
Imbalance	**15**	**15**

Weather station.
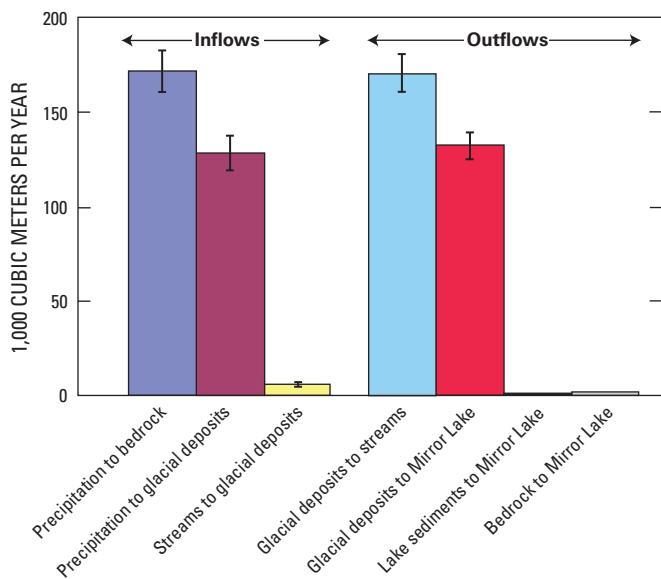
ratios of ground water, precipitation, streamflow, and the lake itself, along with independently measured rates of precipitation, streamflow, and evaporation. This approach yielded an estimate of 95,000 m³/yr of ground-water inflow to Mirror Lake, about twice the value originally calculated.

A third estimate of ground-water inflow was generated from a separate study of the ground-water basin within which Mirror Lake lies (Tiedeman and others, 1997). As part of this study, a computer model of ground-water movement was constructed, producing a water budget for the ground-water basin (fig. 32); therefore, water budgets were available for two distinct accounting units. One accounting unit, Mirror Lake, is actually nested within another accounting unit, the ground-water basin in which the lake resides. A common component in the two water budgets is inflow to the lake. The mutual benefit of these concurrent studies was that data from one

could be used to refine the other. For example, the ground-water-flow model was calibrated by using stream discharge and water levels in wells that were part of the lake study. Conversely, ground-water flow to and from the lake could be calculated from the ground-water-flow model. The model results indicated that 133,000 m³/yr were discharged to Mirror Lake, a value 2.8 times greater than the initial calculation.

The lack of agreement among the three approaches used to estimate ground-water inflow to the lake clearly illustrates some of the uncertainties in water-budget studies. Which estimate of ground-water inflow is correct? Researchers at the site ultimately settled on a rate of 113,000 m³/yr. To balance the final water budget for the lake (table 5), ground-water outflow from the lake was adjusted to compensate for the change in inflow; estimates of precipitation, stream inflow and outflow, and evaporation measurements were all thought to be acceptable.

Results from the Mirror Lake study demonstrate the value of applying multiple approaches to quantify water-budget components. However, it is clear that all water budgets, including the most detailed budgets, contain some degree of uncertainty related to measurement inaccuracies and to our limited ability to make measurements in sufficient spatial and temporal detail. Methods used to calculate the water budget of Mirror Lake were state-of-the-art and the most accurate available. Estimated uncertainties are 5 to 10 percent for precipitation, 10 to 15 percent for evaporation, 5 to 10 percent for streamflow into and out of the lake, and 30 to 50 percent for ground-water inflow and outflow (T.C. Winter, U.S. Geological Survey, and G.E. Likens, Institute of Ecosystem Studies, oral commun., 2006). The overall uncertainty in the water budget is considered to be about 13 percent.

Figure 32. Model calculated water budget for the Mirror Lake ground-water basin from Tiedeman and others (1997). Error bars indicate approximate 95-percent confidence interval.

Surface-water outflow from Mirror Lake was measured with a combined weir and flume.
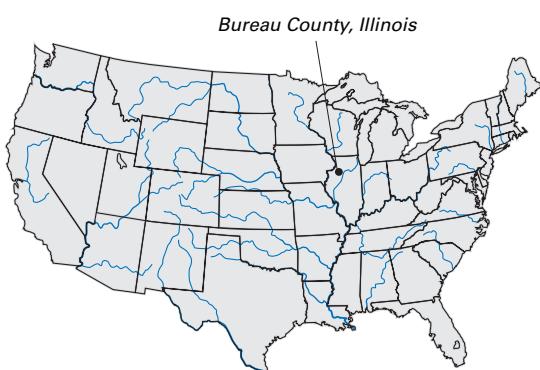
The waste disposal site near Sheffield, Illinois, was in operation from 1967 to 1978.

Water Budget at a Waste Disposal Site in Illinois

A water budget was developed for a radioactive-waste disposal site in Bureau County in northwestern Illinois (fig. 33) for the period of July 1982 through June 1984 (Healy and others, 1989b). The 8-ha site is situated in complexly layered glacial and eolian sediments that range from 10 to 30 m in thickness and overlie a thick sequence (140 m) of low-permeability shales. The disposal site was in operation from 1967 to 1978. During that time, waste was placed in shallow trenches that were subsequently covered with compacted clay covers. The main objective of the study was to estimate rates of water percolation through the trench covers, an issue of concern for regulatory agencies seeking to minimize potential for ground-water contamination. A secondary objective was to assess the utility of water-budget methods for estimating those rates.

Precipitation was measured at three locations (two tipping bucket and one weighing gage). Evapotranspiration rates were estimated on an hourly basis by using a combination Bowen-ratio/aerodynamic profile method. Data collected from the onsite weather station included net radiation, shortwave and longwave radiation, wet- and dry-bulb air temperatures at three heights, soil temperatures and heat flux, and wind-speed and direction (Healy and others, 1989a). Flumes and weirs were used to measure runoff in the ephemeral streams that drained the site. Change in water storage in trench covers was measured with nuclear moisture probes at weekly intervals at three locations. At those same locations, percolation through trench covers (which was considered equivalent to recharge) was estimated with the Darcy method by using pressure heads measured with vertical clusters of tensiometers. Tensiometer readings were obtained electronically with pressure transducers and data recorders at intervals ranging from 5 to 60 minutes.

Figure 33. Location map of Bureau County, Illinois.

Data from onsite weather station were used to estimate evapotranspiration.
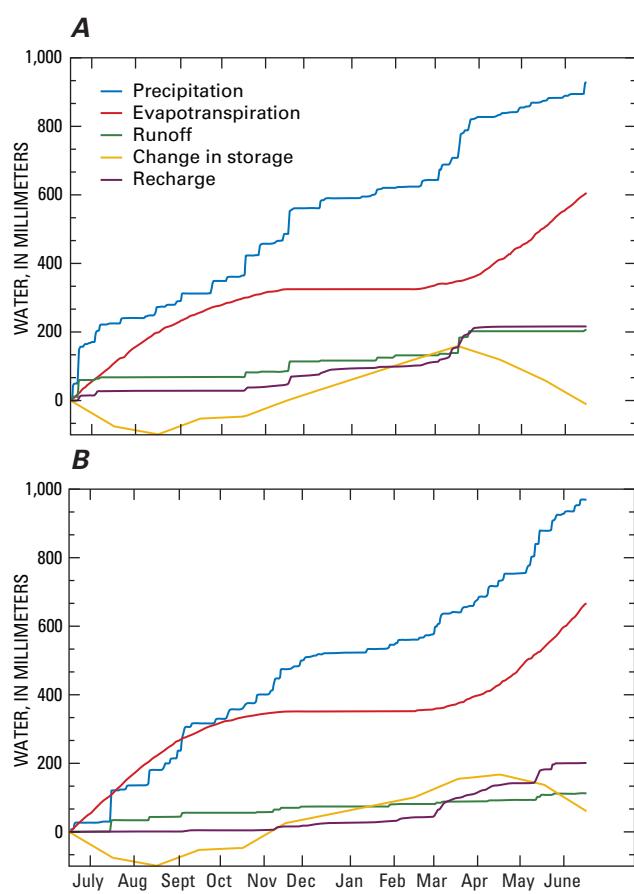
Flumes and weirs were used to measure runoff.

Results of this study illustrate the day-to-day, season-to-season, and year-to-year variability in individual water-budget components. They also demonstrate the inter-dependence among all components. Similar seasonal trends are apparent between the 2 years, but there are some distinct differences (fig. 34). Precipitation totals for the 2 years were similar. The average of 948 mm is close to the long-term average of 890 mm; however, monthly values varied between years. July of the first year was quite wet, but July in the second year saw very little rain. May and June had little rain in the first year but were extremely wet the second year. This variability in precipitation affected other water-budget components, especially runoff and recharge. Runoff was episodic, occurring only in response to large precipitation events, and was more likely to occur if the soil-moisture contents were high. About one-half of the average annual recharge of 208 mm occurred during the months of March and April. Recharge was 54 percent of precipitation for those 2 months in the first year and 55 percent the second year. Other months did not display such consistency: July had recharge of 28 mm in the first year and 0 mm in the second, and May and June had 1 mm in the first year and 64 mm in the second. Evapotranspiration showed consistent seasonal patterns over both years, but daily values during peak summer months were influenced by the availability of soil moisture.

The residual error in the water budget, defined as the difference between precipitation and all other components of the water balance, was –81 mm/yr on average (table 6). Although the magnitude of this number is small relative to precipitation and evapotranspiration, it is large relative to all other water-budget components. Therefore water-budget methods may be suitable for estimating evapotranspiration at this site but problematic for estimating other components of the water-budget equation. Given the variability in weather patterns, the 2-year study period may have been too short to adequately determine a long-term average water budget.

Figure 34.   Water-budget components for (*A*) July 1982 through June 1983 and (*B*) July 1983 through June 1984 for a site in northwestern Illinois (Healy and others, 1989b).

Soil moisture content and pressure head were measured.

Table 6.   Annual values of water-budget components in millimeters for a site in northwestern Illinois (Healy and others, 1989b).

Year	Precipitation	Evapotranspiration	Runoff	Storage change	Percolation into trench	Residual
July 1982 – June 1983	927	606	206	–11	216	–90
July 1983 – June 1984	969	667	113	60	201	–72
2-Year average	948	637	160	24	208	–81

H   Uncertainty in Water-Budget Calculations

All water-budget calculations contain some uncertainty. There are two general sources of this uncertainty: natural variability of the hydrologic cycle and errors associated with measurement techniques. Natural variability occurs in all aspects of the hydrologic cycle. Precipitation patterns are affected by altitude; evapotranspiration and runoff are affected by soil properties, vegetation type and density, surface slope and aspect, depth to ground water, and other factors. Temporal variability in storage and fluxes is largely tied to diurnal, seasonal, and long-term trends in weather. On a daily basis, the pattern of solar radiation generally limits evapotranspiration to daylight hours. Evapotranspiration also is affected by seasonal trends in solar radiation; rates are low during winter months when solar radiation is low, and rates are high during summer months. Seasonal patterns in precipitation exist in many regions. Perhaps the most extreme example is in South Asia. In Mumbai, India, storms during the monsoon season of June

An instrumented field in north-central Oklahoma.

U.S. Department of Energy Atmospheric Radiation Measurement Program

through September account for 94 percent of the average annual precipitation of 180 cm (BBC Weather, *http://www.bbc.co.uk/weather,* accessed on February 12, 2007). Long-term climate change has a large effect on the hydrologic cycle. Ground water from the middle Rio Grande aquifer is as old as 30,000 years (Plummer and others, 2004), indicating that some recharge to the aquifer occurred when the Southwestern United States was experiencing a much wetter climate.

The water-budget equation commonly is used to estimate rates of evapotranspiration or ground-water recharge. A simple analysis of this approach illustrates the importance of considering measurement errors. In this approach, all but one of the water-budget components are measured or estimated independently. The remaining component is assumed equal to the residual of the equation. Consider, for example, an arid region with coarse-grained soils. Typically, there is no surface runoff; precipitation infiltrates the subsurface and is either removed by evapotranspiration or percolates through the unsaturated zone to recharge the underlying aquifer. An appropriate water-budget equation would be:

$$Precipitation = Evapotranspiration + Recharge \qquad\qquad \text{(H1)}$$

Multiple gages reduce the inaccuracy of precipitation estimates.

Suppose an estimate of evapotranspiration is needed. For one year, precipitation was measured at 25 cm and recharge was measured at 3 cm. An evapotranspiration estimate of 22 cm is then derived. If the recharge estimate were in error by 10 percent (recharge was actually 3.3 cm), the uncertainty in the evapotranspiration estimate would be small, less than 2 percent. Even if the recharge estimate were in error by 100 percent (recharge was actually 6 cm and evapotranspiration was 19 cm), the evapotranspiration uncertainty would be less than 15 percent. Now suppose, instead, that we are interested in estimating recharge as the difference between precipitation and evapotranspiration. If evapotranspiration were independently measured at 24 cm, a recharge estimate of 1 cm would be derived. If measurement uncertainty was 10 percent for evapotranspiration, recharge for that year may have been as high as 3.4 cm, a 240-percent difference from the original estimate. As can be seen from this example, if the magnitude of the water-budget residual is small compared to those of the other components, then small uncertainties in other components can result in very large uncertainties in the residual.

Humans and the Hydrologic Cycle

By our very existence, humans, along with all other animals and plants, are a part of the Earth's hydrologic cycle. Therefore, any human activity affects the natural hydrologic cycle. Drinking a glass of water, taking a bath, washing the car—these activities involve a small amount of water, yet they alter the course of that water within its cycle, though perhaps only in minor ways. The activities of humans that affect the hydrologic cycle can be grouped into three overlapping categories: construction of water storage and conveyance structures, land use, and extraction of ground water. These activities are reflected in the hydrologic cycle as a redistribution of water within the atmosphere, land surface, and subsurface, in changes in rates of water flow within and among these compartments, and in a relocation of points of water inflow to and outflow from them. Alterations to the hydrologic cycle may, in turn, lead to changes in natural environments, such as creation of new habitat for fish or loss of wetlands. The following three sections provide a brief overview of the effects humans can have on the hydrologic cycle.

Canal and irrigated fields.

Bureau of Reclamation

Water Storage and Conveyance Structures

Surface-water reservoirs serve many beneficial purposes, providing water for irrigation, domestic use, navigation, hydroelectric power, and recreation. Dams and the reservoirs they create alter the natural movement of water in streams, reducing the number of extreme events, such as floods, and possibly changing stream temperatures (Collier and others, 1996). These alterations may affect downstream ecosystems, benefiting some species but stressing others. Reservoirs create whole new ecosystems, often providing valuable fish habitat. They generally lead to an increase in evapotranspiration and the flow of surface water to the subsurface and to a decrease in total streamflow relative to natural conditions.

Reservoirs, along with an infrastructure of pipelines, canals, and ditches, facilitate the transport of water between watersheds. For some basins, this artificial export or import

Glen Canyon Dam and Lake Powell.

Bureau of Reclamation

may be the single largest component of its water budget. In Colorado approximately 475,000 acre-ft of water from the Colorado River basin is transferred eastward across the Continental Divide each year for agricultural and domestic use (*http://www.water.denver.co.gov/*). This water would have originally flowed to the Pacific Ocean; it is now diverted to the Gulf of Mexico. Canals and ditches and other conduits for transporting water provide the opportunity for exchange of water with the subsurface and the atmosphere.

Other conveyance structures are designed to remove rather than supply water. The Corn Belt of the United States, running through Indiana, Illinois, and Iowa, boasts some of the most productive agricultural land in the world. Yet less than 200 years ago, much of this fertile farmland was natural wetland. Since the early 1800s, farmers have installed tiles and dug ditches to facilitate drainage of the wetlands. Tile drains effectively lower the water table to about 1 meter below land surface, thus providing an adequate environment for crops to grow.

Tile drains are used to lower the water table in some agricultural fields.

Urban development alters water's course through the hydrologic cycle.

Land Use

Land use is perhaps the most important phenomenon affecting water exchange between land surface and the atmosphere. Conversion of native forests, grasslands, and wetlands to agricultural uses constitutes the largest land-use change (in terms of area, at least) in the United States and most other countries. Replacement of native vegetation with agricultural crops leads to changes in patterns of infiltration, evapotranspiration, and ground-water recharge. Irrigation of crops in arid regions has produced an inflow of water to the atmosphere through evapotranspiration that was absent under natural conditions. Clearcutting of rain forests has reduced evapotranspiration rates in large areas of the Amazon River basin. Because of the importance of this region in global circulation patterns, potential climatic effects could extend beyond South America.

Urbanization accounts for the second largest change in land use within the United States. Urban features such as buildings, roads, and parking lots are all impermeable. Thus, they tend to enhance surface runoff of precipitation and reduce infiltration. Runoff from these features may be channeled through storm sewers to streams, leading to increased streamflow and flooding in the worst situations. That runoff also could be funneled to an infiltration gallery leading to an increase in ground-water recharge. Water in urban areas is conveyed to and from users in networks of underground pipes. Invariably, there is some leakage from these pipes. That leakage can be a substantial source of ground-water recharge.
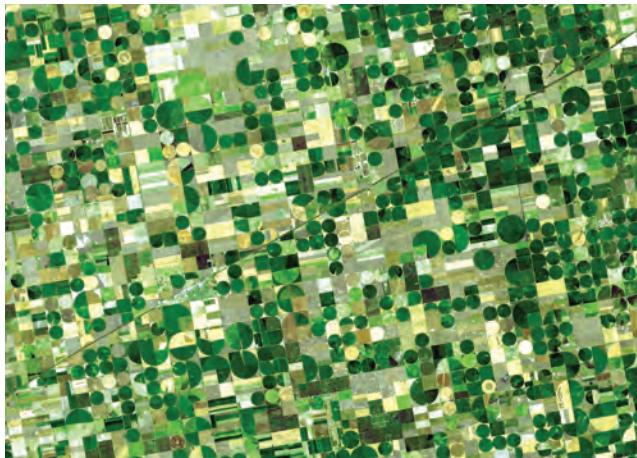
Ground water being pumped for irrigation of rice field.

Satellite image of irrigated fields in western Kansas.

Trickle irrigation conserves water.

Ground-Water Extraction

Throughout history, humans relied primarily upon surface water to satisfy their needs for water. Storage reservoirs were constructed, streams were diverted, and canals were built to convey the water to the areas of need, usually agricultural fields or urban areas. Over the past 200 years, humans have become more reliant on ground water to supply their needs. Extraction of ground water, whether for domestic, agricultural, or industrial uses, is balanced by a reduction in ground-water storage, a reduction in natural discharge, or an increase in recharge. For any particular aquifer, all of these phenomena can occur simultaneously, but change in storage (indicated by changing ground-water levels) is usually more easily determined than changes in discharge or recharge.

Many aquifers within the United States have experienced widespread declines in ground-water levels over the last several decades. Declining water levels indicate a reduction in subsurface water storage, and they may result in reduced ground-water flow to wetlands and streams. Streams that normally gain water from the subsurface could be transformed into losing streams. Effects such as these can sometimes be seen instantaneously—for example, a stream drying up when a well pump is turned on. More commonly, the effects are prolonged in time and difficult to quantify. Similarly, the effects of reduced ground-water discharge on stream and wetland ecosystems may become apparent only over extended periods of time.

Natural Resources Conservation Service

Development of a new irrigation well in west-centrl Florida triggered hundreds of sinkholes over a 20-acre area. (See person in center for scale.)

Most ground water that is extracted for irrigation is evapotranspired back to the atmosphere shortly after it is applied to the land surface. However, a percentage of irrigation water, called irrigation excess or return flow, may run off to a stream or may infiltrate, percolate through the unsaturated zone, and eventually become ground water again. Most ground water removed from the subsurface for domestic use eventually ends up returning to the saturated zone through septic leach fields or is discharged to surface-water bodies from wastewater-treatment plants.
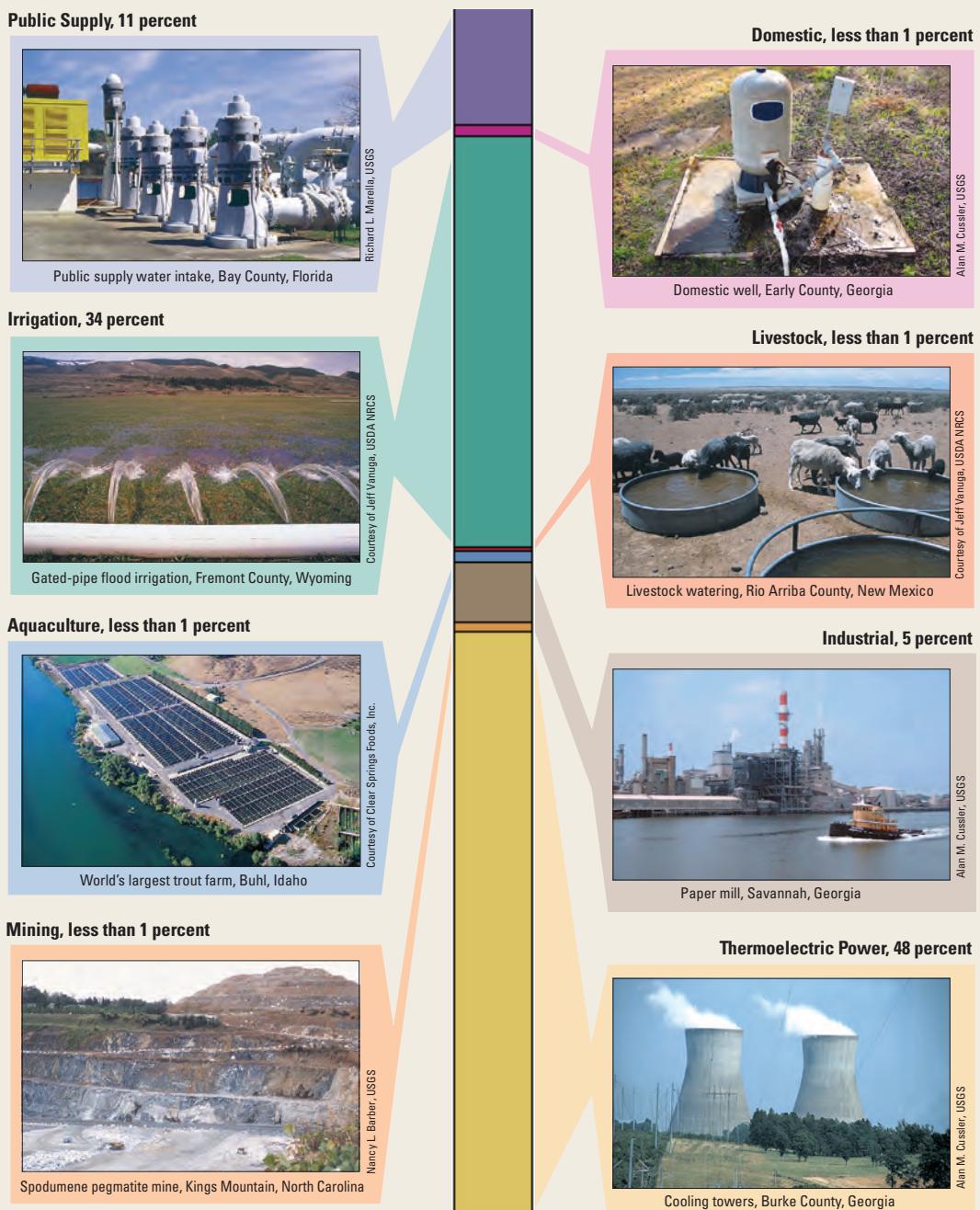
A 1942 photograph of a reach of the Santa Cruz River south of Tucson, Arizona, shows stands of mesquite and cottonwood trees along the river (left photograph, Arizona Game and Fish Department). A replicate photograph of the same site in 1989 shows that riparian vegetation has largely disappeared (right photograph, Robert H. Webb). Data from two nearby wells indicate that the water table has declined more than 30 meters due to pumping; this pumping appears to be the principal reason for the loss of vegetation.
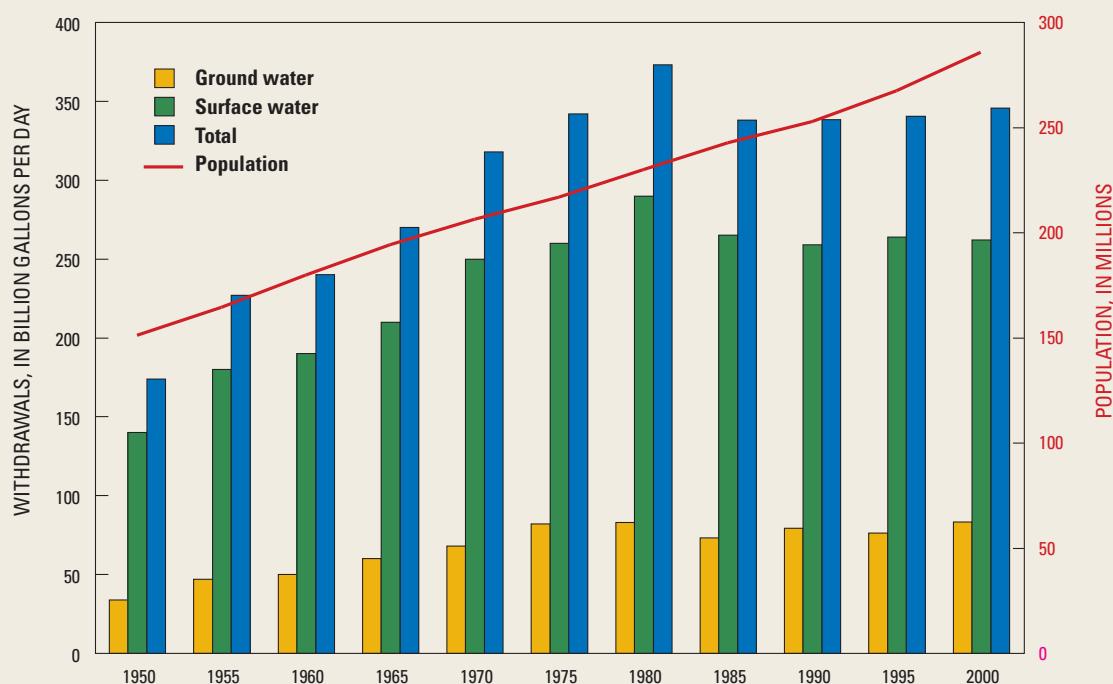
I Water Use and Availability

Humans need water. But just how much water do we need? Every day in the United States 345 billion gallons on average is withdrawn from ground- and surface-water sources for human use (Hutson and others, 2004). This is equivalent to more than 1,000 gal/d for every person in the country—about 40 bathtubs full. We do not usually take that many baths, so how is this water used? The largest use (48 percent) is by thermoelectric power plants, for cooling and steam generation. Other uses, as shown in figure *I–1*, are for irrigation of agricultural lands, domestic needs, industry, mining, aquaculture, and livestock. Water satisfies a myriad of thirsts; a daily bath is but a few drops in the water-use bucket.

Public Supply, 11 percent

Public supply water intake, Bay County, Florida

Richard L. Marella, USGS

Irrigation, 34 percent

Gated-pipe flood irrigation, Fremont County, Wyoming

Courtesy of Jeff Vanuga, USDA NRCS

Aquaculture, less than 1 percent

World's largest trout farm, Buhl, Idaho

Courtesy of Clear Springs Foods, Inc.

Mining, less than 1 percent

Spodumene pegmatite mine, Kings Mountain, North Carolina

Nancy L. Barber, USGS

Domestic, less than 1 percent

Domestic well, Early County, Georgia

Alan M. Cussler, USGS

Livestock, less than 1 percent

Livestock watering, Rio Arriba County, New Mexico

Courtesy of Jeff Vanuga, USDA NRCS

Industrial, 5 percent

Paper mill, Savannah, Georgia

Alan M. Cussler, USGS

Thermoelectric Power, 48 percent

Cooling towers, Burke County, Georgia

Alan M. Cussler, USGS

Figure I-1.   *Percent of total water withdrawals for major categories within the United States (from Hutson and others, 2004).*

Box I **Water Use and Availability**   **59**

Total water withdrawals in the United States have been stable since the mid-1980s. However, on a per capita basis, total withdrawals have decreased over the same period (fig. I–2). This is likely the result of improved techniques that require less water for power generation and advances in irrigation efficiency. Technologies such as low-flow bathroom fixtures and water-saving appliances in homes aid in conserving local water supplies. In New York, for example, water use has declined from about 200 gal/d per person in 1990 to less than 140 gal/d in 2003 (City of New York, Department of Environmental Protection, *http://www.nyc.gov/html/dep/html/droughthist.html* accessed on December 18, 2006). On the national scale, however, the savings realized in the home are minimal when compared to thermoelectric and agricultural use. Conservation and improved efficiency of water use may be driven by economic and water-quality issues as well as by water supply. Energy costs for pumping water continue to rise, and stricter water-quality standards for water discharges have been put in place at the Federal and State levels. Periods of drought may prompt water managers to impose limits on water use. Farmers, ranchers, manufacturers, energy providers, and individual consumers adapt their water use to technological advances, changing regulations, and market forces.

Figure I-2.   *Freshwater withdrawal and population in the United States, 1950–2000 (from Hutson and others, 2004).*

"Compilation of water-use information on a regular basis provides information on the amount of water used by humans, yet there has been no corresponding assessment of water availability within the United States. In addition, the water needs of biota that inhabit waterways, wetlands, flood plains, and other environments are largely unknown" (National Science and Technology Council, 2004). Water budgets of watersheds, aquifers, and surface-water bodies are essential tools for assessing the availability of water for both human and environmental needs. It is useful, therefore, to look at the way water budgets are determined, the uncertainty inherent in those budgets, and the effect of human activity on water budgets. Episodes of drought and floods are reminders that water availability can change substantially over time. It is therefore important to consider the temporal variability in the movement and storage of water in the hydrologic cycle and the relation of that variability to water use.

J   Water Budgets of Political Units
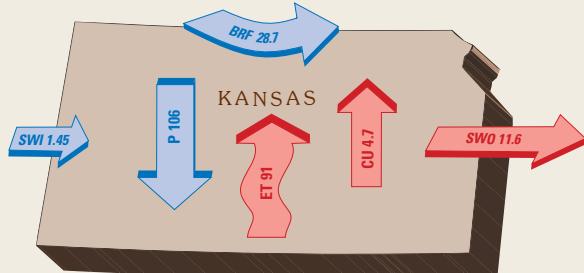
Water in most areas is managed by governmental units, be it countries, States, counties, or water districts. Concerns of these entities include how much water they have, how much water they use, and how, when, and at what rate water supplies are replenished. Throughout this report, the uncertainty inherent in water-budget calculations is demonstrated. That uncertainty is compounded when boundaries of an accounting unit are not aligned with hydrologic boundaries. Such is often the case with governmental units where humans have defined political boundaries that may crisscross natural watershed boundaries and partition aquifers. These boundaries may also follow rivers, resulting in the rivers being shared by competing entities. Measurement of surface- and ground-water flow across political boundaries presents unique and sometimes contentious challenges.

The average annual water budget for the State of Kansas is depicted in figure *J–1* for a period from the 1970s into the 1980s. Average annual precipitation within the State is about 27 inches. Equating inflow and outflow, with the assumption that annual change in storage is relatively small, the water budget can be expressed as:
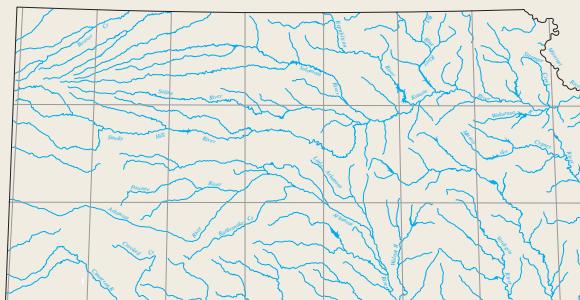
Precipitation + Surface-Water Inflow =
Evapotranspiration + Surface-Water Outflow   (J1)

State boundaries cross tens, if not hundreds, of streams that flow from Colorado and Nebraska into Kansas and a similar number of streams that flow out of Kansas to Oklahoma and Missouri (fig. *J–2*). Only a small percentage of those streams have stream gages to monitor flow, and few of those gages are located at State lines. Thus, uncertainty in estimating total surface inflows and outflows for the State of Kansas may be quite high.

Conspicuous by their absence in the above equation and in figure *J–1* are ground-water inflow and outflow. These processes do occur. At the State level, those rates were deemed insignificant relative to other terms. At a local scale, however, ground-water flow into Kansas from Colorado in the High Plains aquifer (McGuire and others, 2003), for example, may be an important component in the water budget of some western Kansas counties.

Figure J–1.   *Average-annual water budget for Kansas, in million gallons per day, 1970s and early 1980s. Abbreviations: BRF, boundary-river flow; CU, consumptive use (evapotranspiration related to human activities, mostly irrigation); ET, evapotranspiration from native plants and nonirrigated agricultural fields; P, precipitation; SWI, surface-water inflow; SWO, surface-water outflow (Carr and others, 1990).*

Figure J–2.   *Principal rivers, Kansas (Paulson and others, 1991).*

The Missouri River flows in or along seven States.

The High Plains aquifer lies in parts of eight States.

Water Budgets and Management of Hydrologic Systems

The global hydrologic cycle is a continuum; the atmosphere, land surface, and subsurface are intrinsically linked by the water cycling through them. Each of these compartments is influenced in one way or another by human activities, either directly or as a complementary reaction to changes in another compartment. The following examples are presented to show why water budgets are important and how they can be used in management of water resources.

Large River System: Colorado River Basin

The Colorado River basin covers 637,000 km^2 in the Southwestern United States (fig. 35). Although its headwaters are in the Rocky Mountains, where snowmelt and rainfall provide substantial quantities of water, the river traverses largely semiarid and arid regions, where relatively small quantities of water are added to the river. Irrigated agriculture and major metropolitan areas in southern California, southern Nevada, southern Arizona, and the Front Range in Colorado make major demands on Colorado River water. In California and Colorado, water from the Colorado River is exported from

Dillon Reservoir, Colorado.
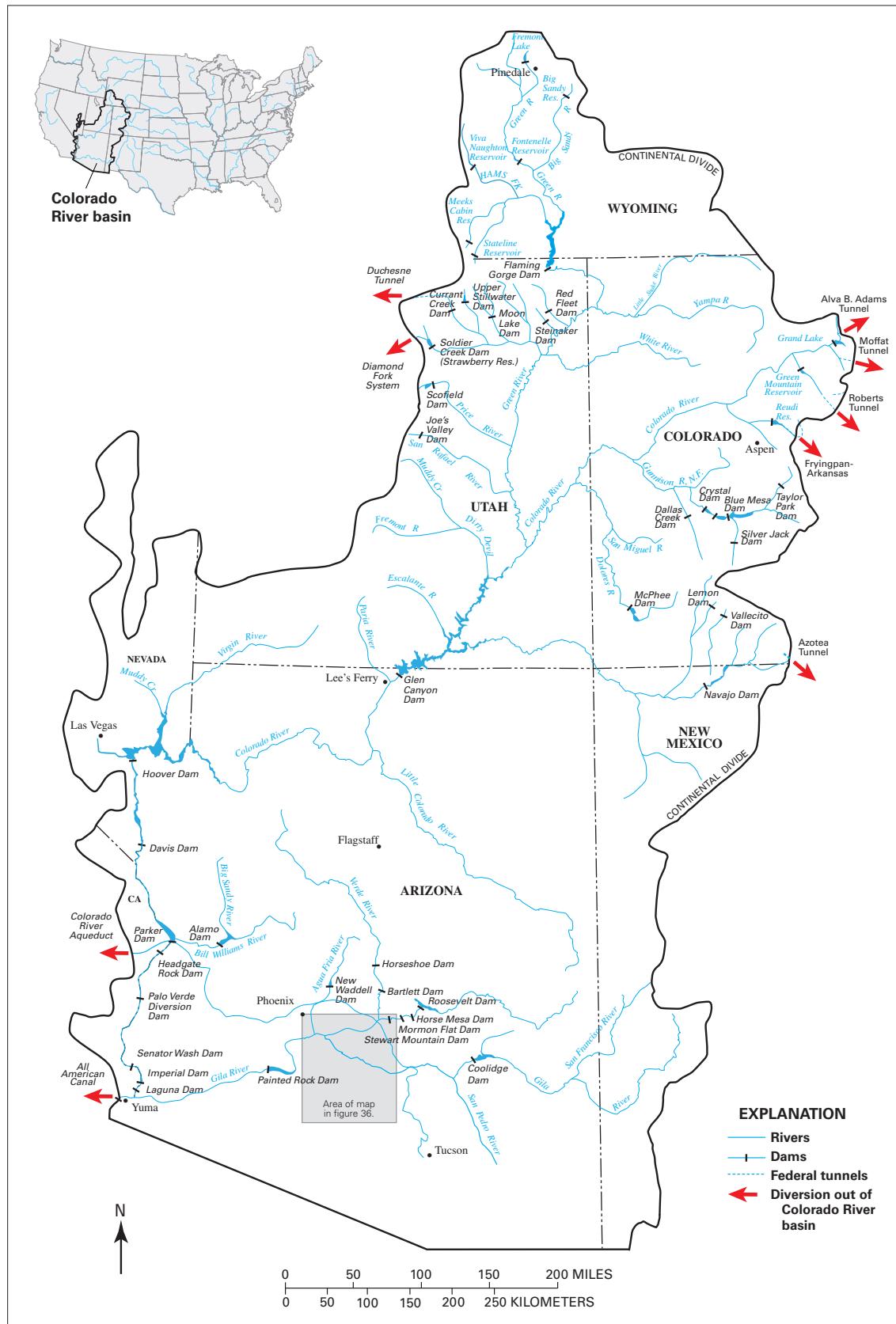
Figure 35.   Map of the Colorado River basin showing locations of major surface-water reservoirs and diversions.

the basin. Because water demands can exceed what the river supplies, technically complex and politically contentious water compacts have evolved over many decades.

Water budgets have been an essential tool for apportioning the Colorado River water among various claimants. Budgets are needed for various types of accounting units that are nested within the basin, including headwater watersheds, agricultural fields, reservoirs, and aquifers. The water budgets for all of these nested units ultimately interact to balance the overall water budget of the basin. The challenge is to manage water resources while explicitly accounting for the inherent uncertainties in water-budget estimates. Two accounting units of the Colorado River basin are briefly discussed in the following sections.

Streamflow in the Colorado River is derived primarily from snowfall in the mountains within its drainage.

Watersheds and Reservoir Management

Reservoirs are key components in managing the water resources of the Colorado River basin. Reservoirs are used to store water when and where it is abundant so the water can be released to downstream areas or exported out of the basin as needed. Reservoirs serve this purpose in many different environments but are perhaps most prominent in semiarid and arid regions. Over the course of each year, operators of reservoirs in the Colorado River basin must make decisions on release rates from reservoirs. Determining how much water to release involves some risk—release too much water and there may not be enough later in the year to satisfy thirsty customers; release too little water and large spring runoff from snowmelt could cause severe flooding.

Selection of optimum water-release rates from a reservoir requires an understanding of the reservoir's water budget. Rates and timing of releases depend on the amount of water currently stored in the reservoir, the anticipated rate and timing of inflow, needs for power generation or flood control, irrigation demands, and other legal and environmental requirements. Watershed models provide a convenient means for predicting rates and timing of inflow; thus, they have become important decision-support tools for many reservoir operators. Watershed models basically are water budgets of watersheds; they calculate water input, storage, and losses of water from a watershed. But like all water budgets, there are uncertainties in watershed model results.

Lake Mead, Arizona and Nevada.

The Bureau of Reclamation (BOR) and the U.S. Geological Survey (USGS) have developed a decision-support system to allow BOR to manage some reservoirs in the Upper Colorado River basin in Colorado (Frevert and others, 2006). The Bureau of Reclamation runs watershed models, developed by the USGS, to predict discharge from the major streams that flow into these reservoirs. A novel approach allows uncertainty in predicted reservoir inflows to be factored into reservoir operations. The watershed model uses current data on snowpack, soil moisture, stream discharge, and historical climate data, such as temperature and precipitation to generate synthetic discharge hydrographs, one for each historical year of data. The idea behind the approach is to estimate what discharge rates are likely to occur in the future under climate conditions similar to those of the past. Results from the watershed model are used to generate probabilities associated with different discharges (for example, 50 percent of the time the volume of discharge would exceed 500,000 m³; 10 percent of the time peak discharge would exceed 1,000,000 m³). The reservoir operator factors these discharges and probabilities, along with estimates and uncertainties of other components of the reservoir's water budget, into decisions on reservoir releases.
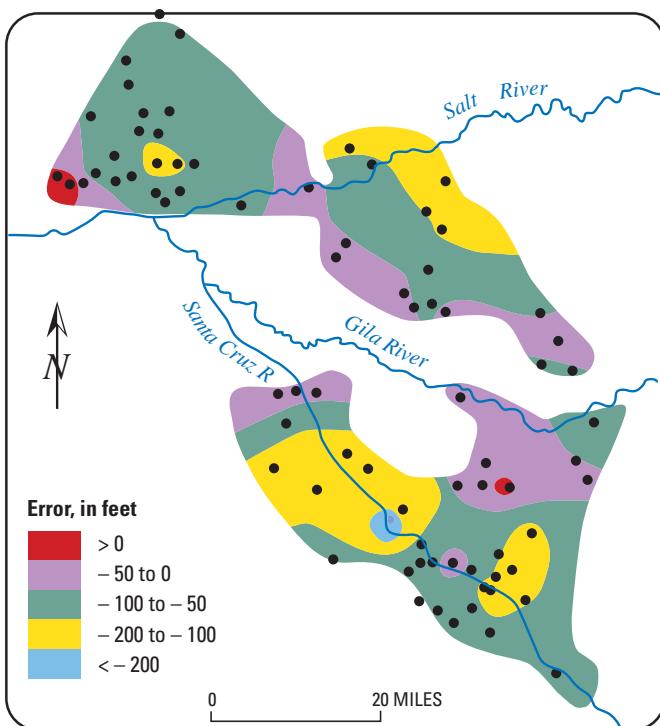
Central Arizona Project aqueduct.

Aquifers in Arizona

The limited amount of surface water in Arizona has led to substantial use of ground water, especially for agriculture. With an arid to semiarid climate and very low rates of recharge, ground-water withdrawals caused water levels in Arizona to decline as early as the 1920s. Declines were more rapid after the 1940s because of the increased availability in rural areas of electricity to power deep-well turbine pumps. Ground-water withdrawals have resulted in reduced discharge to streams and wetlands (Webb and others, 2007) and water-level declines that have caused land subsidence in some areas (Galloway and others, 1999). To help manage ground-water resources, ground-water flow models of aquifers were constructed for many parts of the State. These models, which in effect are water-budget models, were used to predict how ground-water levels would be affected by future aquifer-management practices.

To assess the predictive capabilities of a ground-water flow model of an aquifer in central Arizona (fig. 35), Konikow (1986) compared water levels in 77 wells measured in 1974 with water levels predicted for that year with a model of the Salt River and lower Santa Cruz River basins (Anderson, 1968). The original model was calibrated by using water-level and pumping data collected from 1923 to 1964. Between 1923 and 1964, average ground-water levels declined by an average of about 120 ft. Water levels measured in 1974, 10 years after the ground-water model was completed, differed from those predicted by 50 to 200 ft in large parts of the area (fig. 36).

Error, in feet
- > 0
- -50 to 0
- -100 to -50
- -200 to -100
- < -200

0   20 MILES

Figure 36.   Map of central Arizona showing error in water levels predicted by a ground-water flow model (after Konikow, 1986). Errors are determined as predicted minus measured water levels. Location shown in figure 35.

Agricultural crops require irrigation in Arizona.

Natural Resources Conservation Service

Land surface subsidence in Eloy area, central Arizona.

The discrepancies were attributed to three main factors: (1) differences between where ground-water development was predicted to occur and where it actually occurred, (2) differences in the amount of water predicted to be withdrawn and the amount actually withdrawn, and (3) changes in the hydraulic properties of the aquifer and confining beds caused by dewatering, including land subsidence in a few areas. The Arizona experience illustrates the value of continually updating ground-water-budget models because of the inherent uncertainties in aquifer characterization and in future patterns of ground-water development.

The State of Arizona enacted the Arizona Groundwater Management Code (*http://www.azwater.gov/dwr/Content/Publications/files/gwmgtovw.pdf*, accessed on March 13, 2007) in 1980 to address the sustainability of its limited ground-water resources. The code established a program of ground-water rights and permits with some very demanding provisions. A series of water-management plans created comprehensive conservation targets for areas of severe ground-water depletion. Developers were required to obtain a 100-yr water supply for any new growth. A program for reporting ground-water withdrawal and use was established. The restrictive nature of these provisions reflects serious concern for the availability of water within Arizona. Effective implementation of the code requires development and continual revision of water budgets of the State's aquifers.
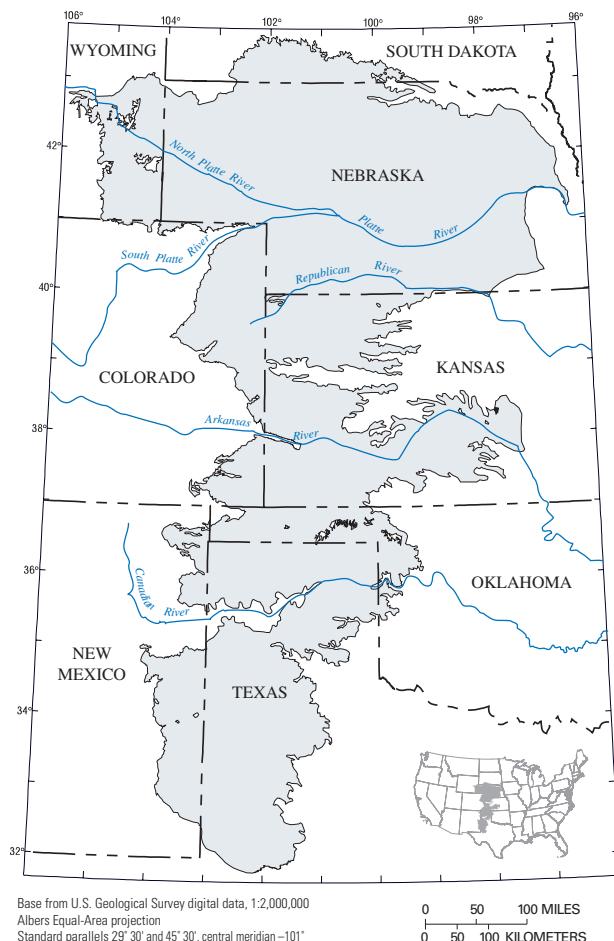
Water well.

Large Aquifer System: High Plains Aquifer

The High Plains aquifer stretches over an area of about 443,000 km^2 in the central plains of the United States (fig. 37). It consists of unconsolidated and consolidated materials, mixtures of sand, silt, and clay, that overlie a sedimentary bedrock surface. The large areal extent and thickness of the aquifer provide storage for huge volumes of water; estimates are as high as 3 billion acre-ft (McGuire and others, 2003). Since the 1940s, large volumes of ground water have been withdrawn to irrigate crops such as corn, soybeans, and cotton. These withdrawals have resulted in substantial declines in ground-water levels beneath large areas, which have in turn led to concerns about ground-water depletion and the future of the aquifer.
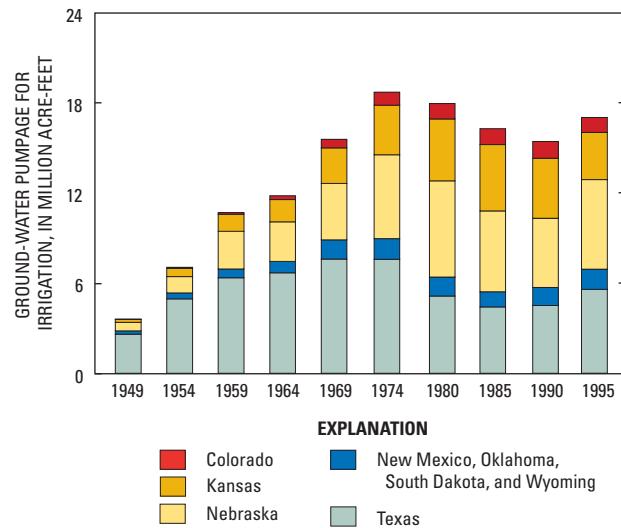
The climate in the High Plains varies from arid to semiarid to subhumid. Average annual precipitation rates range from less than 14 inches in the southwest to about 32 inches in the northeast. There are few perennial streams in the southern and central parts of the High Plains aquifer. Perennial streams, however, are common across the northern part.

Water in the High Plains aquifer has accumulated over thousands of years. Before human development, inflow to the aquifer was some small portion of precipitation that infiltrated, traversed the unsaturated zone, and recharged the aquifer. Estimated recharge rates at that time ranged from 0.10 to 0.50 inch/yr for the southern and central parts and 1 to 2 inches/yr for the northern part of the aquifer. Outflow prior to development was in the form of discharge to streams (or to plants in riparian zones) and hillslope seepage along lateral and eastern aquifer boundaries where the earth materials of the High Plains aquifer form an easily discernible topographic escarpment. While hillslope seepage is a common hydrologic phenomenon, the large spatial scale of this occurrence in the High Plains is noteworthy.
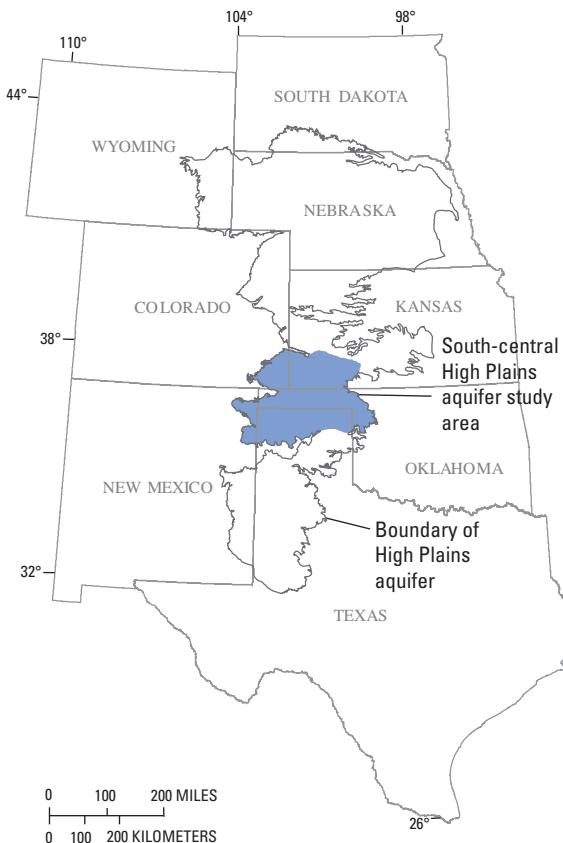
Human development of the High Plains aquifer resulted in large withdrawals of ground water from wells. Total withdrawals and the rate of decline of water levels increased into the mid-1970s (figs. 38, 39) at which time the limitations of this resource became apparent. Since then, improved irrigation efficiency (furrow irrigation has largely been replaced by sprinkler systems) has slowed the rate of decline of ground-water levels. A water budget for a part of the High Plains aquifer was constructed for a time prior to development and another for 1997 (fig. 40; Luckey and Becker, 1999). Over that time, there was an 18-percent reduction in water storage caused primarily by extraction of ground water. There was also a decrease in natural discharge and an increase in recharge from irrigation return flow.

Figure 37. Location map of High Plains aquifer.

Figure 38. Ground-water pumpage from the High Plains aquifer for irrigation by State for selected years, 1949 to 1995 (McGuire and others, 2003).

Figure 39. Water levels over time (1950 – 2005) in observation wells in the High Plains and map showing water-level changes in the High Plains aquifer, predevelopment to 2000 (McGuire and others, 2003).
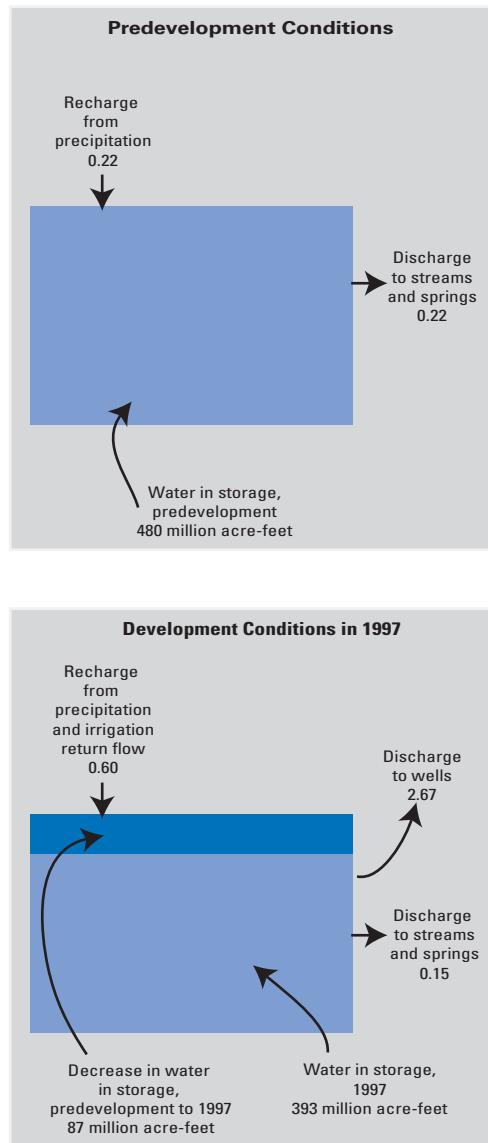
Figure 40. Ground-water budget in the south-central High Plains aquifer study area during predevelopment and in 1997. Values without units are in million acre-feet per year (Luckey and Becker, 1999).

"Accurate estimates of ground-water withdrawals are difficult to obtain. Farmers are not always required to report ground-water withdrawals from their irrigation wells to regulatory agencies, and it is not practical to place flow meters on every well. Indirect methods are often used to estimate withdrawal rates. One such method is based on typical water use for specific crops; another method relates power consumption to pumping rates."

(McGuire and others, 2003)

The loss of water from the entire High Plains aquifer from predevelopment to the year 2000 is estimated to be about 200 million acre-ft, or about 6 percent of its original volume (table 7). Most of this loss occurred in Texas. Ground-water levels have declined over most of the High Plains, but some northern areas have actually experienced a rise in water levels in recent years (fig. 39). These rises are attributed to the use of surface-water supplies for irrigation; excess irrigation water percolates through the unsaturated zone, enhancing recharge rates and augmenting aquifer storage.
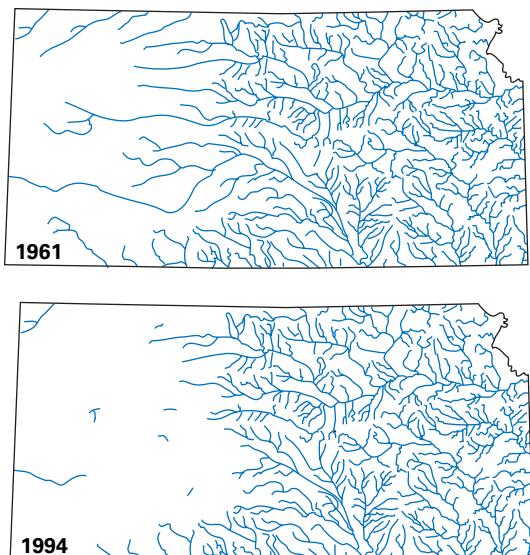
Table 7. Change in water in storage in the High Plains aquifer, predevelopment to 2000 (McGuire and others, 2003).

State	Change in water in storage, in million acre-feet
Colorado	−11
Kansas	−47
Nebraska	4
New Mexico	−8
Oklahoma	−11
South Dakota	0
Texas	−124
Wyoming	0
Eight States	−197

Decades of large withdrawals of ground water have altered historical ground-water discharge patterns. Ground-water discharge to streams has decreased in some areas (fig. 41; Sophocleous, 2000), and hillslope seepage along lateral and eastern aquifer boundaries has been reduced. Effectively, the volume of pumped water, which would have eventually discharged to streams and hillslopes, has been intercepted, and a large percentage of the withdrawn water is lost to evapotranspiration by crops.

Determining the changes in the volume of water stored in an aquifer system such as the High Plains is a difficult task because of its large size, complex geometry, and variable storage properties. Water-level changes shown in figure 39 are based on water-level measurements made in more than 8,000 wells—a large number of wells in an absolute sense, but not so large when the immense expanse of the aquifer is considered. Because pumping for irrigation is seasonal (during the growing season of March through September), these wells were measured in winter or early spring and thus represent maximum water levels for the year by allowing time for water levels to stabilize from the previous irrigation period. Regular synoptic surveys of ground-water levels provide the most useful data with which the health of the aquifer can be assessed. These measurements document changes in storage and provide essential information for calibrating ground-water-flow models.

1961

1994

Figure 41. Major perennial streams in Kansas, 1961 and 1994. Parts of the High Plains aquifer occur in western Kansas where stream channels exhibiting perennial flow have been reduced in length (after Sophocleous, 2000).

High-capacity irrigation well.

Aerial view of center pivot irrigation systems in the Texas Panhandle.

Minimizing reduction in ground-water storage requires a balance between aquifer recharge and discharge. Water budgets are useful tools for assessing this balance. Thus, improved water-budget information can facilitate decisions on water-allocation issues. The accuracy of water budgets can be improved with monitoring of ground-water levels, measurement of streamflow, reliable information on pumping and irrigation rates, and better estimates of rates of natural recharge and of irrigation return flow.

Unfortunately, water-resources management is more complicated than simply balancing the water budget (as if that balancing act were a simple task!). Political and economic issues also are important. The High Plains aquifer system lies within the borders of eight States. Although the States have a common goal of managing aquifer development to prevent depletion and ensure water availability in future years, each State has its own approach for attaining that goal. Irrigation is the economic lifeline of farmers within the High Plains, supporting a vibrant economy since the middle of the 20th century. Cities and towns have grown in response to this economy. Decisions on water management have direct effects

Installing an observation well.

on the lives and livelihoods of the people who reside in this region. Proper management of water resources at any level depends not only on sound science but also on cooperation among political entities, long-term water-management plans, and public involvement and education (Sophocleous, 2000).

A corn field of one acre can give off 4,000 gallons of water per day in evaporation.
One inch of rain or irrigation on that field produces more than 27,000 gallons of water.

Water Budgets and Governmental Units: Lake Seminole
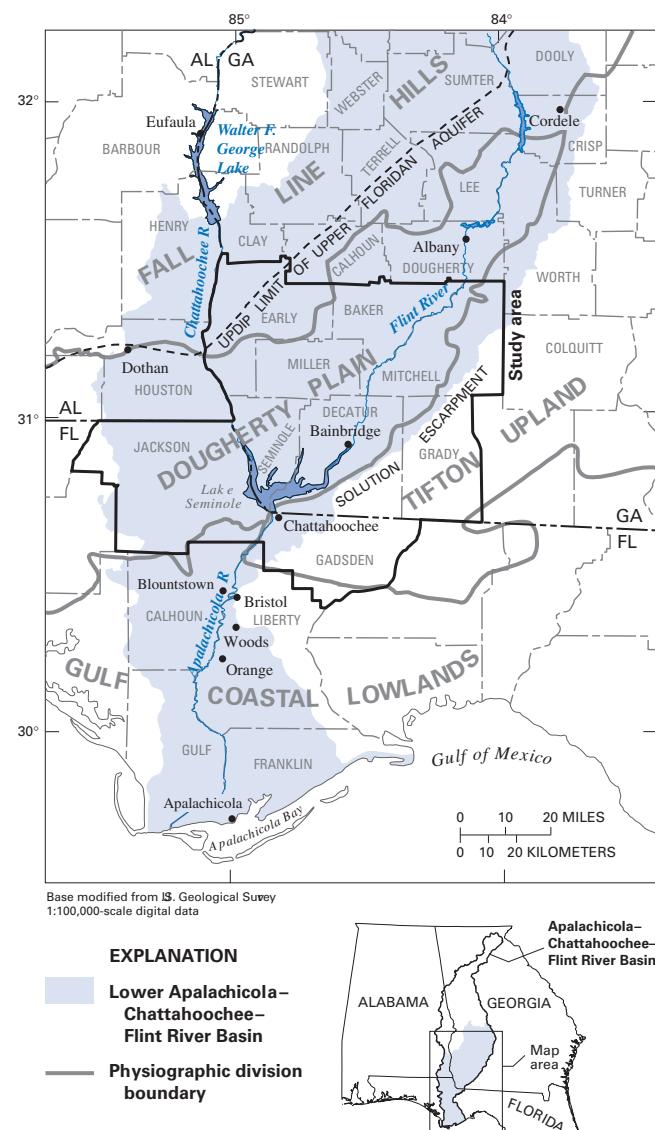
Dams are constructed on rivers to meet a variety of needs. The reservoirs that result from impoundment of flowing waters provide water supply, recreational opportunities, flood control, and hydroelectric power. They also serve as a source of water to maintain navigation levels in waterways. Dams alter the natural flow of water and sediment in a river, thus affecting habitat for biota that live in and along river corridors. Balancing the water needs of humans with those of natural biological resources is an emerging area of concern of reservoir operation.

The Jim Woodruff Lock and Dam on the Apalachicola River in the southeastern United States, built in the late 1940s and early 1950s, impounds Lake Seminole (fig. 42). The lake was intended to aid river navigation, produce hydroelectric power, and provide recreational opportunities. It is fed primarily by flow from the Chattahoochee and Flint Rivers; the contributing watersheds include areas in Alabama, Florida, and Georgia. Outflow from the lake is the major source of freshwater, nutrients, and detritus to the lower Apalachicola River, its estuary, and Apalachicola Bay, an important shellfish fishery.

Water-resource managers in these three States must deal with competing demands for Lake Seminole water. Metropolitan Atlanta draws most of its water from the Apalachicola-Chattahoochee-Flint basin. The rivers in the basin, as well as Lake Seminole itself, are hydraulically connected to the Upper Floridan aquifer system. Irrigation for agriculture in southwestern Georgia is largely derived from ground-water withdrawals. These withdrawals may affect streams flowing into Florida. Alteration of streamflow into Apalachicola Bay is a concern for the fishery in the bay. Allocation of basin water among the three States was of such concern that a compact regarding the water resources of the basin between the U.S. Congress and the States of Alabama, Florida, and Georgia was in effect from 1997 to 2003. The compact has expired and water allocation within the basin may ultimately be decided by the U.S. Supreme Court.
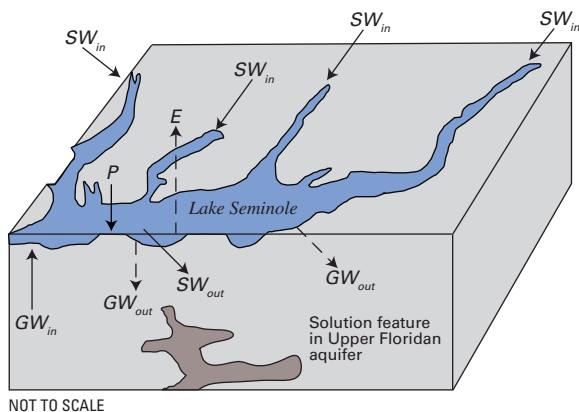
Jim Woodruff Lock and Dam and Lake Seminole.

Figure 42. Location of Lake Seminole, boundaries of the lower Apalachicola-Chattahoochee-Flint River Basin, and physiographic divisions of the Coastal Plain Province in southeastern Alabama, northwestern Florida, and southwestern Georgia (Dalton and others, 2004).

In an effort to understand the effects of the lake on the lower Apalachicola River basin, a detailed water budget of the lake was developed for the period April 2000 to September 2001 (Dalton and others, 2004). The conceptual water budget is depicted in figure 43. Precipitation was monitored at two locations on the lake. Surface-water inflows and outflows were measured at USGS stream-gaging stations. Several small, ungaged tributaries along arms of the reservoir contributed a minor amount of streamflow. Ground-water inflows and outflows were determined from a detailed ground-water-flow model for the area. Evaporation was estimated using an energy-balance method, and lake storage was determined on the basis of daily readings of stage.
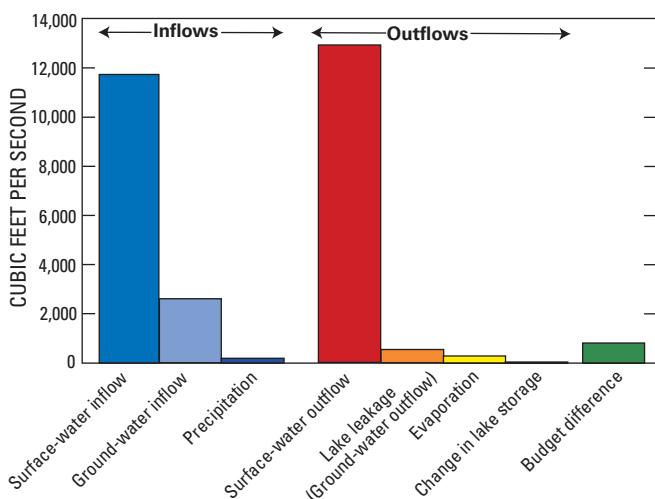
$$P + SW_{in} - SW_{out} + GW_{in} - GW_{out} - E = \Delta S$$

P	= precipitation
SW_{in}	= surface-water inflows
SW_{out}	= surface-water outflows
GW_{in}	= ground-water inflows
GW_{out}	= ground-water outflows
E	= evaporation
ΔS	= change in lake storage

Figure 43. The water budget for Lake Seminole (Dalton and others, 2004).

As shown in figure 44, surface-water flow dominates the Lake Seminole water budget. This is not surprising; the impoundment is operated for short-duration flow augmentation of the Apalachicola River for navigational purposes. The hydraulic connection between the reservoir and the underlying aquifer is reflected in the large flow of ground water into Lake Seminole and the leakage of water from the reservoir to ground water in the proximity of the dam. The direct contribution of rainfall to the reservoir is only 1 percent of the

Measuring streamflow.

total inflow to the lake. Loss of water from the reservoir to the atmosphere by evaporation is only 2 percent of the total loss of water. Because of uncertainties inherent in each method used to estimate water-budget components, the difference between water gain and loss does not exactly match the independently determined change in lake storage; the discrepancy is about 4 percent of total inflow.

It is physically impossible to account for every drop of water entering and leaving a reservoir, a drainage basin, an aquifer, or a State. Thus, water managers are faced with making decisions about water in the context of some uncertainty. In the case of Lake Seminole, customary and state-of-the-art

Figure 44. Average monthly water budget for Lake Seminole for the period April 2000 through September 2001 (Dalton and others, 2004).

Weather station in Lake Seminole.

techniques were used to quantify all water-budget components for the study period. Even so, each technique has some degree of uncertainty associated with it (Dalton and others, 2004). The 4-percent discrepancy in the water budget deserves some attention. In light of the uncertainty in each measurement technique, hydrologists would generally consider this to represent good agreement among the different measurements. However, for water managers, this 4 percent (about 560 ft^3/s) is a substantial amount of water to be unaccounted for. Because the study of water budgets in hydrologic systems is not an exact science, sound management decisions based on good science also account for uncertainties inherent in that science.
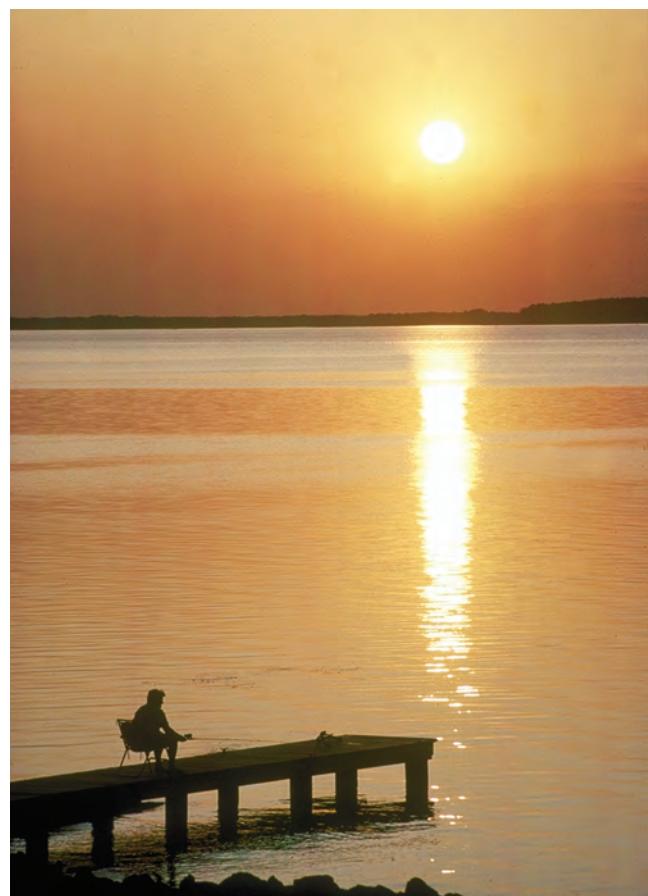
The study of Lake Seminole included an analysis of the relative importance of individual water-budget components to the overall accuracy of the water budget. This analysis, also known as a sensitivity analysis, informs water scientists as well as water managers whether the closure of the water budget could be improved by applying more accurate measurement techniques. For example, if evapotranspiration had been the dominant component of the water budget of Lake Seminole, then future measurements of evapotranspiration with detailed micrometeorological methods would be desirable. Because evapotranspiration was a small part of the water budget, it can be measured with a simpler, less accurate technique requiring fewer resources to implement. Surface-water flow dominates the water budget, so the accuracy of the water budget is closely tied to the accuracy of streamflow measurements. Independent measurements or estimates of groundwater flows also may lead to an improved water budget.

Water-allocation issues among the three States may not be resolved for years to come. On a shorter time scale, water budgets are an integral part of day-to-day dam operations. The Woodruff Dam is the most downstream of a series of six dams

Twilight fishing.

in the Apalachicola-Chattahoochee-Flint basin. Releases of water through dams in this system are coordinated to satisfy, as best possible, a multitude of needs—hydroelectric power, navigation, water quality and quantity, flood control, environment, and recreation. River stage forecasts are provided by the National Weather Service which uses complex simulation models in conjunction with stream and precipitation data obtained in real time through satellite transmission and with predicted precipitation rates. Managers prioritize current water needs in light of these forecasts to set appropriate release rates (*http://water.sam.usace.army.mil/narrativ.htm*, accessed on March 5, 2007).

Maintaining navigable waterways is one of the issues addressed by dam operators.

Wildlife.

Agriculture and Habitat: Upper Klamath Lake

Upper Klamath Lake, in south-central Oregon (fig. 45), is the main source of water for the Bureau of Reclamation's Klamath Project, a project designed to provide water for irrigation of about 180,000 acres of cropland and for two National Wildlife Refuges. The Williamson and Wood Rivers, two streams with a large amount of ground-water discharge, account for more than 50 percent of the water inflow to the lake. Small streams, ground-water seepage, and direct precipitation are secondary sources of inflow. The lake drains to the Link River, the natural outlet for the lake, and eventually contributes to the Klamath River. Water is also diverted from Upper Klamath Lake through the A Canal to provide irrigation water during the growing season. Evapotranspiration losses from the lake are also important. Flow to the Link River is controlled by the Link River Dam, which was completed in 1921. In addition to providing water for irrigation, wildlife refuges, and recreation, water in the lake and associated rivers provides aquatic habitat for a number of species, some that are endangered or threatened, such as the Lost River sucker (*Deltistes luxatus*), shortnose sucker (*Chasmistes brevirostris*), and coho salmon (*Oncorhynchus kisutch*), and some, such as chinook salmon (*Oncorhynchus tshawytscha*), that are of substantial commercial, cultural, and recreational value (Service, 2003).

Upper Klamath Lake, Oregon

Figure 45.   Map showing location of Upper Klamath Lake.

and several aquatic species. In 2001, irrigation withdrawals from the lake were curtailed for hundreds of farmers in order to maintain lake levels to protect endangered sucker habitat and provide instream flows for coho salmon. In late September 2002, tens of thousands of adult prespawned salmon and other fish died in reaches of the Klamath River 200 miles downstream from Upper Klamath Lake even though the BOR released the amount of flow called for in a "dry" hydrologic period in the National Oceanic and Atmospheric Administration's (NOAA) Fisheries' Biological Opinion. The California Department of Fish and Game (2003) attributed the deaths to low streamflow, high water temperatures, a large salmon

Upper Klamath National Wildlife Refuge, Oregon.

U.S. Fish and Wildlife Service

Is there enough water to go around? This is an important question that water managers address in attempting to balance the needs of all parties in the Klamath basin where agriculture and fisheries are important parts of the economy. Farmers need water to grow crops. Native salmon and trout need adequate flows of cool water for reproduction during critical life stages. Endangered suckers need river riparian areas and shoreline habitat for spawning and rearing. National Wildlife Refuges in the basin need water to provide habitat for water fowl migrating along the Pacific Flyway.

In the early 2000s, developing drought conditions in the Klamath basin created water shortages that affected several Native American tribes, the farming community, fishermen,

Klamath River.

U.S. Fish and Wildlife Service

return, and a proliferation of two naturally occurring patho-gens. Lynch and Risley (2003) quantified the September 2002 streamflow and water temperature conditions in the main-stem Klamath River leading up to this fish die-off. A persistent drought in the region and warmer than average weather condi-tions resulted in below-average streamflows in September 2002 (with a 10-year recurrence interval) and warmer-than-average water temperatures (with a 5-year recurrence interval).

An important step in determining if there is enough water to go around is to develop an understanding of the water bud-get of Upper Klamath Lake. The most complete water budget presently available for Upper Klamath Lake is that of Hubbard (1970). Hubbard measured or estimated all major lake inflows and outflows and changes in lake storage during a 3-year period (fig. 46). Hubbard's study has certain limitations: it spanned only 3 years, and the quantitative understanding of the relation between stage and volume of the lake has since been improved. However, because many components of the lake water budget that Hubbard measured have not been measured subsequently on a routine basis, his water balance for Upper Klamath Lake remains the most comprehensive analysis of all the inflows to and outflows from the lake and provides a good example of the value of this type of study.

Measured surface-water inflow and surface-water outflow represented the largest gains and losses of water for the lake. Precipitation falling on the lake accounted for just 7.4 per-cent of total water inputs. Evapotranspiration, determined from pan-evaporation measurements for the lake and by the Blaney-Criddle method for marshes, represented 15.7 per-cent of total water loss from the lake. That evapotranspiration losses exceeded gains from precipitation is a reflection of the semiarid climate of the region. The volume of water stored in the lake was determined by lake-stage measurements and a pre-established stage-volume table (subsequently revised). Lake storage is largely controlled by dam operations and irri-gation withdrawals by BOR and private canals. When viewed on a monthly basis, storage decreased during the growing season (because of releases for downstream flow require-ments, irrigation, and wildlife refuges) and increased from late fall to early spring. On an annual basis, there was little change
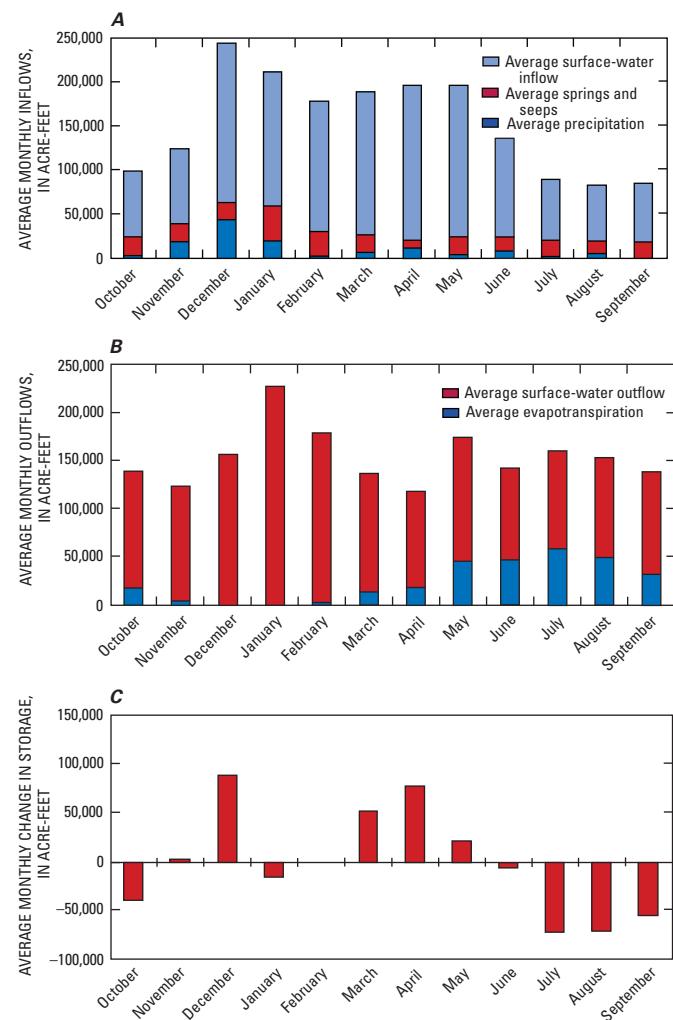
Irrigated fields in the Klamath River Basin.

Bureau of Reclamation

Lost River sucker.

Chinook salmon.

U.S. Fish and Wildlife Service

Figure 46. Average monthly water budget for Upper Klamath Lake for October 1964 through September 1967 (Hubbard, 1970): (*A*) Average monthly inflows, (*B*) average monthly outflows, and (*C*) average monthly change in storage.

U.S. Fish and Wildlife Service

Lower Klamath National Wildlife Refuge.

in storage from year to year, largely because the lake refills in most years. The lake storage and evapotranspiration estimates of Hubbard (1970) have been revised in recent years, but these revisions have not substantially altered the basic understanding of the relative proportions of various inflows to and outflows from Upper Klamath Lake put forth in that report, and it continues to serve as a useful resource for hydrologists working in the basin.

The present knowledge of ground-water inflow to Upper Klamath Lake is largely based on Hubbard's water budget. Ground-water inflow from springs and seeps could not be measured directly; it was estimated as the residual of the water budget equation. Because of this, estimates of ground-water inflow have the largest uncertainties of any of the water-budget components; they reflect the uncertainties in all the other estimates. Month-to-month variability in surface-water flows, precipitation, and evapotranspiration are expected because of seasonal weather patterns. Ground-water inflow, on the other hand, has relatively little seasonal variability but does vary in response to interannual climate cycles. Measurements show that the discharge of large spring complexes near Upper Klamath Lake vary by a factor of 1.5 to 2 in response to decadal climate cycles (M.W. Gannett, U.S. Geological Survey, written commun., 2007). Ground-water discharge directly to the lake likely varies in a similar manner. The average monthly ground-water inflow to the lake for the 36-month study period of Hubbard (1970) was about 21,000 acre-ft.

The Bureau of Reclamation's Annual Operations Plan for Upper Klamath Lake (*http://www.usbr.gov/mp/kbao/*

news/2006_Klamath_Project_Operations_Plan.pdf) is developed in accordance with the U.S. Fish and Wildlife Service and NOAA Fisheries Biological Opinions and a U.S. District Court ruling. The complex plan relies on predictions of water-budget components, especially inflow to the lake for April through September. Forecasts of inflow are provided on a monthly basis by the Natural Resources Conservation Service, which uses current data on snowpack, precipitation, and other meteorological parameters across the drainage basin, stream-flow data, and predictions of weather patterns. These inflow forecasts have large uncertainties early in the calendar year when Annual Operation Plans are being developed because of the difficulty in forecasting weather patterns accurately many months in advance. Constraints in managing the system include maintaining an adequate stage in the lake, ensuring a minimum flow in the Klamath River, delivering water to National Wildlife Refuges at historical rates, and providing farmers with sufficient quantities of irrigation water. Satisfying all these needs is a difficult proposition, especially during extended periods of below-normal precipitation because Upper Klamath Lake does not have multiyear carry-over storage. Accurate and up-to-date assessments of the water budget of the lake would provide valuable support for important water-management decisions on BOR's Klamath Project, particularly if lake-level and downstream-flow requirements are to be better coordinated with current hydrologic conditions in the basin.

Water for Humans and Ecosystems: San Pedro River Ecosystem

The San Pedro River is one of the last free-flowing rivers between Mexico and the United States. Flowing northward from the Mexican State of Sonora through southeastern Arizona to the Gila River (fig. 47), the river maintains one of the most ecologically diverse desert riparian ecosystems on Earth (The Nature Conservancy Web site: *http://www.nature.org*). The ecosystem is home to more than 100 species of mammals (20 species of bats), 40 species of reptiles and amphibians, and 100 species of butterflies. More than 300 species of birds live in or migrate through the river corridor. The U.S. Congress recognized the importance of this ecosystem when it created the San Pedro Riparian National Conservation Area in 1988 (fig. 47) to protect the area.

Protecting the ecosystem means protecting surface water in the river and ground water in the riparian zone. The ecosystem will continue to thrive only through the maintenance of its water resources. Human demands for water within the basin are growing as population increases, posing a classic challenge: Can a limited water resource be properly managed to satisfy the needs of humans and the needs of the environment they treasure? To address this question, a detailed analysis of the ground-water budget for the basin was undertaken by the U.S. Department of the Interior.

The San Pedro River is fed primarily by ground water. Late summer monsoon storms may produce short-term runoff, but the perennial flow in many stream reaches is supported by slow, long-term ground-water discharge. Recharge to the ground-water system derives principally from precipitation falling in the mountains that bound the basin on three sides; this is typical of watersheds in arid and semiarid regions of the United States. Average annual precipitation in the basin is about 16 inches (Pool and Coes, 1999) with rates higher in the mountains than on the basin floor. Recharge water takes years to travel from mountain fronts through the subsurface to the riparian zone that borders the river. Ground water in the ripar-

The San Pedro River flows northward from the Mexican State of Sonora into Arizona.

San Pedro River.

ian zone discharges to the river or is taken up by phreatophytic vegetation.

The Sierra Vista Subwatershed occupies the upper part of the U.S. portion of the river basin. Population in this area was about 68,000 in 2002 and is expected to grow to more than 83,000 by 2011 (U.S. Department of the Interior, 2005). The population is split between dispersed rural residences and urban centers such as Sierra Vista, Bisbee, and Tombstone. Ground water is the only source of water for domestic, industrial, and agricultural use. Development of ground water began in earnest sometime after 1940; by 2002, pumpage was estimated at about 16,500 acre-ft/yr. The ground-water budget for the subwatershed (U.S. Department of the Interior, 2005) is illustrated schematically in figure 48 and can be expressed as:
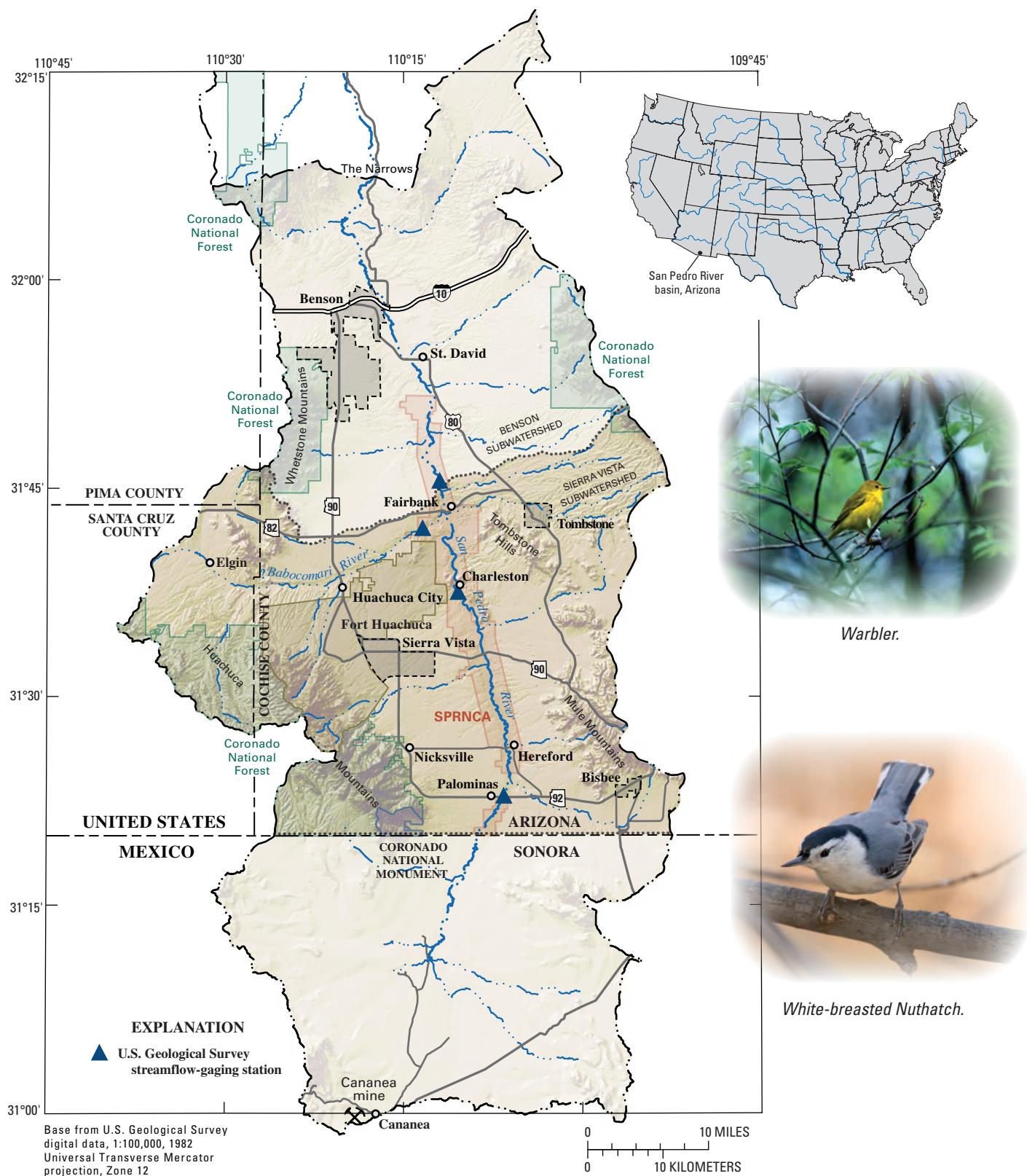
Natural Recharge + Additional Recharge + Ground-Water Inflow
=
Ground-Water Outflow + San Pedro River Base Flow + Evapotranspiration + Pumping + Change in Ground-Water Storage.
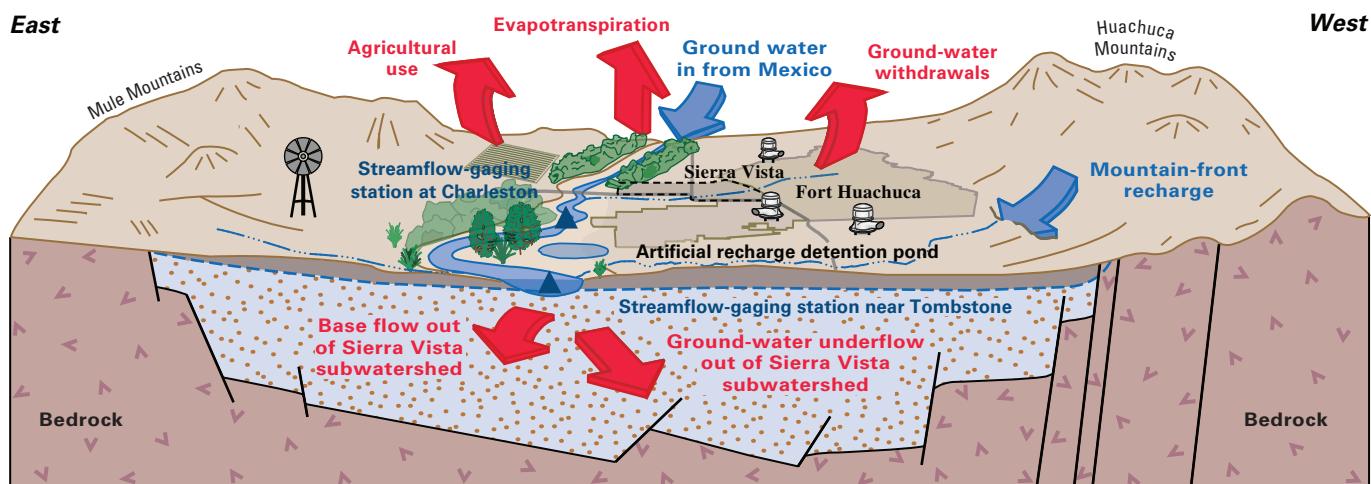
Figure 47.   Location of the upper San Pedro River basin showing the San Pedro Riparian National Conservation Area (SPRNCA).

Figure 48. Simulated annual ground-water budget for the upper San Pedro River basin (U.S. Department of the Interior, 2005).

Additional recharge refers to the return of pumped water to the aquifer through drainage of irrigation water, septic tanks, and enhanced recharge from routing of runoff from impervious areas. That routing could be unintentional, as a result of increased impervious areas such as roads, buildings, and sidewalks. It could also be the result of planned urban infiltration galleries that funnel runoff directly to the ground-water system, thus bypassing the soil zone and avoiding uptake by vegetation.

Table 8 shows values for components of the ground-water budget for a time before ground-water development (1940) and after a period of more than 60 years of development (2002). For the water budget to balance, the increase in pumping between 1940 and 2002 must be offset by one or more other water-budget components. Results of computer simulations of ground-water flow indicated that by 2002 there was a 65 percent decrease in annual ground-water discharge (base flow) to the river, 8,400 acre-ft of water was removed from ground-water storage each year, and there was a slight reduction in evapotranspiration rates. Interestingly, recent estimates of evapotranspiration based on field measurements indicate that current rates (about 10,800 acre-ft/yr) are greater than those estimated for the past (Scott and others, 2006). It is not clear if past estimates, which were not based on field estimates, are in error or if, indeed, riparian evapotranspiration rates have increased. In either regard, the ground-water flow simulations indicated that continued pumping at current rates with no additional recharge will eventually dry up the river (U.S. Department of the Interior, 2005). Such a result would have severe implications for the ecosystem.

Table 8. Annual ground-water budget (in acre-feet) for Sierra Vista subwatershed [predevelopment conditions (1940) from Corell and others (1996) and in 2002 (U.S. Department of the Interior, 2005)]. Net pumping is actual pumping minus that amount of pumped water that returned to the aquifer.

Year	Natural recharge	Ground-water inflow	Ground-water outflow	Evapotranspiration	San Pedro River base flow	Net pumping	Storage change
1940	16,000	3,000	440	8,020	9,540	1,000	0
2002	15,000	3,000	440	7,700	3,250	15,000	−8,400

U.S. Geological Survey stream gage.

The possibility of such an occurrence led to the formation of the Upper San Pedro Partnership (USPP), a group of governmental and private agencies charged with achieving sustainable yield within the basin (U.S. Department of the Interior, 2005). Sustainable yield is defined as "development and use of ground water in a manner that can be maintained for an indefinite time without causing unacceptable environmental, economic, or social consequences" (Alley and Leake, 2004, p. 12). Of course, determining what is or is not acceptable is a subjective matter that may lead to contentious debate. Regardless, the water budget in table 8 provides a starting point for determining sustainable yield. To predict consequences in time and space of future development, results from a ground-water model will be interpreted in the context of various completed and ongoing studies of basin hydrogeology and riparian water needs. In order to ensure continued flow in the San Pedro River and health of the ecosystem, managers are implementing measures designed to conserve water, thereby reducing the population's ground-water demand. At the same time, the USPP seeks to enhance additional recharge by encouraging large-scale artificial recharge. The success of these efforts depends largely on the accuracy of the ground-water budget. Continual refinement of the ground-water budget, as new data become available, is an important aspect of the management plan.

Riparian vegetation along the river stands out in the desert environment.
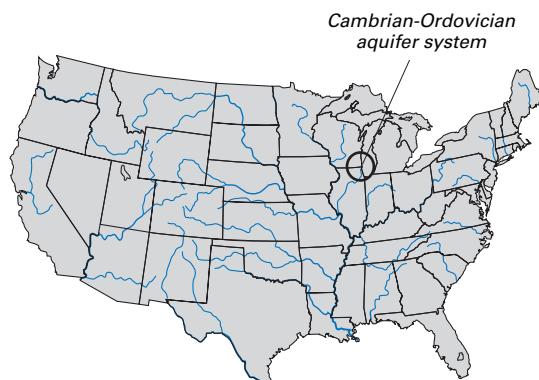
Urban Water Supply: Chicago

A prolific confined aquifer system underlies a large portion of northern Illinois and southern Wisconsin (fig. 49). The system, composed of sandstones and dolomites of Cambrian-Ordovician age, lies at depths of several hundred feet below the urban areas of Chicago and Milwaukee and has long provided drinking water for those areas. Under predevelopment conditions, recharge to the system occurred in areas to the north and west of the cities where aquifer rocks are close to land surface; ground water flowed to the east and southeast, slowly discharging to Lake Michigan and to land surface southeast of Chicago.

The response of this aquifer system to development provides an illustration of the dynamic nature of water budgets and the hydrologic cycle. The first deep well in the Chicago area was drilled in 1864 to a depth of 711 ft; it flowed at land surface at the rate of about 150 gal/minute (Visocky, 1997). In the following decades, ground-water withdrawals from the aquifer system increased sharply, coincident with rising population (fig. 50). The increased pumping resulted in substantial declines in ground-water levels in the aquifer (fig. 51). Around 1980, fears of depleting the system led public water suppliers in the area to shift their source of water to Lake Michigan. Ground-water withdrawals decreased in much of the area, and
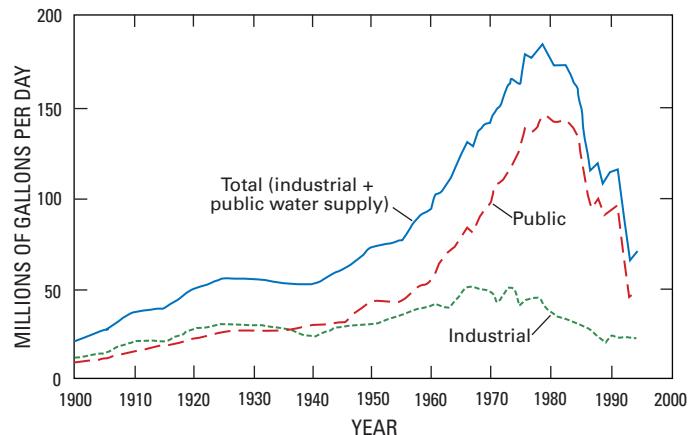
Figure 49.   Approximate location map of Cambrian-Ordovician-age aquifer system in northern Illinois and southern Wisconsin.

ground-water levels showed some recovery through the early 1990s (fig. 52). Burch (1991) predicts that by 2010, water levels will have rebounded substantially in the Chicago area, with rises of more than 600 ft in areas near former pumping centers.

How did changing development practices affect the natural water budget of the aquifer system? The large total withdrawals had multiple effects. Obviously, there was a great
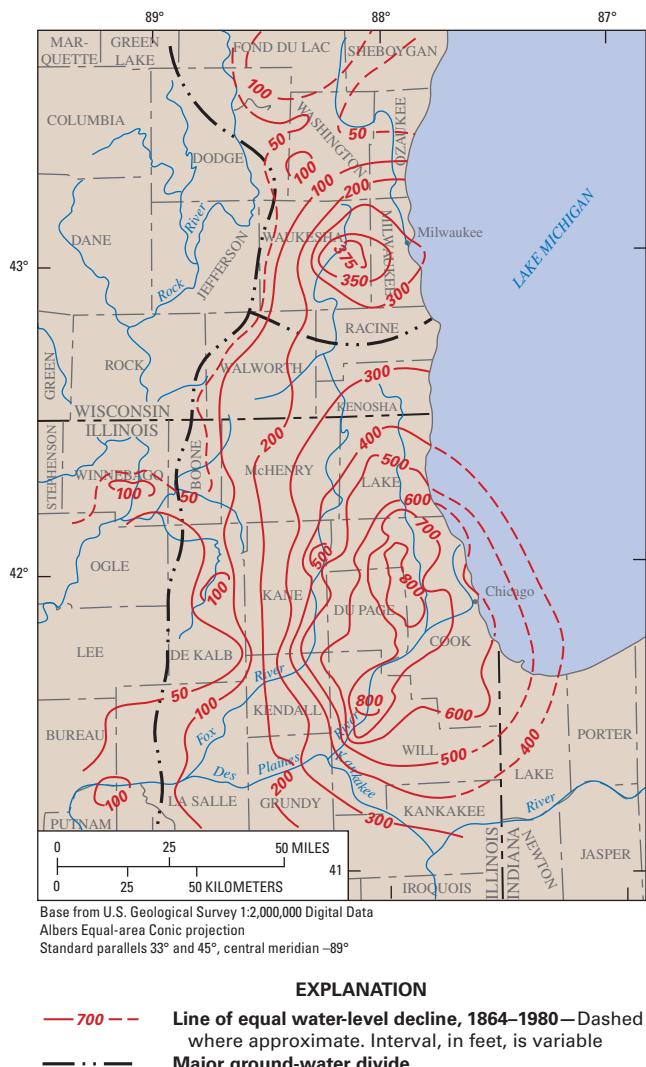
reduction in water stored in the system. The total change in storage could be roughly estimated by using the declines shown in figure 51 if a representative value of storage coefficient were available. Development has reduced the natural rates of outflow to Lake Michigan and other discharge points. In some cases, falling water levels resulted in reversals in hydraulic gradients so that some natural outflow boundaries, such as some lakes near Madison, Wis., actually became inflow boundaries (Burch, 1991). In terms of inflow to the system, predevelopment rates of recharge were likely limited by the ability of the aquifer to accept more water; that is, the system was essentially full and could not accept available recharge. Falling water levels brought on by development have created additional storage capacity, so it is likely that current rates of recharge exceed those of the past. (Of course, as previously discussed, many factors, in particular land-use changes, influence the recharge process.)

Figure 50.   Ground-water withdrawals from the Cambrian-Ordovician aquifer system in the eight-county Chicago area, 1900–94 (Visocky, 1997).
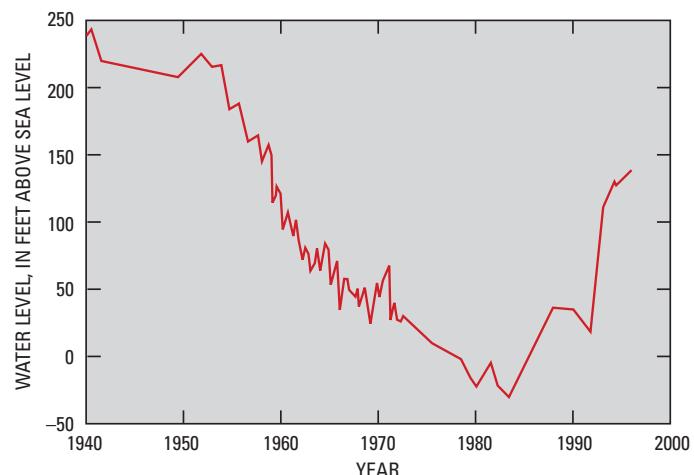
Recharge to deep aquifers that served the Chicago and Milwaukee areas occurs to the north and west of these areas where aquifer rocks are close to land surface.

Pump house in northern Illinois.

EXPLANATION

— *700* — —   **Line of equal water-level decline, 1864–1980**—Dashed where approximate. Interval, in feet, is variable

— · · —   **Major ground-water divide**

Figure 51. Ground-water level declines from 1864 to 1980 in the Cambrian-Ordivician aquifer system, Chicago and Milwaukee areas (Alley and others, 1999).

Base from U.S. Geological Survey 1:2,000,000 Digital Data
Albers Equal-area Conic projection
Standard parallels 33° and 45°, central meridian –89°

Figure 52. Trends in ground-water levels in the deep Cambrian-Ordivician aquifer system. Data are from a well in Cook County near Chicago (after Visocky, 1997).

Outcrops of the St. Peter Sandstone near Monticello, Wisconsin, July 26, 1907.

Humans are part of the hydrologic cycle, ...

U.S. Fish and Wildlife Service

...as are all animals and plants.

U.S. Army Corps of Engineers

Chemical and energy budgets of lakes, rivers, and aquifers are directly linked to water budgets.

Concluding Remarks

A water budget states that the rate of change in water stored in an accounting unit, such as a watershed, is balanced by the rate at which water flows into that unit minus the rate at which water flows out of it. Universally applicable, water budgets can be constructed at any spatial scale—an agricultural field, a wetland, an aquifer, a lake, a watershed, and even the Earth itself and at any temporal scale, from seconds to years to millennia. While theoretically simple, water budgets, in practice, are often difficult to determine. Inherent uncertainties pervade all techniques used to measure water storage and flux. In addition, the dynamic nature of the hydrologic cycle implies that storage and flux terms change over time.

As the human population on Earth continues to grow, so will its demands for water. Balancing the water needs of humans with those of the many ecosystems on Earth will continue to be a challenge. Water budgets provide a means for evaluating the availability and sustainability of a water supply. The link among all components of a water budget serves as a basis for predicting how a natural or human-induced change to one component, such as ground-water extraction, may be reflected in other components, such as streamflow or evapotranspiration. When viewed with an understanding of the underlying hydrologic processes and the uncertainties associated with quantifying those processes, water budgets form a foundation for evaluating water-resources and environmental planning and management options.

Science and technology can assist water-resources and environmental management by addressing important questions related to the hydrologic cycle, water use, water needs, and water availability and sustainability. These questions include:

How much water do humans use?

How much water do ecosystems need to flourish?

How much water is available for humans and ecosystems? Where is this water?

How does the hydrologic cycle naturally change over time?

In what ways do human activities affect the hydrologic cycle?

How will changes in the hydrologic cycle affect water availability and use?

What effects do uncertainties in estimates of water storage and movement have on our understanding of water budgets in general and of the availability and sustainability of water resources in particular?

Steps to developing answers to these questions include monitoring domestic, agricultural, and industrial water use; conducting inventories of ecosystems and their water needs; and undertaking surveys and assessments of water resources. Insight into the variability, both natural and that induced by human activity, of hydrologic cycles can be obtained with improved methods for studying the hydrologic cycle—more accurate instruments, new designs for field studies, alternative methods for interpreting remotely sensed data, new simulation models of water movement through various parts of the hydrologic cycle, and improved methods for predicting water use by humans, plants, and animals. Science and technology can also assist in the development and improvement of decision support systems that allow managers to evaluate various operational options.

Accessible freshwater is a limited resource that humans must share among themselves and with the environment. Supplies of freshwater are not available everywhere. Hence, throughout history humans have constructed systems for conveying and storing water. The resource gained when humans import water to one location is accompanied by a loss of water at another location. In terms of water amounts, every gain is balanced by an equal loss. In terms of economic, cultural, and ecological values, the question of whether gains balance losses must be evaluated by society as a whole. Perhaps the greatest challenge to maintaining a sustainable future for our water resources is the development of policies and laws that balance the many water needs of humans with the water needs of their environment. Water budgets form a foundation upon which those policies and laws can be developed.

Natural Resources Conservation Service

Human activity affects the hydrologic cycle.

Even the most accurate measurements of water fluxes contain some uncertainty.

Holgate Glacier 1909

Holgate Glacier 2004

The hydrologic cycle varies naturally over time.

References Cited

Alley, W.M., and Emery, P.A., 1986, Groundwater model of the Blue River Basin, Nebraska—20 years later: Journal of Hydrology, v. 85, p. 225–249.

Alley, W.M., LaBaugh, J.W., and Reilly, T.E., 2005, Groundwater as an element of the hydrologic cycle, *in* Bear, J., Encyclopedia of Hydrological Sciences, v. 4, pt. 13: New York, John Wiley and Sons, Inc., p. 2215–2228.

Alley, W.M., and Leake, S.A., 2004, The journey from safe yield to sustainability: Ground Water, v. 42, no. 1, p. 12–16.

Alley, W.M., Reilly, T.E., and Franke, O.L., 1999, Sustainability of ground-water resources: U.S. Geological Survey Circular 1186, 86 p.

Anderson, T.W., 1968, Electric analog analysis of ground-water depletion in central Arizona: U.S. Geological Survey Water-Supply Paper 1860, 21 p.

Antweiler, R.C., Goolsby, D.A., and Taylor, H.E., 1995, Nutrients in the Mississippi River, *in* Meade, R.H., ed., Contaminants in the Mississippi River, 1987–92: U.S. Geological Survey Circular 1133, p. 73–86.

Brye, K.R., Norman, J.M., Bundy, L.G., and Gower, S.T., 1999, An equilibrium tension lysimeter for measuring drainage through soil: Soil Science Society of America, v. 63, no. 3, p. 536–543.

Brye, K.R., Norman, J.M., Bundy, L.G., and Gower, S.T., 2000, Water-budget evaluation of prairie and maize ecosystems: Soil Science Society of America, v. 64, no. 2, p. 715–724.

Burch, S.L., 1991, The new Chicago model—A reassessment of the impact of Lake Michigan allocation on the Cambrian-Ordovician aquifer system in northeastern Illinois: Champaign, Ill., Illinois State Water Survey, Research Report 119, 52 p.

California Department of Fish and Game, 2003, September 2002 Klamath River fishkill—Preliminary analysis of contributing factors: Redding, State of California, The Resources Agency, Department of Fish and Game, 63 p.

Carr, J.E., Chase, E.B., Paulson, R.W., and Moody, D.W., 1990, National water summary 1987—Hydrologic events and water supply and use: U.S. Geological Survey Water-Supply Paper 2350, 553 p.

Chow, V.T., Maidment, D.R., and Mays, L.W., 1988, Applied hydrology: New York, McGraw-Hill, 572 p.

Collier, Michael, Webb, R.H., and Schmidt, J.C., 1996, Dams and rivers—A primer on the downstream effects of dams: U.S. Geological Survey Circular 1126, 94 p.

Corell, S.W., Corkhill, Frank, Lovvik, D., and Putnam, F., 1996, A groundwater flow model of the Sierra Vista subwatershed of the upper San Pedro basin, southeastern Arizona: Arizona Department of Water Resources Modeling Report no. 10, 107 p.

Cowardin, L.M., Carter, Virginia, Golet, F.C., and LaRoe, E.T., 1979, Classification of wetlands and deepwater habitats of the United States: U.S. Fish and Wildlife Service Office of Biological Services Report FWS/OBS–79/31, 131 p.

Dalton, M.S., Aulenbach, B.T., and Torak, L.J., 2004, Groundwater and surface-water flow and estimated water budget for Lake Seminole, southwestern Georgia and northwestern Florida: U.S. Geological Survey Scientific Investigations Report 2004–5073, 49 p.

Daly, C.G., Neilsen, R.P., and Phillips, D.L., 1994, A statistical-topographic model for mapping climatological precipitation over mountainous terrain: Journal of Applied Meteorology, v. 33, p 140–158.

Doorenbos, J., and Pruitt, W.O., 1975, Crop water requirements: Irrigation and Drainage Paper, FAO, Rome, 179 p.

Driscoll, C.T., Driscoll, K.M., Mitchell, M.J., and Raynal, D.J., 2003, Effects of acidic deposition on forest and aquatic ecosystems in New York State: Environmental Pollution, v. 123, no. 3, p. 327–336.

Fisher, L.H., and Healy, R.W., 2007, Water movement within the unsaturated zone in four agricultural areas of the United States: Journal of Environmental Quality, in press.

Freeze, R.A., and Banner, J., 1970, The mechanism of natural groundwater recharge and discharge 2. Laboratory column and field measurements: Water Resources Research, v. 6, p. 138–155.

Frevert, D.K., Fulp, T.J., Zagora, E.A., Leavesley, G.H., and Lins, H.F., 2006, Watershed and river systems management program—Overview of capabilities: Journal of Irrigation and Drainage Engineering, v. 132, no. 2, p. 92–97.

Frost, Robert, 1928, West-running brook: New York, Holt and Company.

Galloway, D.L., Jones, D.R., and Ingebritsen, S.E., eds., 1999, Land subsidence in the United States: U.S. Geological Survey Circular 1182, 177 p.

Gat, J.R., and Gonfiantini, R., eds., 1981, Stable isotope hydrology—Deuterium and oxygen-18 in the water cycle: Vienna, International Atomic Energy Agency, 337 p.

Gibran, Kahlil, 1923, The Prophet: New York, A.A. Knopf (1983 ed.), 107 p.

Gilpin, Laura, 1949, The Rio Grande, river of destiny—An interpretation of the river, the land, and the people: New York, Duell, Sloan and Pearce, 243 p.

Healy, R.W., and Cook, P.G., 2002, Using groundwater levels to estimate recharge: Hydrogeology Journal, v. 10, p. 91–109.

Healy, R.W., de Vries, M.P., and Sturrock, A.M., 1989a, Evapotranspiration and microclimate at a low-level radioactive-waste disposal site in northwestern Illinois: U.S. Geological Survey Water-Supply Paper 2327, 44 p.

Healy, R.W., Gray, J.R., de Vries, M.P., and Mills, P.C., 1989b, Water balance at a low-level radioactive-waste disposal site: Water Resources Bulletin, v. 25, no. 2, p. 381–390.

Heath, R.C., 1983, Basic ground-water hydrology: U.S. Geological Survey Water-Supply Paper 2220, 84 p.

Hevesi, J.A., Istok, J.D., and Flint, A.L., 1992, Precipitation estimation in mountainous terrain using multivariate geostatistics. Part I—Structural analysis: Journal of Applied Meteorology, v. 31, no. 7 p. 661–676.

Hubbard, L.L., 1970, Water budget of Upper Klamath Lake, Southwestern Oregon: U.S. Geological Survey Hydrologic Investigations Atlas HA–351.

Huff, F.A., and Schickedanz, P.T., 1972, Space-time uncertainties in precipitation measurement: Proceedings of International Symposium on Uncertainties in Hydrologic and Water Resource Systems, Tucson, Arizona, p. 395–409.

Hutson, S.S., Barber, N.L., Kenny, J.F., Linsey, K.S., Lumia, D.S., and Maupin, M.A., 2004, Estimated use of water in the United States in 2000: U.S. Geological Survey Circular 1268, 46 p.

Institute of Hydrology, 1980, Low flow studies: Walingford, Oxon, United Kingdom, Research Report no. 1.

Jensen, M.E., Burman, R.D., and Allen, R.G., 1990, Evapotranspiration and irrigation water requirements: New York American Society of Civil Engineers Manuals and Reports on Engineering Practice no. 70, American Society of Civil Engineers, 332 p.

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

Katz, B.G., Lee, T.M., Plummer, L.N., and Busenberg, E., 1995, Chemical evolution of groundwater near a sinkhole lake, northern Florida, 1. Flow patterns, age of groundwater and influence of lake water leakage: Water Resources Research, v. 31, no. 6, p. 1549–1564.

Kimball, B. A., Runkel, R.L., Cleasby, T.E., and Nimick, D.A., 2004, Quantification of metal loading by tracer injection and synoptic sampling, 1997–98: U.S. Geological Survey Professional Paper 1652, p. 191–262.

Kitching, R., and Shearer, T.R., 1982, Construction and operation of a large undisturbed lysimeter to measure recharge to the Chalk aquifer, England: Journal of Hydrology, v. 58, p. 267–277.

Konikow, L.F., 1986, Predictive accuracy of a ground-water model—Lessons from a postaudit (Arizona USA): Ground Water, v. 24, no. 2, p. 173–184.

Lahvis, M.A., Baehr, A.L., and Baker, R.J., 1999, Quantification of aerobic biodegradation and volatilization rates of gasoline hydrocarbons near the water table under natural attenuation conditions: Water Resources Research, v. 35, no. 3, p. 753–765.

Lapham, W.W., 1989, Use of temperature profiles beneath streams to determine rates of vertical ground-water flow and vertical hydraulic conductivity: U.S. Geological Survey Water-Supply Paper 2337, 35 p.

Linsley, Jr., R.K., Kohler, M.A., and Paulhus, J.L., 1982, Hydrology for engineers (3d ed.): New York, McGraw-Hill, 508 p.

Liu, J., Chen, J.M., and Cihlar, J., 2003, Mapping evapotranspiration based on remote sensing—An application to Canada's landmass: Water Resources Research, v. 39 1189, doi: 10.1029/2002WR001680.

Lorenz, D.W., and Delin, G.N., 2007, A regression model to estimate regional ground water recharge: Ground Water, v. 45, no. 2, p. 196–208.

Luckey, R.R., and Becker, M.F., 1999, Hydrogeology, water use, and simulation of flow in the High Plains aquifer in northwestern Oklahoma, southeastern Colorado, southwestern Kansas, northeastern New Mexico, and northwestern Texas: U.S. Geological Survey Water-Resources Investigations Report 99–4104, 68 p.

Lvovitch, M.I., 1973, The global water balance: U.S. National Committee for the International Hydrological Decade, National Academy of Science Bulletin, no. 23, p. 28–42.

Lynch, D.D., and Risley, J.C., 2003, Klamath River basin hydrologic conditions prior to the September 2002 die-off of salmon and steelhead: U.S. Geological Survey Water-Resources Investigations Report 2003–4099, 10 p.

Makarov, Oleg, 1988, Preface, in Kelley, K.W., ed., The home planet: Reading, Mass., Addison-Wesley, 256 p.

Matsin, L.G., 1991, The roles of magma and ground-water in the phreatic eruptions at Inyo Craters, Long Valley caldera, California: Bulletin of Volcanology, v. 53, no. 8., p. 579–596.

Mayo, L.R., Trabant, D.C., and March, R.S., 2004, A 30-year record of surface mass balance (1966–95) and motion and surface altitude (1975–79) at Wolverine Glacier, Alaska: U.S. Geological Survey Open-File Report 2004–1069, 114 p.

McGuire, V.L., Johnson, M.R., Schieffer, R.L., Stanton, J.S., Sebree, S.K., and Verstraeten, I.M., 2003, Water in storage and approaches to ground-water management, High Plains aquifer, 2000: U.S. Geological Survey Circular 1243, 51 p.

Meinzer, O.E., 1942, Hydrology, chap. 1 *of* Meinzer, O.E., ed., Physics of the Earth—v. IX: New York, Dover Publications, 712 p.

Molnau, Myron, Rawls, W.J., Curtis, D.L., and Warnick, C.C., 1980, Gage density and location for estimating mean annual precipitation in mountainous areas: Water Resources Bulletin, v. 16, no. 3, p. 428–432.

Nace, R.L., 1967, Are we running out of water?: U.S. Geological Survey Circular 536, 7 p.

National Research Council, 1999, Global environmental change—Research pathways for the next decade: Washington, D.C., National Academy Press, 595 p.

National Science and Technology Council, 2004, Science and technology to support fresh water availability in the United States: Washington, D.C., Office of Science and Technology, 19 p.

Oke, T.R., 1978, Boundary layer climates: New York, Methuen, 372 p.

Paulson, R.W., Chase, E.B., Roberts, R.S., and Moody, D.W., 1991, National water summary 1988–89—Hydrologic events and floods and droughts: U. S. Geological Survey Water-Supply Paper 2375, 591 p.

Phillips, F.M., 1994, Environmental tracers for water movement in desert soils of the American southwest: Soil Science Society of America Journal, v. 58, no. 1, p. 15–24.

Plummer, L.N., 1993, Stable isotope enrichment in paleo-waters in the southeast Atlantic coastal plain, United States: Science, v. 262, p. 2016–2020.

Plummer, L.N., Bexfield, L.M., Anderholm, S.K., Sanford, W.E., and Busenberg, E., 2004, Hydrochemical tracers in the middle Rio Grande Basin, USA—Conceptualization of groundwater flow: Hydrogeology Journal, v. 12, no. 4, p. 359–388.

Plummer, L.N., Busenberg, E., Böhlke, J.K., Nelms, D.L., Michel, R.L., and Schlosser, P., 2001, Groundwater residence times in Shenandoah National Park, Blue Ridge Mountains, Virginia, USA—A multi-tracer approach: Chemical Geology, v. 179, p. 93–111.

Pool, D.R., and Coes, A.L., 1999, Hydrogeologic investigations of the Sierra Vista subwatershed of the upper San Pedro Basin, Cochise County, southeast Arizona: U.S. Geological Survey Water-Resources Investigations Report 99–4197, 38 p.

Quattrochi, D.A., and Luvall, J.C., 1999, Thermal infrared remote sensing for analysis of landscape ecological processes—Methods and applications: Landscape Ecology, v. 14, p. 577–598.

Rantz and others, 1982, Measurement and computation of streamflow, v. 1, Measurement of stage and discharge: U.S. Geological Survey Water-Supply Paper 2175, 284 p.

Rasmussen, W.C., and Andreasen, G.E., 1959, Hydrologic budget of the Beaverdam Creek basin, Maryland: U.S. Geological Survey Water-Supply Paper 1472, 106 p.

Renken, R.A., Cunningham, K.J., Zygnerski, M.R., Wacker, M.A., Shapiro, A.M., Harvey, R.W., Metge, D.W., Osborn, C.L., and Ryan, J.N., 2005, Assessing the vulnerability of a municipal well field to contamination in a karst aquifer: Environmental and Engineering Geoscience, v. 11, no. 4, p. 319–331.

Roethke, Theodore, 1948, "The Lost Son" and other poems: N.Y., Doubleday, 120 p.

Rosenberg, N.J., Blad, B.L., and Verma, S.B., 1983, Microclimate the biological environment, (2d ed.): New York, John Wiley and Sons, 495 p.

Rutledge, A.T., 1998, Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records—Update: U.S. Geological Survey Water-Resources Investigations Report 98–4148, 43 p.

Scanlon, B.R., Healy, R.W., and Cook, P.G., 2002, Choosing appropriate techniques for quantifying groundwater recharge: Hydrogeology Journal, v. 10, p. 18–39.
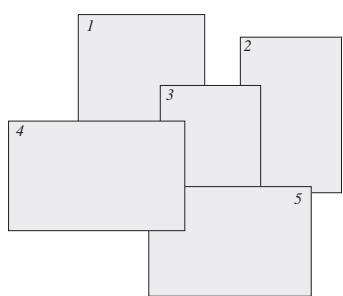
Schulze-Makuch, Dirk, Carlson, D.A., Cherkauer, D.S., and Malik, Peter, 1999, Scale dependency of hydraulic conductivity in heterogeneous media: Ground Water, v. 37, no. 6, p. 904–919.

Scott, R.L., Williams, D.G., Goodrich, D.C., Cable, W.L., Levick, L.R., McGuire, Roberta, Gazal, R.M., Yepez, E.A., Ellsworth, Patrick, and Huxman, T.E., 2006, Determining the riparian ground-water use within the San Pedro National Conservation area and the Sierra Vista Subwatershed, *in* Leenhouts, J.M., Stromberg, J.C., and Scott, R.L., eds., Hydrologic requirements of and consumptive ground-water use by riparian vegetation along the San Pedro River, Arizona: U.S. Geological Survey Scientific Investigations Report 2005–5163, p. 107–142.

Sellers, W.D., 1965, Physical climatology: Chicago, University of Chicago Press, 272 p.

Seo, D.J., 1998, Real-time estimation of rainfall fields using rain gage data under fractional coverage conditions: Journal of Hydrology, v. 208, p. 25–36.

Seo, D.J., Breidenbach, J.P., and Johnson, E.R., 1999, Real-time estimation of mean field bias in radar rainfall data: Journal of Hydrology, v. 223, p. 131–147.

Service, R.F., 2003, 'Combat biology' on the Klamath: Science, v. 300, p. 36–39.

Sophocleous, M.A., 2000, From safe yield to sustainable development of water resources—The Kansas experience: Journal of Hydrology, v. 235, p. 27–43.

Steuer, J.J., and Hunt, R.J., 2001, Use of a watershed-modeling approach to assess hydrologic effects of urbanization, North Fork Pheasant Branch basin near Middleton, Wisconsin: U.S. Geological Survey Water-Resources Investigations Report 2001–4113.

Stevens, Wallace, 1937, "The man with the blue guitar" and other poems: New York, A.A. Knopf, 82 p.

Stonestrom, D.A., and Constantz, Jim, 2003, Heat as a tool for studying the movement of ground water near streams: U.S. Geological Survey Circular 1260, 96 p.

Taylor, C.J., and Alley, W.M., 2001, Ground-water-level monitoring and the importance of long-term water-level data: U.S. Geological Survey Circular 1217, 68 p.

Thornthwaite, C.W., and Mather, J.R., 1955, The water balance: Publications in Climatology, v. VIII, no. 1, Drexel Institute of Technology, Laboratory of Climatology, 86 p.

Tiedeman, C.R., Goode, D.J., and Hsieh, P.A., 1997, Numerical simulation of ground-water flow through glacial deposits and crystalline bedrock in the Mirror Lake area, Grafton County, New Hampshire: U.S. Geological Survey Professional Paper 1572, 50 p.

Updike, John, 1989, Self–Consciousness: Memoirs, New York, A.A. Knopf.

U.S. Department of the Interior, 2005, Water management of the regional aquifer in the Sierra Vista subwatershed, Arizona—2004: Washington, D.C., Report to Congress, 36 p.

Visocky, A.P., 1997, Water-level trends and pumpage in the deep bedrock aquifers in the Chicago region, 1991–1995: Champaign, Illinois State Water Survey, Circular 182, 32 p.

Vorosmarty, C.J., and Sahagian, D., 2000, Anthropogenic disturbance of the terrestrial water cycle: BioScience, v. 50, no. 9, p. 753–765.

Webb, R.H., Leake, S.A., and Turner, R.M., 2007, The ribbon of green—Change in riparian vegetation in the Southwestern United States: Tucson, University of Arizona Press, 470 p.

Winter, T.C., 1984, Geohydrologic setting of Mirror Lake, West Thornton, New Hampshire: U.S. Geological Survey Water-Resources Investigations 84–4266, 61 p.

Winter, T.C., ed., 2003, Hydrological, chemical, and biological characteristics of a prairie pothole wetland complex under highly variable climate conditions—The Cottonwood Lake area: U.S. Geological Survey Professional Paper 1675, 109 p.

Winter, T.C., Harvey, J.W., Franke, O.L., and Alley, W.M., 1998, Ground water and surface water—A single resource: U.S. Geological Survey Circular 1139, 79 p.

Yorke, T.H., and Oberg, K.A., 2002, Measuring river velocity and discharge with acoustic Doppler profilers: Flow Measurement and Instrumentation, v. 13, p. 191–195.

Young, M.H., Wierenga, P.J., and Mancino, C.F., 1996, Large weighing lysimeters for water use and deep percolation studies: Soil Science, v. 161, no. 8, p. 491–501.

J.C. Koch

"It is this backward motion toward the source,
Against the stream, that most we see ourselves in,
The tribute of the current to the source.
It is from this in nature we are from.
It is most of us."

Robert Frost, from West-Running Brook (1928)

Photograph credits:

water inflow,
nes, canals)

Evapotranspiration

ater inflow