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146: Aquifer Recharge

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Aquifer recharge is important both for hydrologic understanding and for effective water resource management. Temporal and spatial patterns of unsaturated-zone processes such as infiltration largely determine its magnitude. Many techniques of recharge estimation exist. Water budget methods estimate all terms in the continuity equation except recharge, which is calculated as the residual. Detailed hydrologic models based on water-budget principles can produce recharge estimates at various scales. Empirical methods relate recharge to meteorologic and geographic parameters for a specific location. Surface-water methods include stream-hydrograph analyses to estimate baseflow (groundwater discharge) at lower elevations in a watershed, which is taken to equal the recharge that has occurred at higher elevations. Subsurface methods include analysis of water-table fluctuations following transient recharge events, as well as diverse unsaturated-zone methods. The zero-flux plane method determines the recharge rate from the change in water storage beneath the zero-flux depth, a boundary between water moving upwards due to evapotranspiration and water moving downward due to gravity. Lysimeter methods use buried containers filled with vegetated soil to mimic natural conditions. Water exiting the bottom is considered to be recharge. Darcian methods for estimating flux densities use unsaturated hydraulic conductivities and potential gradients, indicating recharge rates under appropriate conditions. Chemical mass-balance methods use conservative tracers that move with recharging water. Tracer concentrations in deep unsaturated-zone water, together with tracer input rates, indicate recharge rates. Distinct chemical "markers" can indicate travel times, hence, recharge rates. Thermal methods use heat as a tracer. Moving water perturbs temperature profiles, allowing recharge estimation. Geophysical methods estimate recharge based on water-content dependence of gravitational, seismic, and electromagnetic properties of earth materials.

INTRODUCTION

Defined as water that moves from the land surface or unsaturated zone into the saturated zone, aquifer recharge is vitally important for understanding the hydrologic cycle as well as for applications to water-resource management. The definition used here excludes saturated flow between aquifers so it might be more precisely termed "aquifer-system" or "saturated-zone" recharge. This definition avoids double-accounting in large-scale studies. Recharge is commonly expressed as a volume [L^3], typical units being m^3 or acre-ft. Recharge rate expresses either a flux [L^3T^{-1}] into a specified portion of aquifer, or a flux density [LT^{-1}] (volume per unit surface area) into an aquifer at a point. Over the long term, recharge naturally balances the total losses of water from the saturated zone. In some systems that are characterized by low permeabilities and sensitivity to climatic change, this long-term balance can be discerned only on millennial timescales.

Because it represents replenishment of aquifers critical to maintaining water supplies and ecosystems, recharge has obvious practical importance, especially where ground water is extracted for human use. It is a vital component for evaluating sustainability, as streamflow is for surface water. In the hydrologic balance of an aquifer, recharge processes act in opposition to discharge processes, so the relative magnitudes of recharge and discharge rate give a basic indication of the health of the aquifer and related systems.

Quantitative estimation of recharge rate contributes to the understanding of large-scale hydrologic processes. It

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is important for evaluating the sustainability of groundwater supplies, though it does not equate with a sustainable rate of extraction (Bredehoeft *et al.*, 1982). Where contamination of an aquifer is a concern, the flux of water into the aquifer is essential information; estimating the recharge rate is a first step toward predicting the rate of solute transport to the aquifer. In cases where advection dominates the transport of contaminants (*see* **Chapter 152, Modeling Solute Transport Phenomena, Volume 4**), little additional hydraulic information may be needed to estimate travel times and solute fluxes. Moreover, recharging fluxes and their distribution need to be known for assessment of aquifer vulnerability to contamination, prediction of zones of significant contamination, and evaluation of remedial measures.

Our aim in this article is to provide an understanding of recharge in terms of its processes, its role as a component of the hydrologic cycle, and means for its local or regional estimation. We approach it mainly from the perspective of natural science, emphasizing recharge from natural processes, but also with attention to anthropogenic sources such as irrigation, wastewater disposal, and deliberate augmentation of groundwater. The discussion of estimation techniques makes up the largest part of this article, and we have included much description and explanation of basic recharge processes within this discussion.

RECHARGE IN NATURE AND APPLICATIONS

Sources and Basic Processes

The nature of the water source to a large extent determines which subsurface processes are relevant or dominant in the transport of water to the aquifer. In addition to precipitation, which is commonly the dominant source, other possible sources include surface-water bodies, irrigation, and artificial recharge.

In some cases, perennial or ephemeral surface water in the form of rivers, canals, and lakes may enter the aquifer directly (Winter *et al.*, 1998). Surface-water bodies, however, are not always recharge sources; they can instead be associated with aquifer discharge. A reach of a stream is considered "gaining" or "losing" depending on whether aquifer discharge or recharge dominates over that reach. If there is no unsaturated zone, recharge and discharge may proceed by analogous processes, in each case dictated by the net driving force (usually a combination of gravity and the pressure gradient) at the bottom of the surface-water body. Surface water in direct contact with the saturated zone can be treated in terms of surface water-groundwater interaction, for example, with stream-gauging. This is described further below and in **Chapter 145, Groundwater as an Element in the Hydrological Cycle, Volume 4**. In some cases, especially in drier climates, water from a body of surface water first travels through the unsaturated zone before reaching the aquifer.

Where recharge occurs from precipitation or other sources through an unsaturated zone, general expectations are (i) that gravity is the dominant driving force for recharge (as matric-pressure gradients would typically act to oppose recharge or would balance out over time) and (ii) that the hydraulic conductivity is markedly less in the unsaturated than in the saturated zone. It is not always possible to make a direct analogy between the recharge and discharge processes as with gaining and losing streams. Capillary forces may draw aquifer water upward into the unsaturated zone, but its future transport and ultimate discharge from the aquifer occur to a large extent through saturated-zone processes. Thus unsaturated-zone processes are critical to the evaluation of recharge in practical circumstances.

Deliberate or artificial-recharge operations frequently employ surface-water ponds or streams to put water into the subsurface, though in some applications the water for recharge is injected at significant depth. Natural and artificial recharge mostly involve the same types of processes, though artificial recharge may easily create situations that would be unusual in nature, for example, a thick unsaturated zone under a pond. Irrigation and other water-use practices can contribute to recharge, typically in ways analogous to precipitation or surface water, depending on the mode of application.

Variation

Recharge rates vary considerably in time and space. Recharge often occurs episodically in response to storms and other short-term, high-intensity inputs. Some, perhaps most, of the water that becomes recharge may be spatially concentrated in narrow channels as it passes through the unsaturated zone. Time- and space-concentration are important because for available water at the land surface, there is competition among alternative possible fates such as recharge, runoff, and evapotranspiration. For a given amount of infiltration, temporal concentration favors recharge because it entails shorter residence times for water in the portions of the soil from which evapotranspiration takes place. Similarly, a larger fraction will become recharge if it is concentrated in narrow channels such as fingers or macropores, not only because this tends to hasten its passage through the unsaturated zone, but also because the water then occupies less of the volume of soil from which evapotranspiration takes place.

Temporal Distribution of Recharge

Temporal variation occurs in general with seasonal or short-term variations in precipitation, runoff, or evapotranspiration. Evapotranspiration, in particular, is important because it may extract most or all of the water that infiltrates. The time of greatest recharge is normally when much water is present, so that the processes other than recharge, especially evapotranspiration, are overwhelmed. Because flow rates are much greater in soil at high than at low water content (*see* **Chapter 150, Unsaturated Zone Flow Processes, Volume 4**), the downward flow important for recharge is much greater at high water contents. The slower vertical flow occurring with moderate or low water contents allows more water to remain for extraction from the soil without recharge.

Temporal concentration commonly occurs during storms, floods, and snowmelt. Variability is especially evident in thin unsaturated zones, where recharge may occur within a short time, often much less than one year (e.g. Delin *et al.*, 2000), and sometimes much less than one day (e.g. Gburek and Folmar, 1999). In deep unsaturated zones, recharge may be homogenized over many years such that it occurs with constant flux even though fluxes at shallow depths are erratic. Figure 1 illustrates temporal effects with an example from Dreiss and Anderson (1985).

Variation over time can also be significant over a longer term. Climatic change, for example, can systematically alter recharge patterns over hundreds or thousands of years. Sometimes, especially in arid regions with thick unsaturated zones, evidence for such change may be present in the observable distributions of water and other substances within the unsaturated zone (Tyler *et al.*, 1996).

Spatial Distribution of Recharge

Spatial variation occurs with climate, topography, soil, geology, and vegetation. For example, a decrease of slope or increase of soil permeability may lead to greater infiltration and greater recharge.

Topography influences recharge in several ways. It controls the generation of runoff. Steep slopes, especially if they are convex, may decrease infiltration and recharge. Swales and other concavities that tend to collect surface water may increase infiltration and recharge. Spatial concentration and increased recharge typically occur in depressions and channels, where higher water contents promote rapid movement by increasing the hydraulic conductivity (K), the amount of preferential flow, and the downward driving force at a wetting front.

Vegetation influences recharge mainly through its water-distributing activity. Where water is a significant limiting factor of vegetation growth, roots usually extract most of the water that infiltrates. This may be the case in all but the most humid climates. The result is that evapotranspiration may nearly equal infiltration, leaving recharge as a small fraction of infiltration. Artificial control of vegetation may counter natural influences. Agricultural practices, for example, may leave land fallow part of the time, decreasing total annual evapotranspiration from what it would be naturally. The depth range over which roots can extract water is frequently an important issue. Depending on the plant species, soil, and climate, roots may be present and active at depths of 10 m or more.

Another effect of evapotranspiration is that by drying the root zone, it enhances gradients that may be drawing water upward from the water table, in effect causing discharge through the unsaturated zone. The depth of the water table is a major influence in this process. Such discharge is likely to be greater if the water table is within or near the root

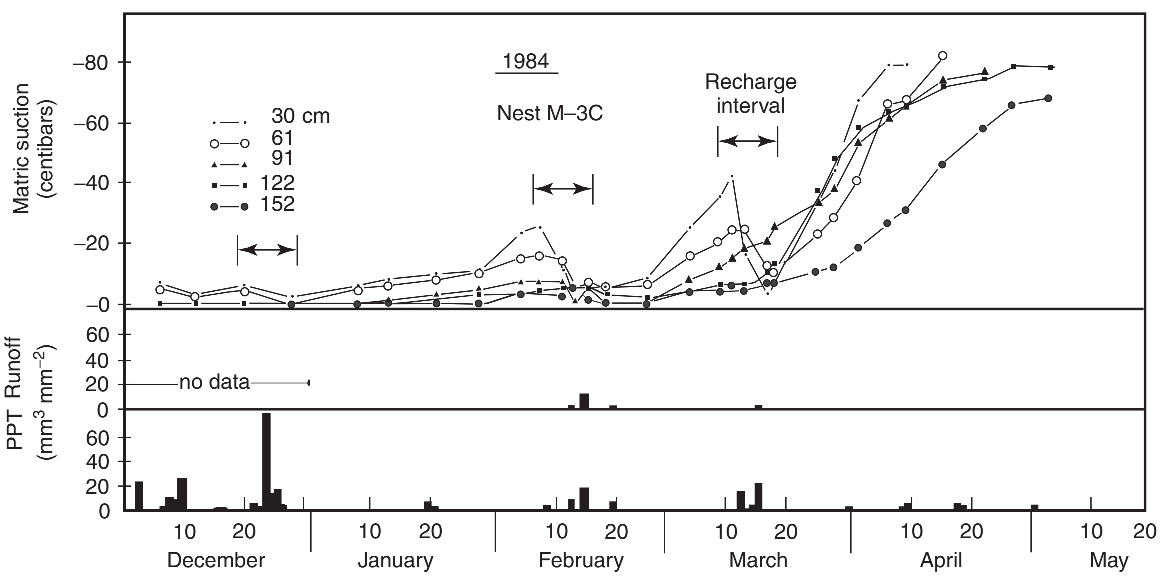

Figure 1 Measured precipitation (PPT), runoff, and matric pressure at several depths for a five-month period on a marine terrace (Reprinted from Ground Water, the National Ground Water Association. Copyright 1985)

zone. The process is thus affected by water-table variation, whether caused by natural processes or artificial recharge or withdrawal.

The character of the subsurface is also a major influence on recharge. Fractures, preferential flow, karst features, and so on, can lead to enhanced recharge by increasing spatial concentration. Features that act to hold water near the surface tend to increase evapotranspiration, and decrease recharge. For example, stratification of the unsaturated zone retards downward flow. Soil water hysteresis also acts during redistribution of water in the unsaturated zone to hold water high in the profile; therefore soils with more pronounced hysteresis of soil water retention may have greater evapotranspiration and less recharge (*see* **Chapter 150, Unsaturated Zone Flow Processes, Volume 4**). The medium also affects the thickness of the zone most affected by evapotranspiration; the presence of bedrock at shallow depth, for example, may confine roots to a thin zone and allow more water to escape their influence by flowing into the bedrock.

Recharge Distribution and Applications

The importance of the variability of recharge in time and space depends on the nature of the application as well as the processes. Water supply issues related to depletion of an aquifer as a whole usually make use of averages over a large area, generally an entire watershed or aquifer. Usually this type of application also requires a time average suitable for predicting long-term sustainability. Thus the period of averaging may include centuries since the last major climatic change. Issues of contaminant transport from a point source usually need a point or highly localized estimate of recharge rate, for a time period about as long as the contaminant would remain present in significant quantities. Aquifer vulnerability assessments and diffuse-source contamination (e.g. from agricultural pesticides) often require knowledge of the distribution of recharge rates over a specified area with high spatial resolution.

Certain time and space scales are inherent in the relevant conceptual models to be chosen or developed. For example, with a deep unsaturated zone in a granular medium, virtually all temporal fluctuations may be averaged out before water enters the saturated zone, making a decadal or longer scale appropriate. If there is a predominance of fracture flow or a shallow unsaturated zone, however, the conceptual model might need to allow for variations on a daily basis.

Methods of estimating recharge produce estimates pertaining to different time and space scales. Many methods fundamentally assess recharge in a way that is highly localized or specific in time and space while others produce recharge estimates that are temporally or areally averaged. For point applications, areal estimates will generally not be suitable unless spatial variability is slight, or unless it is possible to delineate features on the ground that might permit evaluation of likely deviations from the areal average. Additional data relating to the finer scale would be needed to estimate recharge at a given point. For areal applications when the available recharge estimates are point estimates, the area of interest can be divided into subareas within which recharge is more likely to be uniform, and an integrated estimate may be produced using point data in each subarea. Thus, for a given application, a good conceptual model of the hydrologic system and its recharge processes is essential for selecting recharge estimation methods, and for integrating point measurements to represent an area.

ESTIMATION OF RECHARGE

Because of the extreme variability of recharge rates and the fact that many relevant processes cannot be directly observed, recharge is usually difficult to estimate, especially when it involves unsaturated flow. There is no single instrument with the simplicity of a thermometer or pressure gauge that measures unsaturated-zone flux, so estimates are normally based on a combination of various types of data. Particular difficulties in the unsaturated zone are the extreme nonlinearity of basic unsaturated hydraulic properties, the typically extreme heterogeneity of these properties and other relevant features of the unsaturated zone, and the frequent occurrence of preferential flow (*see* **Chapter 150, Unsaturated Zone Flow Processes, Volume 4**). In addition, important processes that compete for water in the unsaturated zone, such as evapotranspiration, often are also difficult to evaluate. A quantitative estimate of recharge may depend on a combination of data that themselves have substantial uncertainty. For these reasons it is wise to apply multiple methods and compare their results.

Much software for simulating surface water, groundwater, and unsaturated-zone transport has been developed that is useful in investigations of aquifer recharge. Public-domain software related to various methods described in this article is available at `http://water.usgs.gov/software/index.html`.

Water-budget Methods

Water-budget methods comprise the largest class of techniques for estimating recharge. The water budget is an integral component of any conceptual model of the system under study. Hence, analysis of the water budget is of great value in any recharge study. Simple water budgets can be readily constructed and refined as needed. As noted by Lerner *et al.* [1990], a good method for estimating recharge should explicitly account for all water that does not become recharge.

The water balance for a control volume, such as an aquifer or a watershed, can be stated as (Scanlon *et al.*, 2002)

$$P + Q_{on}^{sw} + Q_{on}^{gw} = ET^{sw} + ET^{uz} + ET^{gw} + Q_{off}^{sw} + Q_{off}^{gw} + Q_{bf} + \Delta S^{snow} + \Delta S^{sw} + \Delta S^{uz} + \Delta S^{gw} \quad (1)$$

where superscripts refer to surface water, groundwater, unsaturated zone, or snow, P is precipitation and irrigation; Q_{on} and Q_{off} are water flow into and out of the basin; Q_{off}^{sw} is surface-water runoff; Q_{bf} is baseflow (groundwater discharge to streams or springs); ET is evapotranspiration; and ΔS is change in water storage. Units may be any of those presented earlier for recharge. Following our definition, as water that enters the saturated zone, recharge can be written as (Schicht and Walton, 1961):

$$R = \Delta Q^{gw} + Q_{bf} + ET^{gw} + \Delta S^{gw} \quad (2)$$

where R is recharge and ΔQ^{gw} is the difference between groundwater flow out of and into the basin. This equation simply states that water arriving at the water table (i) flows out of the basin as groundwater flow, (ii) discharges to the surface, (iii) is lost to evapotranspiration, or (iv) remains in the subsurface, augmenting storage. Substitution into (1) results in another useful form of the water balance:

$$R = P + Q_{on}^{sw} - Q_{off}^{sw} - ET^{sw} - ET^{uz} - \Delta S^{snow} - \Delta S^{sw} - \Delta S^{uz} \quad (3)$$

Water-budget methods include all techniques based, in one form or another, on one of these water balance equations.

Applications of various forms of equations 1 to 3 abound in the literature. An attractive feature of these methods is their flexibility; applications are often simplistic, but can be quite complex. Derivation of equations 1 to 3 requires few assumptions on the mechanisms that control individual components. Hence, through careful adaptation the equations can be applied over a wide range of space and timescales. For example, the water budget equation can be applied to study water movement in a soil lysimeter at scales of centimeters and seconds. At the same time, the water-budget equation is an integral part of Global Climate Models (GCMs) that predict global climate changes over periods of centuries.

The most common way of obtaining an estimate of recharge by these methods is the indirect or "residual" approach whereby all of the variables in the water budget equation, except R, are either measured or estimated and R is set equal to the residual. The accuracy of the recharge estimate is then dependent on the accuracy with which the other water-budget components can be measured or estimated. Consider a simple example where the only significant components of equation 3 are recharge, precipitation, and unsaturated zone ET and ΔS. Using the residual approach, we can write:

$$(R + \varepsilon_R) = (P + \varepsilon_P) - (ET^{uz} + \varepsilon_{ET}) - (\Delta S^{uz} + \varepsilon_S) \quad (3a)$$

Where the ε terms represent measurement errors. The magnitude of the error in the recharge estimate, ε_R, could be as large as $|\varepsilon_P + \varepsilon_{ET} + \varepsilon_S|$. In arid settings where R is generally a small fraction of P or ET, $|\varepsilon_R|$ can exceed the true value of R, even if the other error terms are relatively small. This can be a serious limitation on application of the residual approach. Timescales for application of water-budget methods are important; more frequent tabulations are likely to improve accuracy. Averaging over longer time periods tends to dampen out extreme precipitation events and hence underestimate recharge. For example, in an arid region, potential ET averaged over a week or a month will usually greatly exceed average P over the same period. But on a daily basis, P can exceed potential ET – a necessity for recharge to occur. Narayanpethkar *et al.* (1994) used a water-budget method to derive an estimate of 23 mm yr^{-1} of recharge in a region of India; Steenhuis *et al.* (1985) estimated recharge to be 400 mm at a site in the eastern US.

Direct water-budget approaches are based on measurements of changes in water storage within some compartment (such as a soil column) and partitioning of those changes to recharge and possibly to other components such as evapotranspiration. Included in this class are the zero-flux plane method and the use of lysimeters, both discussed in a subsequent section.

Modeling Methods

Hydrologic models constitute an important class of water-budget methods. The commonly used groundwater and surface-water flow equations are, in fact, variations of equations 1 to 3. Several modeling approaches have been used to estimate recharge.

Watershed models have been applied at a variety of spatial scales. Some of these models provide a single, lumped value of recharge for the watershed. Others allow disaggregation of the watershed into distinct hydrologic zones that may receive varying amounts of recharge. Watershed models generally require daily climatic data and information on land-surface features and land use. Streamflow data, and less frequently groundwater data, are used to calibrate the models. Bauer and Mastin (1997) used a watershed model to estimate recharge in three small watersheds (average drainage area 0.4 km^2) in western Washington. On the other end of the spatial scale, Arnold *et al.* (2000) applied the SWAT model to the entire upper Mississippi River Basin (492 000 km^2). Recharge estimates in different parts of that basin were between 10 and 400 mm yr^{-1}. Figure 2

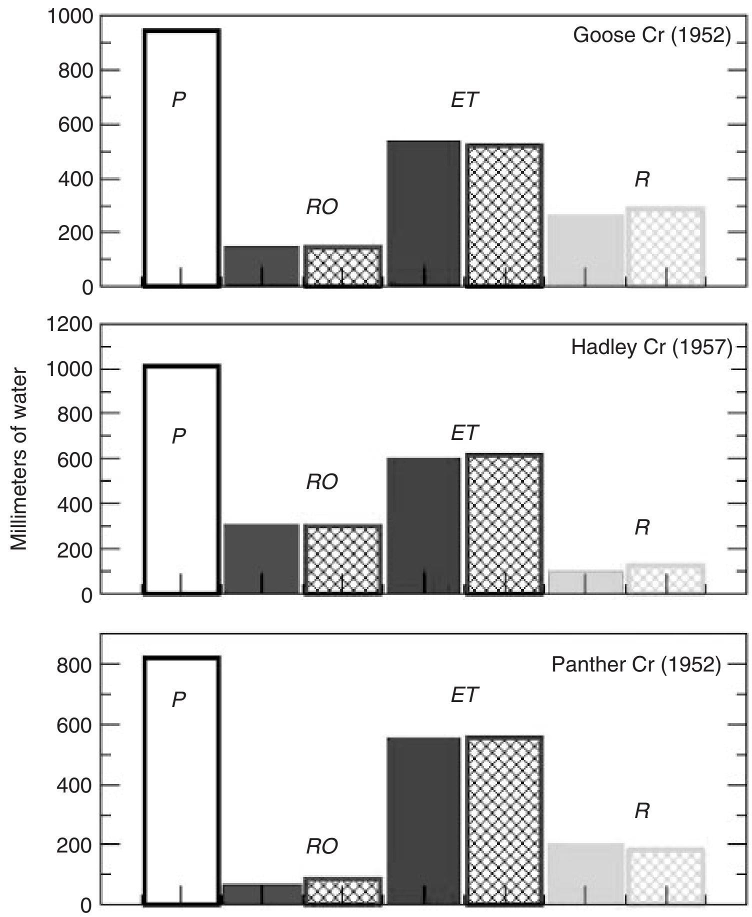

Figure 2 Water-budget components for three watersheds in Illinois: actual (Schicht and Walton, 1961), shown as solid bars; and predicted with the SWAT model (Arnold and Allen, 1996), shown as hatched bars. A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

shows one-year water budgets, including recharge, for 3 small streams in Illinois as measured by Schicht and Walton (1961) and as predicted by Arnold and Allen (1996) using the SWAT model.

Recharge is an integral component of groundwater flow models, and estimates of recharge can be generated through model calibration. Recent incorporation of automatic inverse modeling techniques (Hill *et al.*, 2000) to these models has facilitated this process. A concern with these inverse techniques, however, is the possibility that they could generate nonunique solutions (i.e. different parameter values could produce identical results). This situation can be avoided by calibrating with data on both groundwater head and flux, or by independent determination of hydraulic conductivity. Tiedeman *et al.* (1998) developed estimates of recharge for the Albuquerque Basin in New Mexico using this approach.

One-dimensional models of vertical flow through the unsaturated zone have also been used to estimate recharge rates. As with watershed models, daily climate data are usually used. Information on hydraulic properties of sediments within the unsaturated zone is also required. The complexity of these models varies widely, from a simple reservoir (bucket) model to analytical or numerical solutions to Richards' equation. Example applications include those of Scanlon and Milly (1994), Kearns and Hendrickx (1998), and Flint *et al.* (2002).

A powerful feature of all models is their predictive capability. They can be used to gauge the effects of future climate or land-use changes on recharge rates. Sensitivity analyses can also be used to identify model parameters that most affect computed recharge rates (*see* **Chapter 155, Numerical Models of Groundwater Flow and Transport, Volume 4** and **Chapter 156, Inverse Methods for Parameter Estimations, Volume 4**).

Empirical equations, such as the Maxey-Eakin equation (Maxey and Eakin, 1949), are often used to obtain quick and inexpensive estimates of recharge. These equations, usually developed for specific regions, relate recharge to more easily measured phenomena such as precipitation, temperature, elevation, vegetation, and soil type. While the accuracy of these equations may not be sufficient for some studies, the simplicity of the approach facilitates application over large areas using a Geographic Information System (GIS) and remotely sensed data.

Methods Based on Surface-water Data

Several methods are available for estimating recharge from data collected in surface-water bodies. Most of these methods require data on stream discharge at multiple points or times. Stream loss or seepage to the subsurface can be determined indirectly as the difference in discharge between a downstream and an upstream measurement point. This method is referred to as the Channel Water Balance Method and is appropriate for application only on losing streams. Stream or transmission loss does not necessarily equate with recharge. Water lost from a stream could remain as bank storage, be consumed by evapotranspiration, or become recharge. Stream loss falls into the category termed "potential" recharge by Lerner *et al.* (1990). Seepage estimates obtained by this method represent a value that is integrated over the length of stream channel between the two measurement points. Surface/groundwater exchange at specific points in a surface-water body can be directly measured with seepage meters (Kraatz, 1977; Lee and Cherry, 1978).

Stream flow hydrographs for gaining streams can be analyzed to estimate what portion of that discharge can be attributed to baseflow (also known as *groundwater discharge*). The rate of groundwater discharge is closely related to the rate of recharge; however, the two are not necessarily equal (equation 2). It is unlikely that all recharge in a basin would eventually be discharged to streams. Recharge may be lost to evapotranspiration, pumpage, or groundwater flow out of the basin. Care should be taken in understanding runoff processes in any given watershed. Some basins are unsuitable for hydrograph analysis. In particular, basins with upstream regulation, significant groundwater withdrawals, topographically flat terrain, karstification, or losing stream reaches should be avoided. Hydrograph analysis is usually performed for relatively humid regions where streams are perennial, with well-sustained base flow. Analytical techniques include hydrograph separation (Sloto and Crouse, 1996), chemical hydrograph separation (Hooper *et al.*, 1990), recession-curve displacement (Rorabaugh, 1964; Rutledge, 1998), and analysis of flow-duration curves (Kuniansky, 1989). Using an automated hydrograph recession-curve analysis computer program, Rutledge and Mesko (1996) analyzed discharge records from 89 basins in the eastern US and obtained recharge estimates between 152 and 1270 mm per year. Daniel and Harned (1998) analyzed 161 water years of record from 16 surface-water stations with the local minimum method (Pettyjohn and Henning, 1979) and recession-curve displacement and found that the former method provided estimates that were on average 21% less than those of the recession-curve displacement method.

Methods Based on Groundwater Data

Information on groundwater levels and how those levels vary in time and space can be used to estimate recharge. The most widely used of these methods is the water-table fluctuation (WTF) method (Healy and Cook, 2002). Other methods in this group include variations of the Darcy Methods (as discussed below), flow net analysis (Cedergren, 1977), and various semiempirical modeling approaches (e.g. Su, 1994; Wu *et al.*, 1997).

The WTF method uses the following equation to estimate recharge to unconfined aquifers:

$$R = S_y \frac{\mathrm{d}h}{\mathrm{d}t} \approx S_y \frac{\Delta h}{\Delta t} \tag{4}$$

where S_y is specific yield, h is water-table height, $\mathrm{d}h/\mathrm{d}t$ is the derivative of water-table height with respect to time, which is approximated as the difference in h (Δh) over the elapsed time between measurements (Δt). Equation 4 is applied only over periods of water-level rise (i.e. Δh is positive). The method works best over short periods for shallow water tables that display sharp water-level rises and declines. The method inherently assumes that recharge occurs only as a result of transient events; recharge occurring under steady flow conditions cannot be estimated.

Despite its simplicity and an abundance of available data, there are difficulties in application of the WTF method (Healy and Cook, 2002). Phenomena other than recharge can induce fluctuations in the water table. These include evapotranspiration, changes in atmospheric pressure, the presence of entrapped air ahead of a wetting front, extraction or injection of water by pumping, temperature effects, and tidal effects (Todd, 1989). Careful examination of water level records in conjunction with climatic data is necessary in order to correctly identify water level rises that can be attributed to recharge events. Determining values for S_y can be problematic. There are many methods for measuring or estimating values of S_y (e.g. laboratory methods, field methods, water-budget methods, and empirical methods), but all of these have a degree of associated uncertainty. Rasmussen and Andreasen (1959) applied this method to a small basin in the eastern US and determined that over a 2-year period annual recharge averaged 541 mm. Figure 3 shows average weekly water levels and precipitation for that study. The dashed lines in Figure 3 are the extrapolated recession curves (traces that the well hydrograph would have followed in the absence of the rise-producing precipitation). Δh is measured as the difference between a hydrograph peak and the recession curve at the time of that peak.

The zero-flux plane (ZFP) is the plane within the subsurface where the vertical hydraulic gradient is zero;

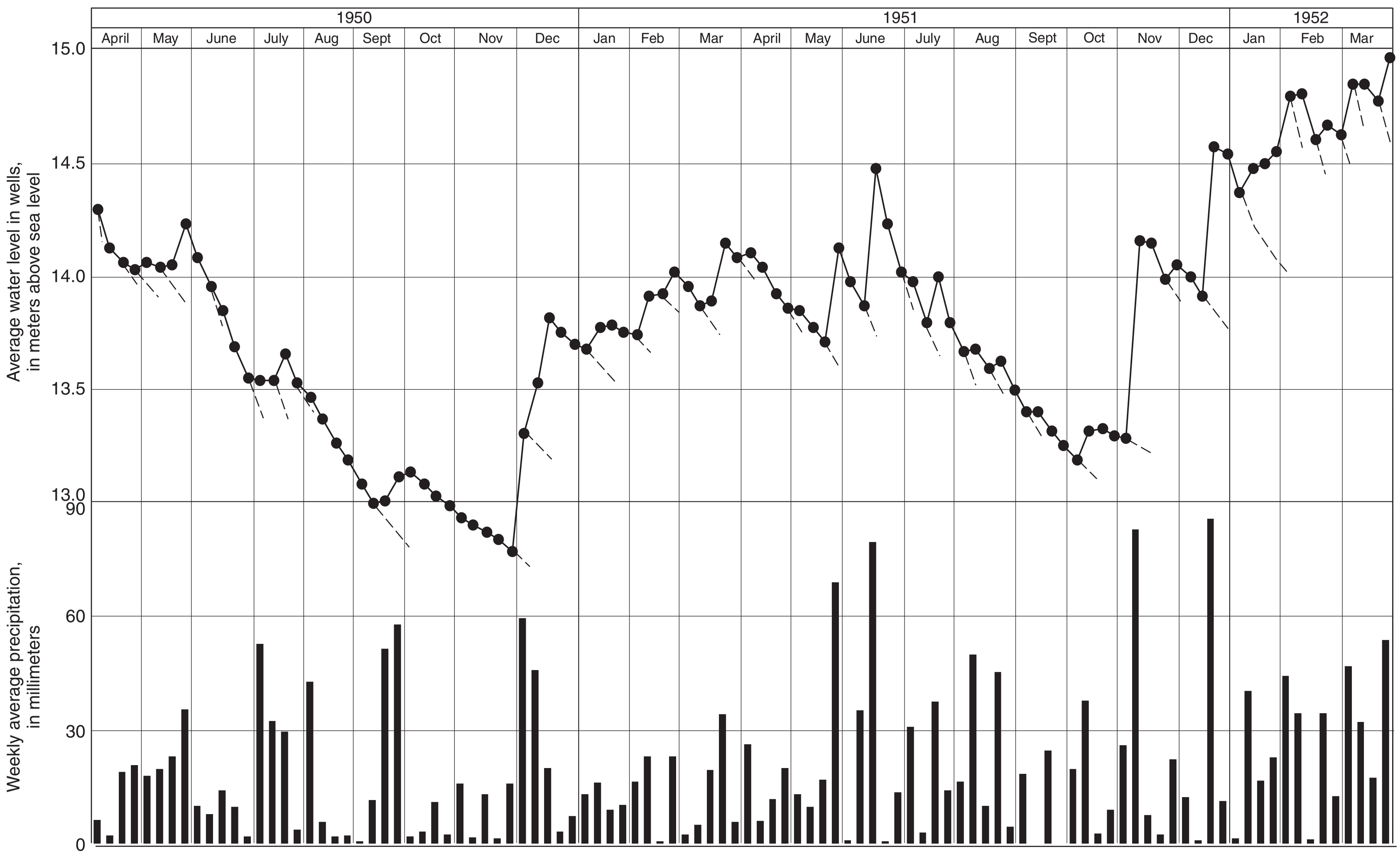

Figure 3 Average water levels in wells and weekly precipitation for Beaverdam Creek basin study of Rasmussen and Andreasen (1959), cited by Healy and Cook (2002)

that is, the plane marks the boundary between water that is moving upwards in response to evapotranspiration and water that is draining downward to become recharge. In applying this method, recharge rate is equated with the rate of change in water storage (drainage) beneath the ZFP (Richards *et al.*, 1956; Roman *et al.*, 1996). Data requirements for this method may be too onerous for some studies. Frequent measurements of soil matric pressure are needed at several depths to accurately locate the ZFP. Soil moisture content must also be measured over the entire thickness of the unsaturated zone. The method cannot be used when water is moving downward throughout the entire profile (i.e. when the ZFP is at land surface). Sharma *et al.* (1991) applied the ZFP method at eight sites in a semiarid region of Western Australia; estimated recharges rates ranged from 34 to 149 mm yr^{-1}.

Lysimeters are containers filled with soil and possibly vegetation and are placed in an environment that mimics natural conditions (Brutsaert, 1982; Young *et al.*, 1996). Fluxes to and from lysimeters are monitored with drains and water-collection vessels or balances that measure changes in weight. The instruments are usually used to measure *ET*. However, if the base of a lysimeter is below the ZFP, drainage from it can be assumed equal to recharge. Kitching *et al.* (1977) used a lysimeter with a surface area of 100 m^2 to measure recharge rates of 342 to 478 mm yr^{-1} for Bunter Sandstone in England.

Darcian Methods

Applied in the unsaturated zone, Darcy's law gives a flux density (q) equal to hydraulic conductivity (K) times the driving force, which equals the recharge rate if certain conditions apply. The basic requirements of the method are to determine K and the driving force, and then to calculate the downward flux density from Darcy's law in a form such as

$$q = -K(\theta)\left(\frac{d\psi}{dz} + \rho g\right) \tag{5}$$

where θ is the water content, ψ is the matric pressure, z is the vertical coordinate, ρ is the density of water, and g is the gravitational acceleration. K would have SI units of $m^2\,Pa^{-1}\,s^{-1}$. Another common form of equation (5), with ψ in head units, has the gravitational term equal to unity. **Chapter 150, Unsaturated Zone Flow Processes, Volume 4** gives additional information about equation (5) and its variables. To use (5) in estimating recharge, accurate measurements are necessary to know $K(\theta)$ adequately under field conditions at the point of interest. For the total driving force, gravity being known, it is essential to determine the matric-pressure gradient or to demonstrate that it is negligible. For purposes requiring areal rather than point estimates, additional interpretation and calculation are necessary. Even when properly applied, Darcian methods do not necessarily indicate total recharge; some types of preferential flow are inherently non-Darcian and if significant would need to be determined separately.

In the simplest cases, in a region of constant downward flow in a deep unsaturated zone as illustrated in Figure 4(a), gravity alone drives the flow. With a sample from the zone of uniform ψ, K measurements at the original field water content directly indicate the long-term average recharge rate. The depth required for flow to be steady depends on the climate, medium, and vegetation. Fluxes

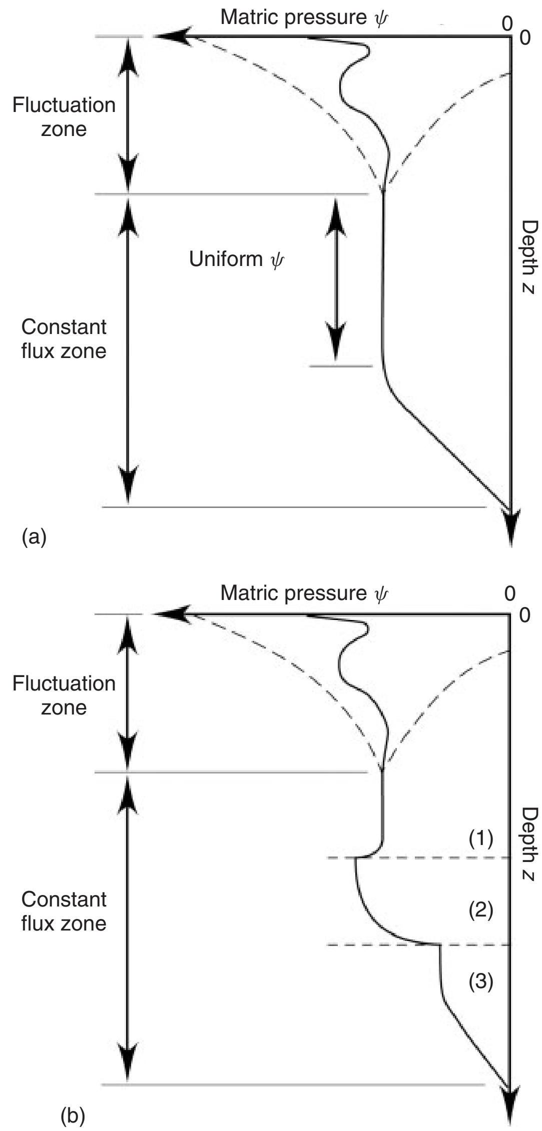

Figure 4 Hypothetical profiles of matric pressure as a function of depth in an unsaturated zone deep enough that its lower portion has a constant downward flux of water in (a) a uniform and (b) a layered profile. Dashed curves indicate possible extremes in the upper portion. The lower horizontal line in each diagram indicates the position of the water table

and water contents fluctuate within the root zone, and possibly to considerably greater depths. Evidence from field experiments (e.g. Nixon and Lawless, 1960) and theoretical assessment of the damping of moisture fluctuations as they move downward (Gardner, 1964) suggest that depths of a few meters are often adequate. In some cases, however, moisture fluctuations have been detected at depths of tens of meters (e.g. Jones, 1978). An obvious way to evaluate steadiness is to measure water content or matric pressure in the zone of interest for a period of a year or more. Where that is not possible, the evaluation of such factors as the homogeneity of the unsaturated zone and the distribution of water within it may help to indicate the degree of steadiness with depth.

In early applications of this technique (Enfield *et al.*, 1973; Stephens and Knowlton, 1986), limited accuracy of K estimates was a significant problem. Accurate laboratory K measurements on core samples, as by the steady-state centrifuge method, are of great value (Nimmo *et al.*, 1994). Figure 5 shows an example of measured hydraulic conductivities with the field water content and its associated K value identified. It happens for sample M30 that two measured $K(\theta)$ points have θ nearly equal to its field value, but even where this is not the case, having measurements at values both higher and lower than the field θ permits the appropriate K to be ascertained by interpolation.

In the examples of Figure 5, as is common, the $K(\theta)$ relation is very steep in the relevant range, meaning that uncertainty in the field θ leads to a greatly magnified uncertainty in the inferred recharge rate. To have reasonable confidence in the recharge estimate, accurate knowledge of the *in situ* water content of the sampled material is essential. In the cases illustrated, it was necessary to work with the water content of the identical soil mass just after sampling, corrected for evaporation between sampling and the first weighing. The oven-dry weight of that same soil must be known, here obtained by drying after K measurements were completed and corrected for losses of dry matter during experimental operations.

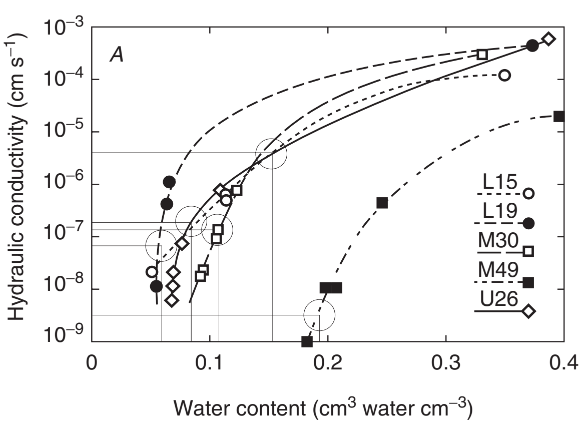

Figure 5 Steady-state centrifuge measurements of hydraulic conductivity versus water content, for five core samples from the unsaturated zone below a desert wash. To indicate recharge flux, the original field water content of each sample is indicated with its corresponding hydraulic conductivity (Reproduced from (Nimmo *et al.*, 2002) by permission of American Geophysical Union)

Variations of the steady-state method are possible for some cases that fall short of the ideal of constant and uniform flow and moisture conditions within a finite portion of the unsaturated zone. One important case is steady flow in a layered medium, in which θ and matric pressure vary spatially as necessary to give K the value needed at each point to maintain a constant Darcian flux density (Figure 4b). Nimmo *et al.* (2002) demonstrated a method for handling this problem using spatially detailed estimates of unsaturated hydraulic properties to compute the uncertainty ascribable to inadequate knowledge of the actual matric-pressure gradient. In such a case one can use geologic data, to the extent possible, to indicate the spatial variability of hydraulic properties.

A more challenging problem is the case where downward fluxes are not steady in time. Some situations may be treated as sometimes-steady, especially if there is evidence that conditions are nearly constant for an extended period such as a growing season. In others, the deviations may be small enough in magnitude that the error from assuming true steadiness is tolerable. It must be acknowledged however, that these approximations compromise not just the magnitude of uncertainty but also the possibility of claiming the inferred recharge rate as a long-term average.

For the general case in which flow is not steady, the appropriate Darcian methods are more complex. Transient water contents and matric pressures must be measured in addition to K (Freeze and Banner, 1970; Healy, 1989). The field site needs to be instrumented with devices for these measurements (*see* **Chapter 150, Unsaturated Zone Flow Processes, Volume 4**). With knowledge of K and the gradient of matric pressure, Darcy's law (equation 5) straightforwardly indicates the flux density at a given time and position, and its downward component is interpretable as recharge if the setting and conditions are such that fates other than recharge can be ruled out. Transient recharge computed with Darcy's law can relate to storms or other short-term events, or provide data for integration into temporal averages. Considerable equipment and labor is required for this, however. It would not usually indicate the long-term average recharge rate accurately, but it can have a valuable payoff in giving a detailed picture of the behavior of water in the unsaturated zone.

Chemical Tracer Methods

Geochemical tracers that have been used for recharge estimation include tritium (^{3}H), oxygen-18 (^{18}O), and deuterium (^{2}H), which are constituents of the water molecule

(H_2O); naturally occurring anions such as chloride (Cl^-) and bromide (Br^-); agriculturally introduced chemicals including nitrate ($NO_3{}^-$); applied organic dyes such as fluorescein ($C_{20}H_{12}O_5$); and dissolved gases including chlorofluorocarbons (CFCs), sulfur hexafluoride (SF_6), and noble gases such as helium (He) and argon (Ar). Concentrations of these constituents in pore water are related to recharge by applying chemical mass-balance equations, by taking advantage of distinctive temporal patterns in the composition of infiltrating water, or by determining the "age" of the water (i.e. the time since the water was in contact with the atmosphere). Geochemical tracers provide point and areal estimates of recharge. Of the many geochemical tracers that have been used for research purposes, only a few have found widespread use at scales relevant to aquifer recharge. They are the main subjects of this section.

The chemical mass-balance approach assumes that the tracer is conservative and moves with pore water, without significant retardation or acceleration due to electrochemical interactions with solids. Chemical mass-balance methods can be applied at basin scales using only groundwater and atmospheric-deposition data, (e.g. Dettinger, 1989; Anderholm, 2000). When applied to the unsaturated zone, chemical mass-balance estimates are obtained by equating the time-averaged flux of tracer across the land surface with the tracer flux at a depth sufficient to be unaffected by evapotranspiration and other temporally varying influences:

$$q_{m,z_0} = C_z q_z \tag{6}$$

Here q_{m,z_0} is the average flux of tracer across the land surface (mass of tracer per unit land surface per unit time), C_z is the tracer concentration in pore water at the evaluation depth z (mass of tracer per unit volume of pore water), and q_z is the water-flux density at z (volume of water per unit bulk area normal to the flux direction per unit time). As with the Darcian method (equation 5), q_z equals the recharge rate if certain conditions apply.

Soluble tracers reach the land surface in atmospheric deposition of dust (dry fall) and in precipitation. Infiltration of runoff (streamflow and sheetwash) supplies additional tracer where overland flow occurs. Irrigation water, agricultural chemicals, and engineered structures provide additional tracer to cultivated areas and areas of artificial recharge. The total flux of tracer across the land surface is the sum of the fluxes from constituent sources, that is

$$q_{m,z_0} = C_p P + C_r R + C_i I + C_f F + C_a A \tag{7}$$

where C_p is the effective tracer concentration in precipitation (i.e. including tracer in dry deposition), C_r is the concentration of tracer in infiltrating runoff, C_i is the concentration of tracer in irrigation water, C_f is the concentration in fertilizer and other agricultural applications, C_a is the concentration in artificial-recharge water, and P, R, I, F, and A are the volumes of precipitation, runoff, irrigation, agricultural applications, and artificial-recharge water that infiltrate per unit time per unit land surface. The first term on the right-hand side represents atmospheric sources, the second term overland-flow sources, and the last three terms anthropogenic sources. One or two terms usually dominate in a given environment.

Substituting equation (7) into equation (6) and rearranging yields the water-flux density at depth z (q_z) in terms of measured quantities, that is.

$$q_z = \frac{(C_p P + C_r R + C_i I + C_f F + C_a A)}{C_z} \tag{8}$$

If z is in a zone of steady flux, that is, if climatic and other inputs are approximately stationary over relevant timescales (dictated by travel times between the land surface and the aquifer), then q_z is equal to the recharge rate at the sampled location. Application of equation (8) also assumes negligible preferential flow (or sufficient sampling to provide adequate averaging of preferential pathways), in addition to conservative transport (Wood, 1999; Scanlon, 2000). Collectively, these assumptions imply that tracer fluxes at depth are in approximate steady-state equilibrium with the time-averaged tracer fluxes across the land surface.

Recharge rates can also be determined from apparent travel velocities of a tracer (solute) "marker" through the unsaturated zone, using (Saxena and Dressie, 1984; Gvirtzman and Magaritz, 1986; Williams and Rodoni, 1997)

$$q_z = \bar{\theta}\frac{(z_2 - z_1)}{(t_2 - t_1)} \tag{9}$$

where $\bar{\theta}$ is the average volume fraction of water between z_1 and z_2. Here z_1 and z_2 are the depths of a solute marker at times t_1 and t_2. Like the mass-balance method (equation 8), the tracer-velocity method requires that the marker move conservatively with the water and that water contents at depth remain constant through time.

The most widely used geochemical tracer for recharge estimation is chloride. Chloride is abundant in nature, conservative in hydrologic settings, and readily analyzed. Chloride mass-balance methods have proven especially useful in arid and semiarid regions (Stone, 1984; Edmunds *et al.*, 1988; Allison *et al.*, 1994; Phillips, 1994; Prudic, 1994; Tyler *et al.*, 1996; Roark and Healy, 1998; Maurer and Thodal, 2000; Izbicki *et al.*, 2002; Stonestrom *et al.*, 2003b). Figure 6 (adapted from Stonestrom *et al.*, 2003b) shows an example of a conservative tracer (chloride) applied in three distinct hydrologic settings in an arid environment: beneath native xeric vegetation, beneath an ephemeral stream, and beneath an irrigated crop. Results in Figure 6 are plotted in integral form (as cumulative chloride versus cumulative water, starting at the land surface).

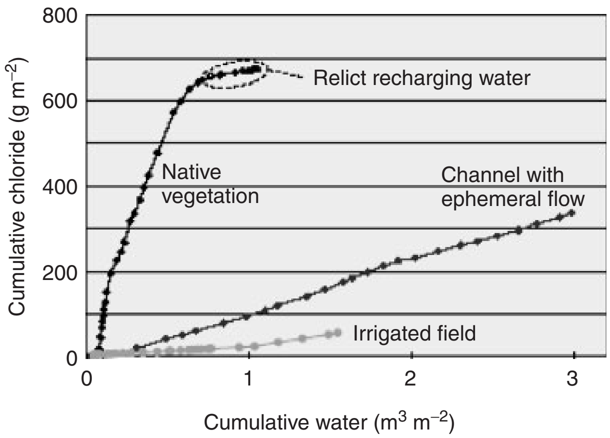

Figure 6 Concentrations of chloride in pore water, given by the slopes of the curves, indicate aquifer recharge for quasi-steady conditions. Recharge beneath irrigated field is higher than beneath ephemeral channel. Profile beneath native vegetation has multiple slopes reflecting climate-induced changes in recharge. Example from the Amargosa Desert, Nevada (Stonestrom *et al.*, 2003b). A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

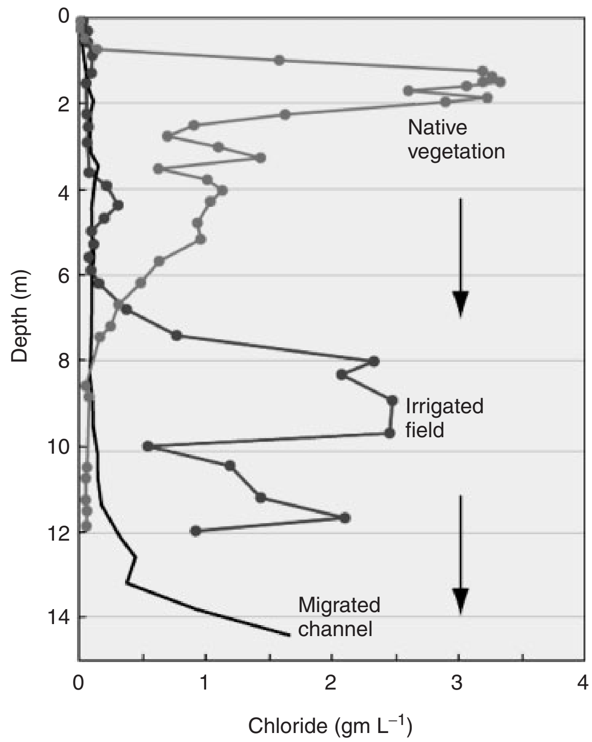

Figure 7 Travel-time of a solute marker indicates aquifer recharge under certain conditions. Here accumulations of chloride, common beneath native vegetation in arid and semiarid environments, served as the marker. Recharge following conversion to agriculture mobilized the marker in one case. Recharge following flood-induced channel migration mobilized the marker in another. Example from the Amargosa Desert, Nevada (Stonestrom *et al.*, 2003b). A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

Plotted this way, linear relations that extrapolate through the origin are consistent with the assumption that transport of water and solute at depth are in approximate equilibrium with average net inputs across the land surface. Slopes of the relations (equal to pore-water concentrations) are inversely proportional to recharge rates. Profiles beneath the channel and irrigated field meet the criteria for application of the method; the profile beneath native xeric vegetation does not. Water in the deep unsaturated zone here is relict infiltration from a prior climatic period. Figure 7 (adapted from Stonestrom *et al.*, 2004) shows displaced chloride "markers" used in the tracer-velocity method (equation 9).

Distinct isotopic "species" of water also serve as tracers for determining recharge. The isotopic composition of precipitation varies with altitude, season, storm track, and other factors. Recharge estimates using distinct isotopic species of water employ temporal or geographic trends in infiltrating water (Saxena and Dressie, 1984; Gvirtzman and Magaritz, 1986; Williams and Rodoni, 1997). Isotopically distinct water can be used in mass-balance or tracer-velocity methods (equations 8 and 9). Deuterium and oxygen-18 labeled species are conservative tracers. Mass-balance methods can be applied to tritium-labeled species taking radioactive decay into account.

Tritium and other nonconservative tracers can indicate the length of time that water has been isolated from the atmosphere, that is, its "age". In principle equivalent to tracer-velocity methods, age-dating methods are usually applied to the saturated zone. Recharge rates can be inferred from groundwater ages if mixing is small. If ages are known along a flow line, the recharge rate is given by

$$q = \bar{\theta} L (A_2 - A_1)^{-1} \qquad (10)$$

where $\bar{\theta}$ is the average volumetric water content along the flow line, A_1 and A_2 are the ages at two points, and L is the separation length. One point is often located at the water table. Preindustrial water can be dated by decay of naturally occurring radioisotopes, including carbon-14 and chlorine-36 (Leaney and Allison, 1986; Phillips *et al.*, 1986). The abundance of tritium, chlorine-36, and other radioisotopes increased greatly during atmospheric weapons testing, labeling precipitation starting in the 1950s (Knott and Olimpio, 1986). Additional anthropogenic tracers for dating postindustrial recharge include chlorofluorocarbons, krypton-85, SF_6 and agricultural chemicals (Ekwurzel *et al.*, 1994; Davisson and Criss, 1996; Busenberg and Plummer, 2000).

Heat-based Methods

Heat is a useful tracer for estimating aquifer recharge. Perturbations of purely conductive propagation of temperature fluctuations from the land surface into the subsurface indicate recharge rates in shallow profiles beneath streams and other sources of water (Rorabaugh, 1954; Lapham, 1989; Stonestrom and Constantz, 2003). Water that moves below the reach of seasonal temperature oscillations perturbs the geothermal temperature distribution at depth. The degree of perturbation thus indicates recharge rates in deep unsaturated zones (Rousseau *et al.*, 1999). Measured temperature profiles are used in analytical or numerical inversions of the equation(s) governing the coupled transport of heat and water, solving for water flux at the water table. Numerical inversions can treat sediment heterogeneities as well as arbitrary initial and boundary conditions.

Daily or annual temperature fluctuations can often be approximated by sinusoidal functions (Figure 8). If recharging fluxes are steady and materials homogeneous, the

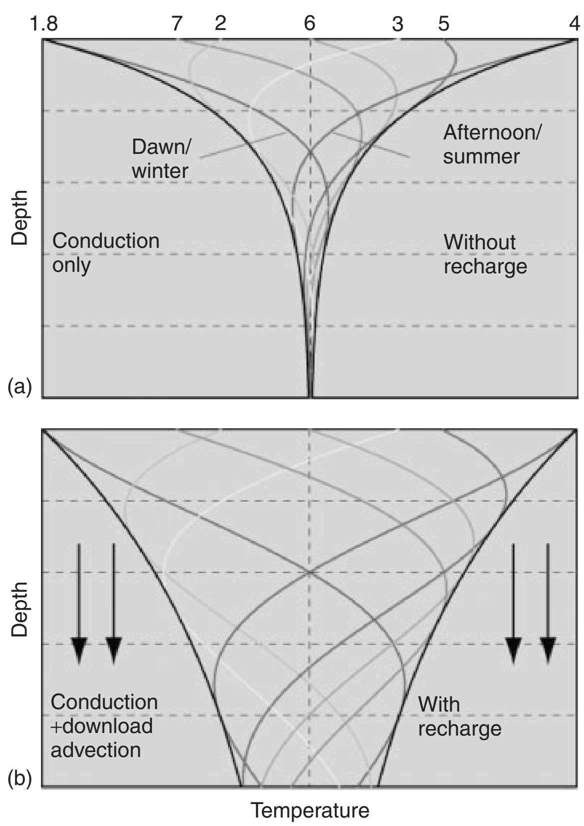

Figure 8 Recharging water advects diurnal and seasonal temperature fluctuations deeper into the profile than conduction alone. Numbered lines show successive temperature profiles over one daily or annual cycle (profiles 1 and 8 are identical, but separated by one cycle). The amount of downward shift in the bounding envelope depends on the rate of aquifer recharge. A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

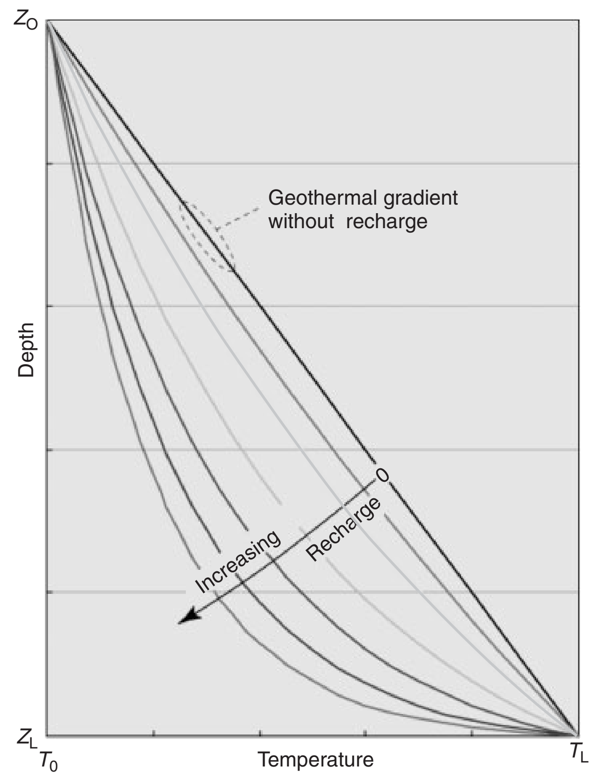

Figure 9 Recharging water in the deep unsaturated zone perturbs the steady geothermal profile produced by conduction alone. The degree of departure from the purely conductive profile indicates the amount of recharge. T_0 and T_L are temperatures at depths, Z_0 and Z_L, beneath the maximum penetration of seasonal fluctuations. A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

recharge rate q can be obtained by inversion from the phase shift (b) and attenuation (a) of the temperature waves as they propagate downward from the land surface (Stallman, 1965)

$$T(z) = T_0 + \Delta T \exp(-az) \sin\left(\frac{2\pi}{\tau - bz}\right) \tag{11}$$

where a and b are related to thermal-conduction and advection constants K' and V' by

$$a = \left[\left(\frac{K' + V'^4}{4}\right)^{\frac{1}{2}} + \frac{V'^2}{2}\right]^{1/2} - V'$$

and

$$b = \left[\left(\frac{K' + V'^4}{4}\right)^{\frac{1}{2}} - \frac{V'^2}{2}\right]^{1/2}$$

Here T_0 and ΔT are the mean and amplitude of the temperature signal at the land surface, V' and K' are defined as

$$K' = \frac{\pi C_b}{\kappa_b \tau}$$

and

$$V' = \frac{q C_w}{2\kappa_b}$$

C_b and C_w are the volumetric heat capacities of the bulk medium and water, κ_b is the thermal conductivity of the bulk medium, and τ is the period of forcing. Any self-consistent system of units (such as SI units) can be used to compute the recharge flux q, which is directly proportional to V'.

Below the depth of seasonal fluctuations, temperature profiles will be bowed away from linear profiles by recharging water, with the degree of curvature related to the magnitude of recharge (Figure 9). In dimensionless form (Bredehoeft and Papadopoulos, 1965)

$$\frac{[T(z) - T_0]}{[T_0 - T_L]} = \frac{[\exp(q/\lambda) - 1)]}{[\exp(Lq/\lambda) - 1)]} \tag{12}$$

where T_L is the temperature at the water table, at depth L, and $\lambda(= \kappa_b/C_b)$ is the thermal diffusivity of the bulk medium. As with equation (11), any self-consistent units can be used to compute the recharge flux q by inversion of the measured temperature profile, in this case the steady profile at depth.

Other Geophysical Methods

Many geophysical techniques provide data relevant to recharge based on the water-content dependence of gravitational, seismic, or electromagnetic properties of earth materials. For quantitative estimates of recharge rate, repeated high-precision gravity (microgravity) surveys indicate changes in mass due to recharge events (Pool and

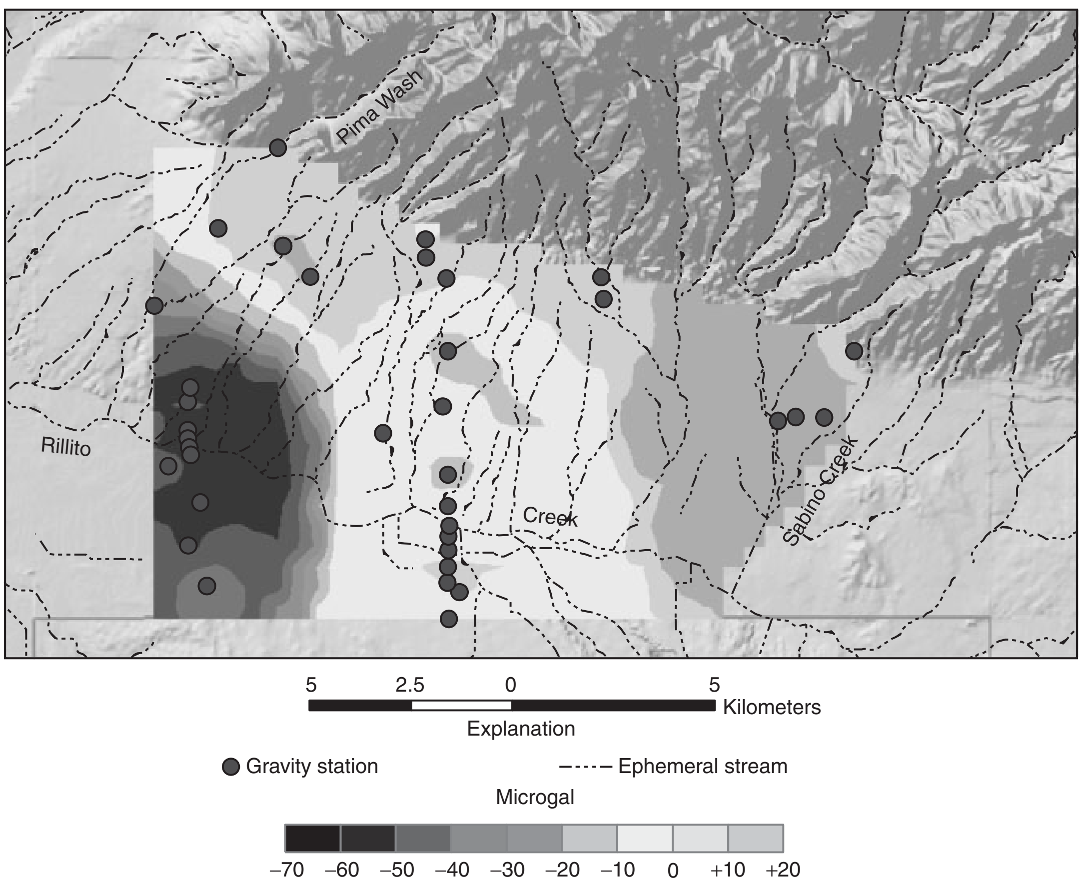

Figure 10 Changes in mass from aquifer recharge can be measured by repeated microgravity surveys. The example shows positive increases in gravity due to rising groundwater levels along the mountain front beneath tributaries to the ephemeral Rillito river, Arizona between June 1999 and March 2002. Pumping-related decreases in groundwater mass are evident downstream, beneath the main stem (unpublished data of Donald R. Pool, USGS, Tucson, AZ). A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

Schmidt, 1997; Figure 10). Similarly, repeated surveys using seismic or ground-penetrating-radar equipment show recharge-induced changes in water-table elevation (Haeni, 1986; Bohling *et al.*, 1989). Electrical resistance tomography can image zones of high-water content. Resistance tomographs have been used to map recharge areas in three dimensions based on land-surface measurements (Stonestrom *et al.*, 2003a). In addition to surface-based techniques, cross-bore tomographic imaging produces detailed three-dimensional reconstructions of water distribution and movement during periods of transient recharge (Daily *et al.*, 1992).

The next wave of advances in recharge estimation may involve remote sensing (RS) (see Part 05). Although still in the developmental stage, the broad spatial coverage afforded by satellite and aerial-borne instruments makes RS methods particularly attractive. Some of these methods attempt to directly determine changes in the amount of water stored in the subsurface, such as Synthetic Aperture Radar Interferometry (InSAR) (Hoffmann *et al.*, 2001; Lu and Danskin, 2001) and remotely sensed changes in gravity. Other RS methods contribute more indirectly to recharge estimation, for example, where RS data indicate certain components of the water budget equation, such as precipitation and evapotranspiration, or parameter values required in simulation models, such as soil properties, vegetation type and density, and land-use practices.

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