Choosing appropriate techniques for quantifying groundwater recharge

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Abstract Various techniques are available to quantify recharge; however, choosing appropriate techniques is often difficult. Important considerations in choosing a technique include space/time scales, range, and reliability of recharge estimates based on different techniques; other factors may limit the application of particular techniques. The goal of the recharge study is important because it may dictate the required space/time scales of the recharge estimates. Typical study goals include water-resource evaluation, which requires information on recharge over large spatial scales and on decadal time scales; and evaluation of aquifer vulnerability to contamination, which requires detailed information on spatial variability and preferential flow. The range of recharge rates that can be estimated using different approaches should be matched to expected recharge rates at a site. The reliability of recharge estimates using different techniques is variable. Techniques based on surface-water and unsaturated-zone data provide estimates of potential recharge, whereas those based on groundwater data generally provide estimates of actual recharge. Uncertainties in each approach to estimating recharge underscore the need for application of multiple techniques to increase reliability of recharge estimates.

Résumé Il existe différentes techniques pour quantifier la recharge; toutefois, il est souvent difficile de choisir les techniques appropriées. Les points importants pour le choix d'une technique sont l'échelle de temps et d'espace, la gamme de valeurs et la validité des estimations de la recharge basées sur différentes techniques; d'autres facteurs peuvent limiter l'application de techniques particulières. Le but des études de la recharge est important parce qu'il peut imposer les échelles de temps et d'espace des estimations de recharge. Les buts de ces études concernent habituellement l'évaluation des ressources en eau, qui requiert des informations sur la recharge à des échelles spatiales étendues et sur des durées comptées en dizaines d'années, et l'évaluation de la vulnérabilité des aquifères aux contaminations, qui exige des informations détaillées sur la variabilité spatiale et les écoulements préférentiels. La gamme des taux de recharge qui peuvent être estimés par différentes approches doit être adaptée aux valeurs attendues de la recharge sur le site. La validité des estimations de recharge faites par des techniques différentes est variable. Des techniques s'appuyant sur des données concernant les eaux de surface et la zone non saturée fournissent des estimations de recharge potentielle, tandis que celles basées sur les données des eaux souterraines donnent généralement des estimations de la recharge réelle. Les incertitudes de chaque approche d'estimation de la recharge mettent en relief la nécessité d'appliquer des techniques multiples pour accroître la validité des estimations de la recharge.

Resumen Existen diversas técnicas para cuantificar la recarga, pero elegir las apropiadas es a menudo difícil. Entre las consideraciones a tener en cuenta, hay que citar las escalas espacial y temporal, el rango y la fiabilidad de las estimaciones de la recarga obtenidas por medio de técnicas diferentes; hay otros factores que pueden limitar la aplicación de técnicas particulares. El objetivo de un estudio de recarga es importante, ya que puede condicionar las escalas temporal y espacial de las estimaciones. Los objetivos típicos comprenden la evaluación de recursos, cosa que requiere información de la recarga para escalas espaciales extensas y escalas temporales cifradas en décadas, y la evaluación de la vulnerabilidad del acuífero a la contaminación, para lo que hace falta información detallada sobre la variabilidad espacial y el flujo preferente. Se debería contrastar el rango de los valores estimados de recarga mediante enfoques diferentes con los valores esperados en un emplazamiento. La fiabilidad de las estimaciones basadas en técnicas diferentes es variable. Así, las técnicas basadas en datos de aguas superficiales y zona no saturada proporcionan estimaciones

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del potencial de recarga, mientras que las que se fundamentan en datos de aguas subterráneas proporcionan estimaciones de la recarga real. La incertidumbre asociadas a cada método corrobora la necesidad de aplicar varias técnicas distintas para aumentar la fiabilidad de las estimaciones de la recarga.

Keywords Groundwater recharge/water budget · Numerical modeling · Arid regions · Unsaturated zone · Groundwater age

Introduction

Determining which of a wide variety of techniques is likely to provide reliable recharge estimates is often difficult. Various factors need to be considered when choosing a method of quantifying recharge. A thorough understanding of the attributes of the different techniques is critical. The space/time scales of recharge estimates are important because different study goals require recharge estimates over different space and/or time scales. Whereas some studies focus on recharge estimates for water-resource assessment (Luckey et al. 1986; Kearns and Hendrickx 1998), others concentrate on estimates for contaminant transport or aquifer vulnerability to contamination (Egboka et al. 1983; Flury et al. 1994; Scanlon and Goldsmith 1997). Although spatial variability in recharge at local and intermediate scales may not be critical for water-resource assessment, it is critical for contaminant transport, because focused recharge and preferential flow allow contaminants to migrate rapidly through the unsaturated zone to underlying aquifers. Locating areas of low recharge is important for radioactive and hazardous-waste disposal sites (Tyler et al. 1996; Scanlon et al. 1997). Aquifer vulnerability to contamination is controlled to a large extent by recharge-transporting contaminants to aquifers (Robins 1998); therefore, delineation of zones of high recharge is critical to defining zones that are vulnerable to contamination. The time scales of recharge estimates are also important. Information on recharge at decadal time scales is generally required for water-resource planning, whereas time scales required for contaminant transport range from days to thousands of years, depending on the particular contaminants being considered.

Background information on potential controls on recharge is extremely valuable. The climate, geomorphology (including topography, soil, and vegetation), and geology of a site control the location and timing of recharge and therefore impact the choice of technique for estimating recharge. Humid and arid systems represent end members for different climates and generally require different approaches to quantify recharge. Humid regions are usually characterized by shallow water tables and gaining streams. Aquifers are often full, and groundwater is usually discharged through evapotranspiration and baseflow to streams. Diffuse recharge is dominant. Recharge rates in these regions are often limited by the ability of aquifers to store and transmit water, processes that are strongly affected by subsurface geology. In contrast, deep water tables and losing streams are common in alluvial valleys in arid regions. Focused recharge commonly dominates. Recharge rates are limited in large part by the availability of water at the land surface, which is controlled by climatic factors, such as precipitation and evapotranspiration, and by surface geomorphic features.

The purpose of this paper is to outline the attributes of the various techniques that have been used to quantify recharge, including the applicable space and time scales, the range of recharge rates that have been estimated with each technique, the reliability of the recharge estimates, and important factors that promote or limit the use of a particular technique. Because a detailed review of each technique is beyond the scope of this report, the reader is referred to various papers and books in the literature (e.g., Simmers 1988, 1997; Sharma 1989; Lerner et al. 1990; Lerner 1997; Robins 1998) and in this volume.

Terminology

Infiltration refers to water movement from the surface into the subsurface. In many unsaturated-zone studies, terms such as *net infiltration*, *drainage*, or *percolation* are used to describe water movement below the root zone, and these are often equated to recharge. Although *recharge* can be broadly defined as water that reaches an aquifer from any direction (down, up, or laterally) (Lerner 1997), this paper focuses on downward water movement across a water table. *Diffuse (direct) recharge* refers to recharge derived from precipitation or irrigation that occurs fairly uniformly over large areas, whereas *focused* or *localized recharge* refers to concentrated recharge from depressions in surface topography, such as streams, lakes, and playas. Some classifications restrict the term *localized recharge* to recharge from small depressions, joints, or rivulets and use the term *indirect recharge* for recharge beneath mappable features such as rivers and lakes (Lerner 1997). Rushton (1997) also distinguishes *actual recharge*, estimated from groundwater studies, and which reaches the water table, from *potential recharge*, estimated from surface-water and unsaturated-zone studies, and which is water that has infiltrated that may or may not reach the water table because of unsaturated-zone processes or the ability of the saturated zone to accept recharge.

Background Information for Recharge Estimation

The first stage of a recharge study in an area that has not previously been studied should involve collecting existing data on potential controls on recharge, such as climate, hydrology, geomorphology, and geology. These data are used to develop a conceptual model of recharge in the system. The conceptual model describes location, timing, and likely mechanisms of recharge and provides initial estimates of recharge rates.

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Climate plays a major role in controlling recharge, as shown by differences in recharge sources and rates between arid and humid settings. Preliminary recharge rates for a site can be estimated using available meteorological data and soil hydraulic-parameter data in unsaturated-zone models. Available hydrologic data should also be evaluated, including streamflow data (for evaluating gaining and losing sections of streams) and water-table depth (for determining unsaturated-zone thickness).

Variations in geomorphology reflect differences in topography, vegetation, and soil type, which can affect recharge. The impact of topography on local and regional groundwater flow paths was demonstrated by Tóth (1963). Recharge is generally considered to occur in topographic highs and discharge in topographic lows in humid regions, whereas in arid alluvial-valley regions recharge is usually focused in topographic lows, such as channels of ephemeral streams. The concept of hydrogeomorphic units was originally described by Tóth (1963) and Meyboom (1966, 1967). Delineation of hydrogeomorphic settings on the basis of topographic attributes – including slope classes, breaks in slope, curvature, and elevation – is greatly facilitated by the use of geographic information systems (GIS) and digital elevation models and is used widely in Australia (Salama et al. 1994; Hatton 1998). Vegetation cover is important in assessing recharge potential at a site. Recharge is generally much greater in nonvegetated than in vegetated regions (Gee et al. 1994) and greater in areas of annual crops and grasses than in areas of trees and shrubs (Prych 1998). The impact of vegetation was demonstrated in Australia, where replacement of deep-rooted native Eucalyptus trees with shallow-rooted crops resulted in recharge increases of about two orders of magnitude (<0.1 mm/year for native mallee vegetation to 5–30 mm/year for crop/pasture rotations) (Allison et al. 1990). Therefore, information on land use/land cover is important for evaluating recharge. Irrigated areas should also be identified because irrigation return flow often contributes significant amounts of recharge. Soil texture and permeability are important because coarse-grained soils generally result in higher recharge rates than do fine-grained soils. Cook et al. (1992) notes an apparent negative correlation between clay content in the upper 2 m and the recharge rate. Information on soil type and permeability is available from databases such as STATSGO (US Department of Agriculture 1994).

Information on topography, land use/land cover, and soil types is combined to define geomorphic systems that control recharge. Such systems are closely related to underlying geologic systems. Physiographic provinces represent regions of similar climate and geology that had similar geomorphic history. Examples of such provinces include alluvial-fan, fluvial, eolian, glacial, and coastal regions that may have characteristic recharge attributes. Lerner et al. (1990) describe recharge characteristics of several provinces, including alluvial-fan and fluvial systems and sand, sandstone, and limestone provinces. For example, alluvial fans near the base of mountain fronts usually receive recharge from streams draining adjacent mountains; therefore, techniques that quantify surface-water input to these systems are used to assess the total volume of recharge in alluvial fans. An understanding of the physiographic provinces is useful in developing a conceptual model of recharge in a system, and analogies can be drawn to similar physiographic provinces elsewhere. Insight is thus gained into recharge sources, flow mechanisms, and spatial and temporal variability in recharge.

Techniques for Estimating Recharge

For purposes of discussion, techniques for estimating recharge are subdivided into various types, on the basis of the three hydrologic sources, or zones, from which the data are obtained, namely surface water, unsaturated zone, and saturated zone. This subdivision of techniques is somewhat arbitrary and is probably not ideal. The different zones provide recharge estimates over varying space and time scales. Within each zone, techniques are generally classified into physical, tracer, or numerical-modeling approaches. This overview focuses on aspects of each approach that are important in choosing appropriate techniques, such as the space/time scales, range, and reliability of recharge estimates. The range of recharge rates for different techniques is based on evaluation of the literature and general evaluation of uncertainties and should be considered only approximate. Because many techniques in the different zones are based on the water-budget equation, this topic is described separately.

Water Budget

The water budget for a basin can be stated as:

$$P + Q_{on} = ET + Q_{off} + \Delta S \tag{1}$$

where P is precipitation (and may also include irrigation); Q_{on} and Q_{off} are water flow onto and off the site, respectively; ET is evapotranspiration; and ΔS is change in water storage. All components are given as rates (e.g., mm/day or mm/year). Individual components consist of subcomponents. Water flow onto or off the site is written as the sum of surface flow, interflow, and groundwater flow. ET is distinguished on the basis of the source of evaporated water (surface, unsaturated zone, or saturated zone). Water storage takes place in snow, surface-water reservoirs, the unsaturated zone, and the saturated zone. Rewriting the water-budget equation to incorporate many of these subcomponents results in:

$$\begin{aligned} P + Q_{on}^{sw} + Q_{on}^{gw} = {} & ET^{sw} + ET^{uz} + ET^{gw} + R_0 \\ & + Q_{off}^{gw} + Q^{bf} + \Delta S^{snow} \\ & + \Delta S^{sw} + \Delta S^{uz} + \Delta S^{gw} \end{aligned} \tag{2}$$

where superscripts refer to the subcomponents described above, R_0 (runoff) is surface-water flow off the site, and Q^{bf} is baseflow (groundwater discharge to streams or springs). Groundwater recharge, R, includes any infiltrat-

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ing water that reaches the saturated zone and can be written as (Schicht and Walton 1961):

$$R = Q^{gw}_{off} - Q^{gw}_{on} + Q^{bf} + ET^{gw} + \Delta S^{gw} \tag{3}$$

This equation simply states that all water arriving at the water table either flows out of the basin as groundwater flow, is discharged to the surface, is evapotranspired, or is retained in storage. Substituting this equation into Eq. (2) produces the following version of the water budget:

$$R = P + Q^{sw}_{on} - R_0 - ET^{sw} - ET^{uz} - \Delta S^{snow} - \Delta S^{sw} - \Delta S^{uz} \tag{4}$$

Water-budget methods are those that are based, in one form or another, on a water-budget equation. They include most hydrologic models, such as surface-water and groundwater flow models. For any site, some of the terms in Eq. (4) are likely to be negligible in magnitude and therefore may be ignored.

The most common way of estimating recharge by the water-budget method is the indirect or "residual" approach, whereby all of the variables in the water-budget equation except R are measured or estimated, and R is set equal to the residual. An advantage of water-budget methods is flexibility. Few assumptions are inherent in Eq. (1). The methods are not hindered by any presuppositions as to the mechanisms that control the individual components. Hence, they can be applied over a wide range of space and time scales, ranging from lysimeters (centimeters, seconds) to global climate models (kilometers, centuries).

The major limitation of the residual approach is that the accuracy of the recharge estimate depends on the accuracy with which the other components in the water-budget equation are measured. This limitation is critical when the magnitude of the recharge rate is small relative to that of the other variables, in particular ET. In this case, small inaccuracies in values of those variables commonly result in large uncertainties in the recharge rate. Some authors (e.g., Gee and Hillel 1988; Lerner et al. 1990; and Hendrickx and Walker 1997) have therefore questioned the usefulness of water-budget methods in arid and semiarid regions. However, if the water budget is calculated on a daily time step, P sometimes greatly exceeds ET on a single day, even in arid settings. Averaging over longer time periods tends to dampen out extreme precipitation events (those most responsible for recharge events). Methods for measuring or estimating various components of the water budget are described in Hillel (1980), Rosenberg et al. (1983), and Tindall and Kunkel (1999).

Techniques Based on Surface-Water Studies

The status of recharge related to surface-water bodies depends on the degree of connection between surface-water and groundwater systems (Fig. 1; Sophocleous 2002, this volume). Humid regions are generally characterized by gaining surface-water bodies because groundwater discharges to streams and lakes. In contrast, arid regions are generally characterized by losing surface-water bodies, because surface-water and groundwater systems are often separated by thick unsaturated sections. Therefore, surface-water bodies often form localized recharge sources in arid settings. Recharge can be estimated using surface-water data in gaining and losing surface-water bodies.

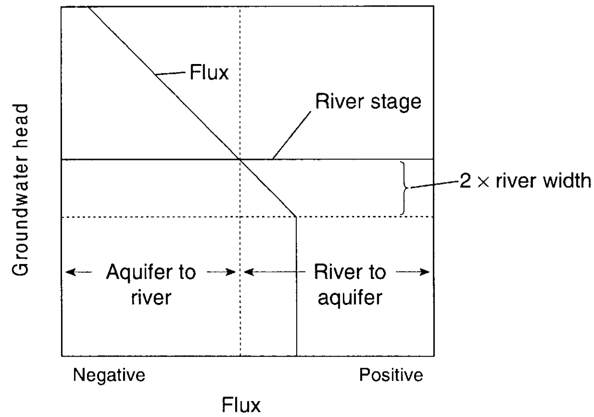

Fig. 1 Water fluxes related to the degree of connection between rivers and aquifers. The aquifer discharges to the river when the groundwater head is greater than the river stage, whereas the river recharges the aquifer when the river stage is greater than the groundwater head. Recharge values generally reach a constant rate when the water-table depth is greater than twice the river width (Bouwer and Maddock 1997)

Physical techniques

Channel-water budget

Surface-water gains or losses can be estimated using channel-water budgets based on stream-gauging data. Lerner et al. (1990), Lerner (1997), and Rushton (1997) provide detailed reviews of this approach. The channel-water budget is described as (Lerner 1997):

$$R = Q_{up} - Q_{down} + \sum Q_{in} - \sum Q_{out} - E_a - \frac{\Delta S}{\Delta t} \tag{5}$$

where R is recharge rate, Q is flow rate, Q_{up} and Q_{down} are flows at the upstream and downstream ends of the reaches, Q_{in} and Q_{out} refer to tributary inflows and outflows along the reach, E_a is the evaporation from surface water or stream bed, and ΔS is change in channel and unsaturated-zone storage over change in time (Δt). The term *transmission loss* refers to the loss in streamflow between upstream and downstream gauging stations (Lerner et al. 1990). This loss reflects potential recharge that can result in an overestimate of actual recharge because of bank storage and subsequent evapotranspiration, development of perched aquifers, and inability of the aquifer to accept recharge because of a shallow water table or low transmissivity (Lerner et al. 1990). Recharge values generally reach a constant rate when the water-table depth is greater than twice the stream width, because flow is generally controlled by gravity at these depths (Bouwer and Maddock 1997; Fig. 1).

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The range of recharge rates that can be measured using this technique depends on the magnitude of the transmission losses relative to the uncertainties in the gauging data and tributary flows. Gauging data in the US are generally considered to be accurate to ±5% (Rantz 1982). Lerner et al. (1990) indicate that measurement errors during high flows are often ±25% and sometimes range from –50 to +100% during flash floods in semiarid regions. The recharge estimates represent average values over the reach between the gauging stations. The temporal scales represented by the recharge values range from event scale (minutes to hours) to much longer time scales that are estimated by summation of individual events.

Seepage meters

Seepage to or from surface-water bodies can be measured by using seepage meters (Kraatz 1977; Lee and Cherry 1978). A seepage meter consists of a cylinder that is pushed into the bottom of the stream or lake. Attached to the cylinder is a reservoir of water; the rate at which water within the cylinder infiltrates is determined by changes in reservoir volume. This method is inexpensive and easy to apply. An automated seepage meter is described by Taniguchi and Fukuo (1993). Uncertainties in estimated fluxes can be determined from replicate measurements. Seepage fluxes measured in different US studies vary greatly, from approximately 1 to 3,000 mm/day at a site in Minnesota (Rosenberry 2000), 9 to 223 mm/day in Minnesota and Wisconsin (Lee 1977), and 12 to 122 mm/day in Nevada (Woessner and Sullivan 1984). Because seepage meters provide point estimates of water fluxes, measurements may be required at many locations for a representative value to be obtained. Time scales range from those based on individual events to days. Recharge over longer times is estimated from a summation of shorter times.

Baseflow discharge

In watersheds with gaining streams, groundwater recharge can be estimated from stream hydrograph separation (Meyboom 1961; Rorabough 1964; Mau and Winter 1997; Rutledge 1997; Halford and Mayer 2000). Use of baseflow discharge to estimate recharge is based on a water-budget approach (Eq. 3), in which recharge is equated to discharge. Baseflow discharge, however, is not necessarily directly equated to recharge because pumpage, evapotranspiration, and underflow to deep aquifers may also be significant. These other discharge components should be estimated independently. Bank storage may complicate hydrograph analysis because water discharging from bank storage is generally derived from short-term fluctuations in surface-water flow and not from areal aquifer recharge and could result in overestimation of recharge. Various approaches are used for hydrograph separation, including digital filtering (Nathan and McMahon 1990; Arnold et al. 1995) and recession-curve displacement methods (Rorabough 1964). The accuracy of the reported recharge rates depends on the validity of the various assumptions. Recharge estimates based on hydrograph separation range from 152 to 1,270 mm/year in 89 basins (Rutledge and Mesko 1996) and from 127 to 635 mm/year in 15 basins (Rutledge and Daniel 1994) in the eastern US. Rutledge (1998) recommended an upper limit on basin size of 1,300 km^2 for application of this method because of difficulties in separating surface-water and groundwater flow and bank-storage effects in larger systems and because of the areally uniform recharge assumption. The minimum time scale is a few months. Recharge over longer times can be estimated by summation of estimates over shorter times. Recent progress has been made on the use of chemical and isotopic techniques to infer the sources of streamflow from end members such as rainfall, soil water, groundwater, and bank storage (Hooper et al. 1990; Christophersen and Hooper 1992). This approach is data intensive, but it provides information that is useful in conducting hydrograph separation. Suecker (1995) used sodium concentrations in a two-component mixing model to determine the subsurface contribution to three alpine streams in Colorado, USA.

Tracer techniques

Heat tracer

Installation and maintenance of stream-gauging stations are expensive and difficult, particularly in ephemeral streams in semiarid regions that are subject to erosion. As an alternative to stream gauging, heat can be used as a tracer to provide information on when surface water is flowing in ephemeral streams and to estimate infiltration from surface-water bodies (Stallman 1964; Lapham 1989; Constantz et al. 1994; Ronan et al. 1998). Monitoring depths vary, depending on time scales, sediment types, and anticipated water fluxes beneath the stream. Diurnal temperature fluctuations are generally monitored at depths of ~0.05 to 1 m for fine-grained material, and 0.3 to 3 m for coarse-grained material. Depths for monitoring annual temperature fluctuations are generally an order of magnitude greater. Measured temperature is used with inverse modeling using a nonisothermal variably saturated flow code, such as VS2DH (Healy and Ronan 1996), to estimate hydraulic conductivity of the sediments. Data analysis is complex, and inverse solutions are not always unique. Percolation rates can be estimated if the hydraulic head is calculated from measured data. Temperature can be monitored accurately and inexpensively using thermistors or thermocouples. Heat dissipation sensors can be used to monitor temperature and matric potential simultaneously in unsaturated media.

The minimum net infiltration rate that can be estimated using heat as a tracer depends on the range of surface-water temperature fluctuations and the time scale considered. Stallman (1964) suggests a minimum recharge rate of ~20 mm/day in the stream bed using diurnal temperature fluctuations, and ~1 mm/day using annual temperature fluctuations in natural media with average heat

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properties. Reported infiltration rates from various studies include 0.05 to 6.4 mm/day (Maurer and Thodal 2000), 18 to 37 mm/day (Bartolino and Niswonger 1999), and 457 mm/day (Lapham 1989). Previous studies generally used a single vertical array of temperature sensors and therefore provide an estimate of one-dimensional flow at a point; however, some ongoing studies include two- and three-dimensional arrays of sensors to provide more realistic three-dimensional flux estimates beneath streams (R. Niswonger, US Geological Survey, personal communication, 2001). Recharge can be estimated for time periods ranging from hours to years.

Isotopic tracers
Stable isotopes of oxygen and hydrogen are used to identify groundwater recharge from rivers and lakes. In regions where rivers have headwaters at high elevations (in mountainous areas), river water is often depleted in stable isotopes relative to local precipitation in adjacent basins. If rivers retain the depleted isotopic signature of the headwaters, the difference in stable-isotope signatures of rivers and local precipitation can be used to determine the relative contribution of these two sources of groundwater recharge. Recharge to groundwater on the Wairau Plain and on the Canterbury Plains, South Island, New Zealand, is attributed primarily to rivers (depleted stable isotopes from mountain catchments), whereas the contribution of locally infiltrated precipitation is much smaller (Taylor et al. 1989, 1992). Groundwater recharge in an aquifer in the Netherlands was attributed to bank infiltration from the Rhine River because of the depleted ^{18}O signature, which is a result of the Alpine catchment basin of the Rhine (Stuyfzand 1989). Isotopic tracers provide information on recharge sources; however, it is generally difficult to quantify recharge rates. The time scales range from seasonal in areas of high flux to hundreds of years in areas of low flux.

Numerical modeling
Watershed (rainfall/runoff) modeling is used to estimate recharge rates over large areas. Singh (1995) reviewed many watershed models, which generally provide recharge estimates as a residual term in the water-budget equation (Eq. 4; Arnold et al. 1989; Leavesley and Stannard 1995; Hatton 1998). The minimum recharge rate that can be estimated is controlled by the accuracy with which the various parameters in the water budget can be measured (~±10%) and the time scale considered. The various watershed models differ in spatial resolution of the recharge estimates. Some models are termed *lumped* and provide a single recharge estimate for the entire catchment (Kite 1995). Others are spatially disaggregated into hydrologic-response units (HRUs) or hydrogeomorphological units (HGUs) (Salama et al. 1993; Leavesley and Stannard 1995). Watershed models are applied at a variety of scales. Bauer and Mastin (1997) applied the Deep Percolation Model to three small watersheds (average size 0.4 km^2) in the Puget Sound in Washington, USA. Average annual recharge rates are 37, 138, and 172 mm for the three basins. Arnold et al. (2000) applied the SWAT model to the upper Mississippi River Basin, USA (492,000 km^2). The basin was divided into 131 hydrologic-response units with an average area of 3,750 km^2. Estimated annual recharge ranged from ~10 to 400 mm. Small-scale applications allow more precise methods to be used to measure or estimate individual parameters of the water-budget equation (Healy et al. 1989). Time scales in models are daily, monthly, or yearly. Daily time steps are desirable for estimation of recharge because recharge is generally a larger component of the water budget at smaller time scales. Other recent applications of watershed models to estimate recharge include Arnold and Allen (1996; recharge rates 85 to 191 mm/year, Illinois, USA), Sami and Hughes (1996; recharge rate ~6 mm/year in a fractured system, South Africa), and Flint et al. (2002, this volume; recharge rate 2.9 mm/year at Yucca Mountain, Nevada, USA).

Techniques Based on Unsaturated-Zone Studies

Unsaturated-zone techniques for estimating recharge are applied mostly in semiarid and arid regions, where the unsaturated zone is generally thick. These techniques are described in detail in Hendrickx and Walker (1997), Scanlon et al. (1997), Gee and Hillel (1988), and Zhang (1998). The recharge estimates generally apply to smaller spatial scales than those calculated from surface-water or groundwater approaches. Unsaturated-zone techniques provide estimates of potential recharge based on drainage rates below the root zone; however, in some cases, drainage is diverted laterally and does not reach the water table. In addition, drainage rates in thick unsaturated zones do not always reflect current recharge rates at the water table.

Physical techniques
Lysimeters
The various components of the soil water budget are accurately measured by using lysimeters (Brutsaert 1982; Allen et al. 1991; Young et al. 1996). Lysimeters consist of containers filled with disturbed or undisturbed soil, with or without vegetation, that are hydrologically isolated from the surrounding soil, for purposes of measuring the components of the water balance. All lysimeters are designed to allow collection and measurement of drainage. Precipitation and water storage are measured separately in drainage lysimeters (also termed pan or nonweighing lysimeters). Weighing lysimeters are generally used for accurate measurements of evapotranspiration. They are constructed on delicate balances capable of measuring slight changes in weight that represent precipitation and water-storage changes. Surface areas of lysimeters range from 100 cm^2 (Evett et al. 1995) to ~300 m^2 for large pan lysimeters (Ward and Gee 1997); depths range from tens of centimeters to 10 to 20 m (Gee

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et al. 1994). If the base of the lysimeter is not deeper than the root zone, measured drainage fluxes overestimate aquifer recharge rates. Therefore, lysimeters are generally unsuitable for areas with deep-rooted vegetation. Recharge rates can be estimated at time scales from minutes to years. The minimum water flux that can be measured using a lysimeter depends on the accuracy of the drainage measurements and the surface area of the lysimeter. For large lysimeters (surface area ~100 m^2), recharge rates of about 1 mm/year can be resolved. The upper flux that can be measured depends on the design of the drainage system but should exceed drainage fluxes in most natural settings. A wide variety of recharge rates has been measured using lysimeters: 342 to 478 mm/year over a 3-year period for the Bunter Sandstone, England (surface area 100 m^2; Kitching et al. 1977), 200 mm/year for the Chalk Aquifer, England (surface area 25 m^2; Kitching and Shearer 1982), and 1 to 200 mm/year in an 18-m-deep lysimeter in a semiarid site (Hanford, Washington, USA; Gee et al. 1992). Most lysimeters have a drainage-collection system that is open to the atmosphere, which creates a seepage-face lower-boundary condition. For thick unsaturated zones, this artifact causes different moisture and pressure-head profiles in the lysimeter relative to those in the adjacent undisturbed area (van Bavel 1961). In an effort to minimize the influence of the bottom boundary, some lysimeters have been built with a porous plate on the bottom that is set at a prescribed pressure head. Lysimeters are not routinely used to estimate recharge because they are expensive and difficult to construct and have high maintenance requirements. They are more suitable for evaluation of evapotranspiration than recharge.

Zero-flux plane

The soil-water budget can be simplified by equating recharge to changes in soil-water storage below the zero-flux plane (ZFP), which represents the plane where the vertical hydraulic gradient is zero. The ZFP separates upward (ET) from downward (drainage) water movement. The rate of change in the storage term between successive measurements is assumed to be equal to the drainage rate to the water table or the recharge rate. The ZFP requires soil matric-potential measurements to locate the position of the ZFP and soil-water-content measurements to estimate storage changes. The ZFP method, first described in Richards et al. (1956), has been used in various studies (Royer and Vachaud 1974; Wellings 1984; Dreiss and Anderson 1985; Healy et al. 1989). The minimum recharge rate that can be measured is controlled by the accuracy of the water-content measurements (generally ±0.01 m^3/m^3). Recharge rates estimated by this method range from 34 to 149 mm/year (eight sites, semiarid region, Western Australia; Sharma et al. 1991), 78 to 300 mm/year (Chalk and Sandstone aquifers, England; Cooper et al. 1990), and 345 to 469 mm/year (Upper Chalk aquifer, southern England; Wellings 1984). The ZFP provides a recharge estimate at the measurement point. Time scales range from event scales to years.

The ZFP technique cannot be used when water fluxes are downward throughout the entire profile or when water storage is increasing, because downward movement of a wetting front generally masks the zero-flux plane. A simplified water-budget approach is generally used for these conditions (Hodnett and Bell 1990; Roman et al. 1996). The ZFP technique is relatively expensive in terms of the required instruments and amount of data collection. This technique works best in regions where large fluctuations exist in soil-water content throughout the year and where the water table is always deeper than the ZFP.

Darcy's law

Darcy's law is used to calculate recharge (R) in the unsaturated zone according to the following equation:

$$R = -K(\theta)\,dH/dz = -K(\theta)\frac{d}{dz}(h+z) = -K(\theta)\left(\frac{dh}{dz}+1\right) \tag{6}$$

where $K(\theta)$ is the hydraulic conductivity at the ambient water content, θ; H is the total head; h is the matric pressure head; and z is elevation. Application of Darcy's law requires measurements or estimates of the vertical total-head gradient and the unsaturated hydraulic conductivity at the ambient soil-water content. The method has been applied in many studies under arid and semiarid conditions (Enfield et al. 1973; Sammis et al. 1982; Stephens and Knowlton 1986) and also under humid conditions (Ahuja and El-Swaify 1979; Steenhuis et al. 1985; Kengni et al. 1994; Normand et al. 1997). For thick unsaturated zones, below the zone of fluctuations related to climate, in uniform or thickly layered porous media, the matric pressure gradient is often nearly zero, and water movement is essentially gravity driven. Under these conditions, little error results by assuming that the total head gradient is equal to 1 (unit-gradient assumption) (Gardner 1964; Childs 1969; Chong et al. 1981; Sisson 1987). The unit-gradient assumption removes the need to measure the matric pressure gradient and sets recharge equal to the hydraulic conductivity at the ambient water content. The unit gradient assumption has been used in many studies (Sammis et al. 1982; Stephens and Knowlton 1986; Healy and Mills 1991; Nimmo et al. 1994).

The minimum recharge rate that can be estimated by using Darcy's law depends on the accuracy of the hydraulic conductivity and head-gradient measurement if the latter is not unity. Accurate measurements of hydraulic conductivity as low as 1×10^{-9} cm/s can be obtained by using the steady-state centrifuge (SSC) method; this value corresponds approximately to 0.3 mm/year (Nimmo et al. 1992). However, problems with sample disturbance and drying during collection and spatial variability in hydraulic conductivity generally result in a measurement limit of about 20 mm/year. Recharge rates determined by the Darcy method range from 37 mm/year in an arid region (New Mexico, USA; Stephens and Knowlton 1986) to about 500 mm/year for an irrigated site having a thin unsaturated zone (near Grenoble, France; Kengni et al. 1994). If hydraulic conductivity is strongly dependent on water content, uncertainty increases (Nimmo et al. 1994). This method provides a point estimate of recharge over a wide range of time scales; however, if applied at significant depths in thick vadose zones, it may represent a larger area. An attractive feature of the Darcy method is that it can be applied throughout the entire year, whereas the ZFP can be applied only at certain times of the year.

Tracer techniques

Applied tracers

Chemical or isotopic tracers are applied as a pulse at the soil surface or at some depth within the soil profile to estimate recharge (Athavale and Rangarajan 1988; Sharma 1989). Infiltration of precipitation or irrigation transports the tracer to depth. Commonly used tracers include bromide, ^3H, and visible dyes (Athavale and Rangarajan 1988; Kung 1990; Flury et al. 1994; Aeby 1998; Forrer et al. 1999). Organic dyes are generally used to evaluate preferential flow (Flury et al. 1994; Scanlon and Goldsmith 1997). Although ^3H is the most conservative of all tracers, its use is prohibited in many areas because of environmental-protection laws. Kung (1990) showed that bromide uptake by plants is often significant, and sorption is important for organic dyes. The subsurface distribution of applied tracers is determined sometime after the application by digging a trench for visual inspection and sampling, or by drilling test holes for sampling. The vertical distribution of tracers is used to estimate the velocity (v) and the recharge rate (R):

$$R = v\theta = \frac{\Delta z}{\Delta t}\theta \tag{7}$$

where Δz is depth of the tracer peak, Δt is time between tracer application and sampling, and θ is volumetric water content. The minimum water flux that can be measured with applied tracers depends on the time between application and sampling and, in the case of surface-applied tracers, the root-zone depth. If the rooting depth is assumed to be 0.5 m and the average water content 0.2 m^3/m^3, a recharge rate of 100 mm/year would be required to transport the applied tracer through the root zone in 1 year. Lower recharge rates can be measured when the tracers are applied below the root zone. The maximum water flux that can be measured depends on the depth to the water table. Recharge rates resulting from excess irrigation were evaluated by Rice et al. (1986) using surface-applied bromide to a bare field followed by 45 cm of irrigation water for 159 days. The resultant recharge rate was 3.3 mm/day, which exceeded that estimated using a water balance by a factor of 5. The discrepancy was attributed to preferential flow. The tritium injection technique (at a depth of 0.7 m) was used to estimate recharge rates that ranged from 6.5 to 100 mm/year in several basins in southern India (Athavale and Rangarajan 1990). Sharma et al. (1985) applied bromide at the soil surface under natural precipitation in a vegetated area and estimated a recharge rate of 224 mm for a 76-day period. Tracers are generally applied at a point or over small areas (10 to 200 m^2). The calculated recharge rates represent the time between application and sampling, which is generally months to years.

Historical tracers

Historical tracers result from human activities or events in the past, such as contaminant spills (Nativ et al. 1995) or atmospheric nuclear testing (^3H and ^{36}Cl) (Fig. 2). These historical tracers or event markers are used to estimate recharge rates during the past 50 years (Phillips et al. 1988; Scanlon 1992; Cook et al. 1994). Industrial and agricultural sources produce contaminants such as bromide, nitrate, atrazine, and arsenic, and these can provide qualitative evidence of recent recharge; however, uncertainties with respect to source location, concentration, and timing of contamination, as well as possible nonconservative behavior of contaminants, make it difficult to quantify recharge. The presence of an event marker in water suggests that a component of that water recharged in a particular time period. The peak concentration of thermonuclear tracers can also be used to estimate water flux by using Eq. (7), where z is approximated by the depth of the peak concentration of the tracer, θ is the average water content above the tracer peak, and t is the time period between the peak tracer fallout and the time the samples were collected. The minimum recharge rate that can be estimated using thermonuclear tracers is about 10 mm/year, because of the time required to move through the root zone (Eq. 7; Cook and Walker 1995). In many areas where these tracers have been used, the bomb pulse peak is still in the root zone (^{36}Cl, 1.8 mm/year, Norris et al. 1987; ^{36}Cl, 2.5 to 3 mm/year, ^3H, 6.4 to 9.5 mm/year, Phillips et al. 1988; ^{36}Cl, 1.4 mm/year, ^3H, 7 mm/year, Scanlon 1992), indicating that water fluxes at these sites are extremely low, which is an important consideration for waste disposal. Because much of this water in the root zone is later evapotranspired, water fluxes estimated from tracers within the root zone overestimate water fluxes below the root zone by as much as several orders of magnitude (Tyler and Walker 1994; Cook and Walker 1995). Deep penetration of thermonuclear tracers has been found in sandy soils in arid settings (^3H, 23 mm/year, Dincer et al. 1974; ^3H, 22 to 26 mm/year, Aranyossy and Gaye 1992). The maximum water flux that can be estimated may be limited by depth to groundwater. For example, if the average water content is 0.1 m^3/m^3 and the time since peak fallout is 40 years, a recharge rate of 50 mm/year would result in a peak at a depth of 20 m. Therefore, this technique is generally unsuitable where recharge rates are much greater than 50 mm/year. Theoretically, the technique could be used for higher recharge rates if the water table were deeper; however, the difficulty of soil sampling at these depths and locating the tracer peak may be prohibitive. Historical tracers provide point estimates of water flux over the last 50 years.

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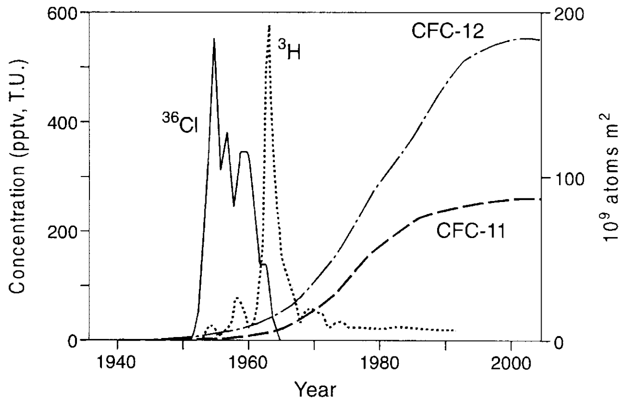

Fig. 2 Input functions for historical tracers, including ^3H, ^{36}Cl, CFC-11, and CFC-12 (^3H data from Ottawa, Canada; ^{36}Cl data, Phillips 2000; CFC data, http://water.usgs.gov/lab/cfc/background/air_curve.html)

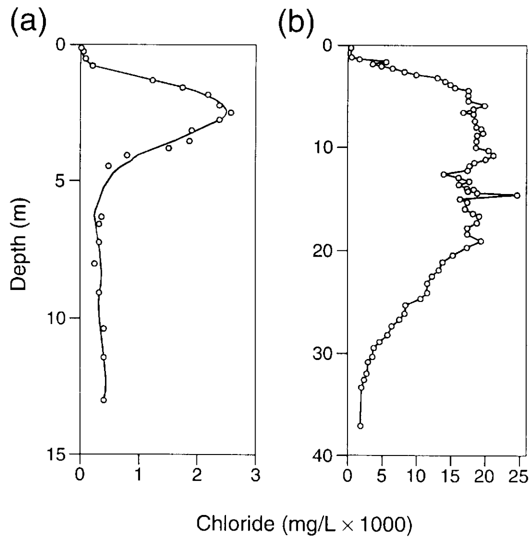

Fig. 3 Typical chloride profiles in unsaturated systems: bulge-shaped chloride profile attributed to **a** paleoclimatic variation (Scanlon 1991) and **b** diffusion to a shallow water table (Cook et al. 1989)

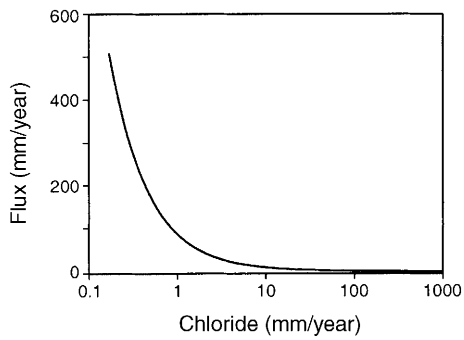

Fig. 4 Sensitivity of drainage calculated using the chloride mass-balance approach to chloride concentrations, based on data from boreholes at a site in the Chihuahuan Desert, Texas, USA (Scanlon 2000)

Environmental tracers – chloride

Environmental tracers such as chloride (Cl) are produced naturally in the Earth's atmosphere and are used to estimate recharge rates (Allison and Hughes 1978; Scanlon 1991, 2000; Phillips 1994). The mass of Cl into the system (precipitation and dry fallout, P) times the Cl concentration in P (C_P) is balanced by the mass out of the system (drainage, D) times the Cl concentration in drainage water in the unsaturated zone (C_{uz}) if surface runoff is assumed to be zero:

$$PC_p = DC_{uz} \qquad D = \frac{PC_p}{C_{uz}} \tag{8}$$

Chloride concentrations generally increase through the root zone as a result of evapotranspiration and then remain constant below this depth. Bulge-shaped Cl profiles at some sites have been attributed to paleoclimatic variations or to diffusion to a shallow water table (Fig. 3). Drainage is inversely related to Cl concentration in the unsaturated-zone pore water (Eq. 8). This inverse relationship results in the Cl mass-balance (CMB) approach being much more accurate at low drainage rates, because Cl concentrations change markedly over small changes in drainage (Fig. 4). The CMB approach has been most widely used for estimating low recharge rates, largely because of the lack of other suitable methods. Water fluxes as low as 0.05 to 0.1 mm/year have been estimated in arid regions in Australia and in the US (Allison and Hughes 1983; Cook et al. 1994; Prudic 1994; Prych 1998). Low recharge rates are reported to be consistent with radioactive decay of ^{36}Cl at a site in the US (Scanlon 2000). Somewhat higher recharge rates have been calculated from Cl concentrations measured in sinkholes in Australia (>60 mm/year; Allison et al. 1985), sand dunes cleared of vegetation in Australia (4 to 28 mm/year; Cook et al. 1994), and sands with sparse vegetation in Cyprus (33 to 94 mm/year; Edmunds et al. 1988). The maximum water flux that can be estimated is based on uncertainties in measuring low Cl concentrations and potential problems with Cl contributions from other sources and is generally considered to be about 300 mm/year. Scanlon and Goldsmith (1997) report uncertainties of an order of magnitude beneath ephemeral lakes (playas) in the US because of uncertainties in Cl input from runon to playas. The CMB approach provides point estimates of recharge rates. Temporal scales range from decades to thousands of years (Scanlon 2000).

Numerical modeling

Unsaturated-zone modeling is used to estimate deep drainage below the root zone or recharge in response to meteorological forcing. Recent advances in computer technology and in computer codes have made long-term simulations of recharge more feasible. A variety of approaches is used to simulate unsaturated flow, including soil-water storage-routing approaches (bucket model; Flint et al. 2002, this volume; Walker et al. 2002, this

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volume), quasi-analytical approaches (Kim et al. 1996; Simmons and Meyer 2000), and numerical solutions to the Richards equation. Examples of codes that use the Richards equation include BREATH (Stothoff 1995), HYDRUS-1D, HYDRUS-2D (Simunek et al. 1996), SWIM (Ross 1990), VS2DT (Lappala et al. 1987; Hsieh et al. 2000), and UNSATH (Fayer 2000). Theoretically, the range of recharge rates that can be estimated using numerical modeling is infinite; however, the reliability of these estimates should be checked against field information such as lysimeter data, tracers, water content, and temperature (Scanlon and Milly 1994; Andraski and Jacobson 2000; Simmons and Meyer 2000; Flint et al. 2002, this volume). Bucket-type models can be used over large areas (Flint et al. 2002, this volume); however, models based on the Richards equation are often restricted to evaluating small areas (≤100 m^2) or to one-dimensional flow in the shallow subsurface (≤15 m depth). Time scales that can be evaluated range from hours to decades; however, many recharge modeling studies evaluate periods of 30 to 100 years because of availability of meteorological information (Rockhold et al. 1995; Stothoff 1997; Kearns and Hendrickx 1998). Because of uncertainties in hydraulic conductivity and nonlinear relationships between hydraulic conductivity and matric potential or water content, recharge estimates based on unsaturated-zone modeling that use the Richards equation may be highly uncertain. Numerical modeling is generally used as a tool to evaluate flow processes and to assess sensitivity of model output to various parameters. Stothoff (1997) evaluated the impact of alluvial-cover thickness overlying fractured bedrock on recharge and reports high recharge rates (≤50% of precipitation) if alluvial-cover thicknesses are less than 25 to 50 cm and little or no recharge for alluvial-cover thicknesses of 50 to 500 cm. The effect of soil texture and vegetation was evaluated by Rockhold et al. (1995) for a 30-year period (1963–1992). Recharge rates range from 0.5 mm/year (sagebrush on sand) to 22 mm/year (bare sand). Recharge rates also vary with soil texture (7.6 mm/year, bare silt loam, to 22 mm/year, bare sand). A similar study was conducted for a 100-year period by Kearns and Hendrickx (1998), who also demonstrate the effects of vegetation and soil texture on recharge rates.

Techniques Based on Saturated-Zone Studies

Most unsaturated-zone techniques provide point estimates of recharge; saturated-zone techniques commonly integrate over much larger areas. Whereas surface-water and unsaturated-zone approaches provide estimates of drainage or potential recharge, saturated-zone approaches provide evidence of actual recharge because water reaches the water table.

Physical Techniques

Water-table fluctuation method

The water-table fluctuation (WTF) method is based on the premise that rises in groundwater levels in unconfined aquifers are due to recharge water arriving at the water table. Recharge is calculated as

$$R = S_y dh/dt = S_y \Delta h/\Delta t \tag{9}$$

where S_y is specific yield, h is water-table height, and t is time. The WTF method has been used in various studies (Meinzer and Stearns 1929; Rasmussen and Andreasen 1959; Gerhart 1986; Hall and Risser 1993) and is described in detail by Healy and Cook (2002, this volume). The method is best applied over short time periods in regions having shallow water tables that display sharp rises and declines in water levels. Analysis of water-level fluctuations can, however, also be useful for determining the magnitude of long-term changes in recharge caused by climate or land-use change. Difficulties in applying the method are related to determining a representative value for specific yield and ensuring that fluctuations in water levels are due to recharge and are not the result of changes in atmospheric pressure, the presence of entrapped air, or other phenomena, such as pumping. The method has been applied over a wide variety of climatic conditions. Recharge rates estimated by this technique range from 5 mm/year in the Tabalah Basin of Saudi Arabia (Abdulrazzak et al. 1989) to 247 mm/year in a small basin in a humid region of the eastern US (Rasmussen and Andreasen 1959). Water-level fluctuations occur in response to spatially averaged recharge. The area represented by the recharge rates ranges from tens of square meters to several hundred or thousand square meters. Time periods represented by the recharge estimates range from event scale to the length of the hydrographic record.

Darcy's law

Darcy's law is used to estimate flow through a cross section of an unconfined or confined aquifer. This method assumes steady flow and no water extraction. The subsurface water flux (q) is calculated by multiplying the hydraulic conductivity by the hydraulic gradient. The hydraulic gradient should be estimated along a flow path at right angles to potentiometric contours. The volumetric flux through a vertical cross section of an aquifer (A) is equated to the recharge rate (R) times the surface area that contributes to flow (S):

$$qA = RS \tag{10}$$

The cross section should be aligned with an equipotential line. The Darcy method has been used by Theis (1937) and Belan and Matlock (1973). The method is easy to apply if information on large-scale, effective hydraulic conductivity and the hydraulic gradient is available. The area should reflect a natural system with minimal pumpage. Recharge estimates based on Darcy's law are highly uncertain because of the high variability of

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hydraulic conductivity (several orders of magnitude). The applicability of laboratory-measured hydraulic conductivities at the field scale is also questionable. This technique can be applied to large regions (~1 to ≥10,000 km^2). The time periods represented by the recharge estimates range from years to hundreds of years.

Tracer techniques

Groundwater dating

Historical tracers or event markers such as bomb-pulse tritium (^3H) are used in both unsaturated and saturated zones to estimate recharge. Tritium has been used widely in the past (Egboka et al. 1983; Robertson and Cherry 1989); however, bomb-pulse ^3H concentrations have been greatly reduced as a result of radioactive decay. In the southern hemisphere, where ^3H concentrations in precipitation were an order of magnitude lower than in the northern hemisphere (Allison and Hughes 1977), it is often difficult to distinguish bomb-pulse ^3H from current ^3H concentrations in precipitation. The use of ^3H to date groundwater is generally being replaced by the use of tracers such as chlorofluorocarbons (CFCs) and tritium/helium-3 ($^3H/^3He$). These gas tracers can be used only as water tracers in the saturated zone, where they can no longer exchange with the atmosphere. The first appearance of tracers such as CFCs or $^3H/^3He$ can be used to estimate recharge rates where flow is primarily vertical, as in recharge areas near groundwater divides. Recharge rates can also be determined by estimating ages of groundwater. Age is defined as the time since water entered the saturated zone. Groundwater ages are readily estimated from CFCs by comparing CFC concentrations in groundwater with those in precipitation (Fig. 2). The age of the groundwater, *t*, is calculated from $^3H/^3He$ data using the following equation:

$$t = -\frac{1}{\lambda}\ln\left[1 + \frac{^3\mathrm{He}_{trit}}{^3\mathrm{H}}\right] \tag{11}$$

where λ is the decay constant (ln $2/t^{1/2}$), $t^{1/2}$ is the ^3H half life (12.43 years), and $^3He_{trit}$ is tritiogenic ^3He. Use of this equation assumes that the system is closed (does not allow ^3He to escape) and is characterized by piston flow (no hydrodynamic dispersion).

In unconfined porous-media aquifers, groundwater ages increase with depth, the rate of which depends on aquifer geometry, porosity, and recharge rate (Cook and Bohlke 2000). The vertical groundwater velocity decreases with depth to zero at the lower boundary of the aquifer. The age increases linearly with depth near the water table and nonlinearly at greater depths. Near the water table, the influence of the aquifer geometry is greatly reduced. The recharge rate can be determined by dating water at several points in a vertical profile, calculating the groundwater velocity by inverting the age gradient, extrapolating the velocity to the water table if it is not measured near the water table (Cook and Solomon 1997), and multiplying the velocity by the porosity for the depth interval [similar to Eq. (7)].

The range of recharge rates that can be estimated by using groundwater dating depends on the ranges of ages that can be determined. CFCs and $^3H/^3He$ are used to determine groundwater ages up to approximately 50 years, with a precision of 2 to 3 years (Cook and Solomon 1997). Radioactive decay of ^{14}C can be used to estimate groundwater ages of 200 to 20,000 years. The estimated recharge rates are average rates over the time period represented by the groundwater age. Groundwater recharge rates of 100 to 1,000 mm/year have been determined using $^3H/^3He$ (Schlosser et al. 1989; Solomon et al. 1995) and CFCs (Dunkle et al. 1993; Cook et al. 1998). Recharge rates much less than 30 mm/year are difficult to determine accurately using these tracers because of problems associated with diffusion of ^3He into the unsaturated zone (Cook and Solomon 1997) and the difficulty of obtaining discrete samples near the water table. Dispersive mixing can result in ±50% uncertainty in $^3H/^3He$ ages prior to 1970, when ^3H input varied markedly during the bomb pulse (Solomon and Sudicky 1991). Recharge rates of 0.1 to 100 mm/year have been determined using ^{14}C (Vogel 1967; Leaney and Allison 1986; Verhagen 1992), although diffusional transport at very low recharge rates probably means a lower limit to the method of about 1 mm/year (Walker and Cook 1991). The method is most accurate where piezometers have been completed with relatively short well screens. Recharge rates calculated using groundwater dating spatially integrate recharge over an area upgradient from the measurement point. Therefore, spatial scales can range from local (decameter scale) if samples are collected near a groundwater divide (Szabo et al. 1996) to regional (kilometer scale; Pearson and White 1967).

Horizontal flow velocities can be estimated from radioactive decay of ^{14}C or ^{36}Cl in a confined aquifer. These data can be used to estimate recharge rates (*R*):

$$R = \nu nA/S \tag{12}$$

where *v* is velocity, *n* is porosity, *A* is the cross-sectional area of the confined aquifer where the velocity is determined, and *S* is the surface area of the recharge zone. If necessary, corrections should be made for any leakage to or from the confined aquifer. Using this method, recharge rates of ~50 mm/year were estimated for the Carrizo aquifer in Texas, USA, on the basis of ^{14}C data from Pearson and White (1967), assuming no vertical leakage to the confined aquifer.

There are various restrictions to the use of these tracers to date groundwater. CFCs can be used only in rural areas not affected by septic tanks because of contamination associated with industrial and residential areas, whereas $^3H/^3He$ can be used in contaminated and uncontaminated areas. The concentrations of $^3H/^3He$ and CFCs at the water table are assumed to be equal to those in the atmosphere. Any difference in concentrations would result in errors in the estimated ages. Cook and Solomon (1995) conclude that errors are negligible if the unsaturated-zone thickness is ≤10 m. Other issues that need to be considered when using these tracers include the effect

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of excess air for both $^{3}H/^{3}He$ and CFCs and the effect of recharge temperature, sorption, and degradation on CFCs. Sampling for $^{3}H/^{3}He$ and CFCs is complex, and analysis is relatively expensive.

Environmental tracers – chloride
The chloride mass-balance (CMB) approach can be used in the unsaturated and saturated zones to estimate groundwater recharge. The CMB approach was originally applied in the saturated zone by Eriksson and Khunakasem (1969) to estimate recharge rates (30 to 326 mm/year) on the Coastal Plain of Israel. Recharge rates estimated from groundwater Cl concentrations range from 0 to 8 mm/year in South Africa (Sami and Hughes 1996), 11 mm/year in the Southern High Plains, USA (Wood and Sanford 1995), 13 to 100 mm/year in southwestern Australia (Johnston 1987), and 150 to 660 mm/year in northeastern Australia (Cook et al. 2001). Recharge rates based on groundwater Cl by Johnston (1987) are as much as two orders of magnitude greater than those based on Cl in the unsaturated-zone pore water. The discrepancy between the two rates is attributed to preferential flow. Slightly higher recharge rates can be estimated using Cl in groundwater than in soil water because extraction of water from the soil generally requires additional dilution. The CMB approach spatially integrates recharge over areas upgradient from the measurement point. Spatial scales range from ~200 m (G.A. Harrington, CSIRO, personal communication, 2000) to several kilometers (Wood and Sanford 1995). The time scales range from years to thousands of years.

Numerical modeling

Early estimates of groundwater recharge were made by graphical analysis of flow nets for both unconfined and confined aquifers (Cedegren 1989); however, this approach has largely been replaced by groundwater flow models. Groundwater-model calibration or inversion is used to predict recharge rates from information on hydraulic heads, hydraulic conductivity, and other parameters (Sanford 2002, this volume). Because recharge and hydraulic conductivity are often highly correlated, model inversion using hydraulic-head data only is limited to estimating the ratio of recharge to hydraulic conductivity (Fig. 5). The reliability of the recharge estimates depends on the accuracy of the hydraulic-conductivity data. Because hydraulic conductivity ranges over several orders of magnitude, estimation of recharge rates using model calibration may not be very accurate. The estimated recharge is often nonunique because the same distribution of hydraulic heads can be produced with a range of recharge rates, as long as the ratio of recharge to hydraulic conductivity remains the same (Fig. 5). Recharge and hydraulic conductivity are fixed for steady-state simulations, whereas transient simulations reproduce temporal variations in recharge that further constrain the recharge estimates.

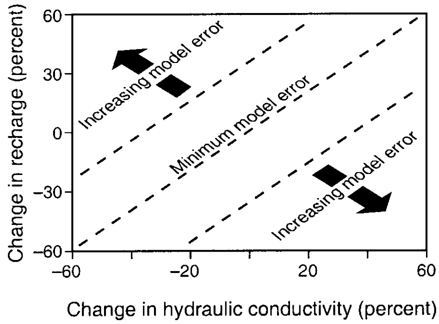

Fig. 5 Relation between change in recharge and change in hydraulic conductivity. Groundwater-model calibration using hydraulic heads only provides information on the ratio of recharge to hydraulic conductivity. Model error can be minimized using a wide range of recharge and hydraulic-conductivity values if the ratio between recharge and hydraulic conductivity is constant. (Modified from Luckey et al. 1986)

Recent studies have used joint inversions that combine hydraulic heads and groundwater ages to further constrain inverse modeling of recharge (Reilly et al. 1994; Szabo et al. 1996; Portniaguine and Solomon 1998). Manual trial and error procedures or automated procedures, which use nonlinear regression between measured and simulated data, are used. Whereas hydraulic heads are sensitive to the ratio of recharge to hydraulic conductivity, groundwater ages are sensitive to the ratio of recharge to porosity (Portniaguine and Solomon 1998). Use of both head and age data provides constraints on recharge, hydraulic conductivity, and porosity. Because these three parameters are highly correlated, a unique solution requires information on one of them. Porosity generally varies much less than recharge or hydraulic conductivity; therefore, porosity can be estimated for the system (Portniaguine and Solomon 1998). Automated inversions provide information on the nonuniqueness of the solutions. Joint inversions were used to estimate zonal recharge rates that range from 10 to 2,000 mm/year at a site in the US (Portniaguine and Solomon 1998). Spatial scales are generally much greater than those for unsaturated-zone modeling and range from several meters squared to 1,000,000 km^2 or greater. Time scales generally range from days to 100 years because of the availability of hydrologic data.

Mixing-cell models (compartment models, lumped models, and black-box models) have been used to delineate sources of recharge and estimate recharge rates on the basis of chemical and isotopic data. The hydrologic system is treated as a series of interconnected cells or compartments, which are fully mixed internally. Each cell can have more than one input and output. Fluxes between cells are varied iteratively until a good fit between measured and simulated hydrologic, chemical, and/or isotopic data is obtained. An estimate of the mean recharge to the system is calculated by dividing the volume of the system by the mean residence time. Allison and Hughes (1975) used a mixing-cell model based on conservation of ^{3}H to assess the relative contribution of

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lateral inflow from a mountain range and recharge through the unsaturated zone. Yurtsever and Payne (1986) used a nine-compartment mixing-cell model having shallow, intermediate, and deep reservoirs to reproduce discharge and ^{3}H concentrations in a large karst system in southern Turkey. Hydrochemical and isotopic data were used by Adar et al. (1992) to define a mixing-cell model in the Arava Valley, Israel. Multivariate cluster analysis was used to define recharge sources and delineate mixing cells. Mass-balance equations were developed for each cell on the basis of conservation of water, dissolved chemical species, and isotopes. These equations were solved simultaneously for unknown recharge rates into the various cells.

Comparison of Range of Recharge Rates and Spatial and Temporal Scales of the Various Techniques

The various techniques for quantifying recharge differ in the range of recharge rates that they estimate (Fig. 6) and the space and time scales they represent (Figs. 7 and 8). The range of recharge rates estimated with a particular technique should be evaluated on a site-specific basis by conducting detailed uncertainty analyses that include uncertainties in the conceptual model and in the input and output parameters. The ranges shown in Fig. 6 are based primarily on measured ranges from the literature discussed previously and provide some indication of possible ranges for each technique. Numerical-modeling approaches can generally be used to estimate any range in recharge rates; however, the reliability of these recharge estimates should be evaluated in terms of the uncertainties in the model parameters. Some techniques have definite restrictions on the recharge rates that they can estimate. Surface-applied and historical tracers in the unsaturated zone require a minimum recharge rate to transport the tracers through the root zone. In addition, historical tracers in the saturated zone, such as ^{3}H/^{3}He, require a minimum recharge rate of ~30 mm/year to confine the ^{3}He. Use of environmental tracers, such as Cl, is one of the few techniques that can estimate very low recharge rates and is generally more accurate in this range. The upper range of recharge rates shown for the various techniques (Fig. 6) generally reflects the measured rates in the literature and may not reflect a true upper limit for the technique. Upper recharge limits for applied and historical tracers in the unsaturated zone may reflect limitations of the thickness of the unsaturated zone or the ability to locate these tracers at depth. In many cases where recharge rates are high, the unsaturated zone is not very thick. Analytical uncertainties in Cl measurements and uncertainties in Cl inputs restrict the upper range of recharge rates that can be estimated with the CMB technique.

The surface areas represented by the recharge estimates vary markedly among the different techniques (Fig. 7). In general, many techniques based on unsaturated-zone data provide point estimates or represent relatively small areas, whereas some of the surface-water techniques and many of the groundwater approaches represent much larger areas. Surface-water techniques such as seepage meters and heat tracers provide point estimates of recharge. Watershed modeling can be used to estimate recharge rates over a large range of scales, as shown by previous studies (up to 500,000 km^2; Arnold et al. 2000). Although many of the unsaturated-zone techniques provide point estimates of recharge, such estimates may represent much larger areas, as shown by comparisons of point estimates from several basins (Phillips 1994) and by relating point data to geomorphic settings (Scanlon et al. 1999a) in the southwestern US. In addition, electromagnetic induction has proved to be a useful tool to regionalize point estimates (Cook et al. 1992; Scanlon et al. 1999b). Saturated-zone studies spatially integrate recharge fluxes over large areas. This spatial integration is important for water-resource assessments where large-scale estimates of recharge are often required. Using tracers to date water near groundwater divides may provide estimates of local recharge rates.

The time scales represented by recharge rates are variable (Fig. 8). Many surface-water approaches provide recharge estimates on short time scales (event scales), and estimates over longer time scales are ob-

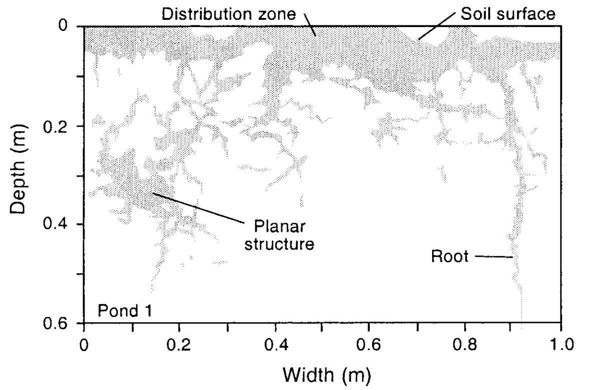

Fig. 6 Preferential flow shown by FD&C (Food Drug and Cosmetic) blue dye in structured soils (Scanlon and Goldsmith 1997)

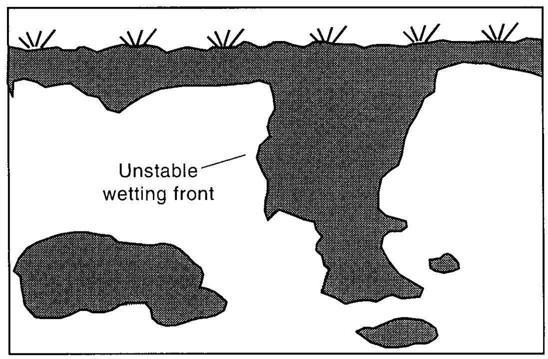

Fig. 7 Unstable flow in the shallow subsurface at a site in the Netherlands (Hendrickx and Dekker 1991)

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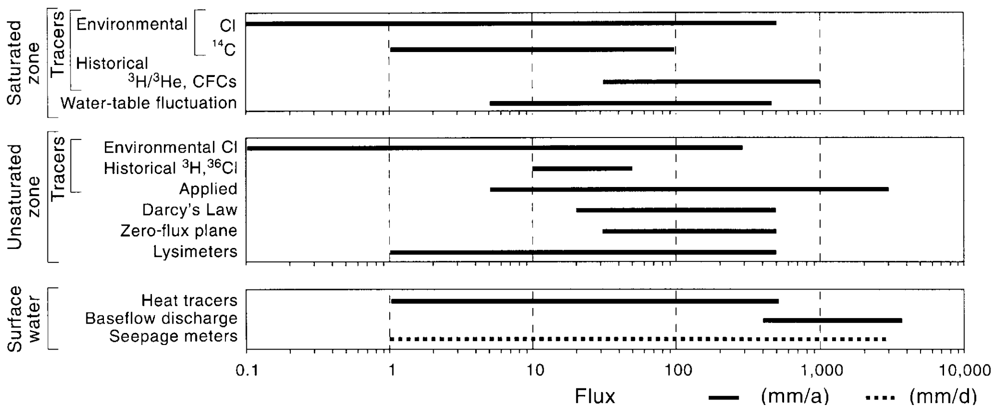

Fig. 8 Range of fluxes that can be estimated using various techniques

tained by summing those from individual events. Unsaturated-zone techniques, such as lysimeters, zero-flux plane, and applied tracers, and saturated-zone techniques, such as water-table fluctuations, provide recharge estimates on event time scales also. These techniques are restricted to providing recharge estimates for the length of the monitoring record. Numerical-modeling approaches can be used to predict recharge over any time scale; however, recharge estimation based on climatic data is generally restricted to about 100 years. The only techniques that can provide integrated, long-term estimates of recharge are tracers such as ^{36}Cl, ^{3}H, $^{3}H/^{3}He$, CFCs, ^{14}C, and Cl. Tracers are very useful for estimating net recharge over long time periods but generally do not provide detailed time series information on variations in recharge.

Application of Multiple Techniques

Because of uncertainties associated with each approach for estimating recharge, the use of many different approaches is recommended to constrain the recharge estimates. In many cases, different approaches complement each other and help refine the conceptual model of recharge processes. Examples of multiple approaches include the use of various tracers in unsaturated zones (e.g., Cl, ^{36}Cl, and ^{3}H; Scanlon 1992; Cook et al. 1994; Nativ et al. 1995; Tyler et al. 1996; Prych 1998) and saturated zones (CFC-11, CFC-12, $^{3}H/^{3}He$; Ekwurzel et al. 1994; Szabo et al. 1996). Other studies have combined soil physics and environmental tracers (Scanlon et al. 1999a) and also numerical modeling (Scanlon and Milly 1994; Fayer et al. 1996).

Ideally, as many different approaches as possible should be used to estimate recharge. Techniques based on data from surface water and unsaturated and saturated zones can also be combined. Sophocleous (1991) shows how unsaturated-zone water-balance monitoring could be combined with water-table fluctuations to increase the reliability of recharge estimates. Some studies have used catchment-scale surface-water models to provide estimates of recharge to groundwater models (Davies-Smith et al. 1988; Handman et al. 1990); however, such an approach assumes that no time lag occurs between infiltration and groundwater recharge. More recently, surface-water and groundwater models have been integrated, such as the SWAT and MODFLOW codes by Sophocleous and Perkins (2000). This integrated model provides a framework for the total system that can be used to check continuity and better constrain model parameters. Parameter optimization is conducted by calibrating against multiple targets, such as groundwater levels, streamflow data, and other data, that should result in more reliable results than obtained when using watershed or groundwater models separately. In this integrated model, recharge is constrained by an overall water budget for the surface-water system, and stream–aquifer interactions are constrained by the watershed model.

Recharge Estimation Related to Aquifer Vulnerability to Contamination

Spatial variability in recharge and preferential flow through the unsaturated zone are critical issues for contaminant transport and for assessing aquifer vulnerability to contamination. The occurrence of focused flow and preferential flow should be identified because they greatly reduce transport times through the unsaturated zone. Quantifying recharge in areas subjected to preferential flow is quite complicated. Whereas preferential flow may be demonstrated using unsaturated-zone tracer experiments, recharge rates in areas of preferential flow should be based on saturated-zone data because the saturated zone generally integrates contributions from preferred pathways.

Preferential flow includes macropore flow and unstable flow (Steenhuis et al. 1994). Macropore flow refers to flow along noncapillary-size openings such as those in fractured chalk (Nativ et al. 1995), fractured basalts, karst, and structured fine-grained sediments (Flury et al. 1994). Macropore flow occurs mostly in humid sites that have high precipitation (Gish and Shirmohammadi 1991) or in arid settings subjected to ponding. Applied tracer experiments using organic dyes are often conducted at or

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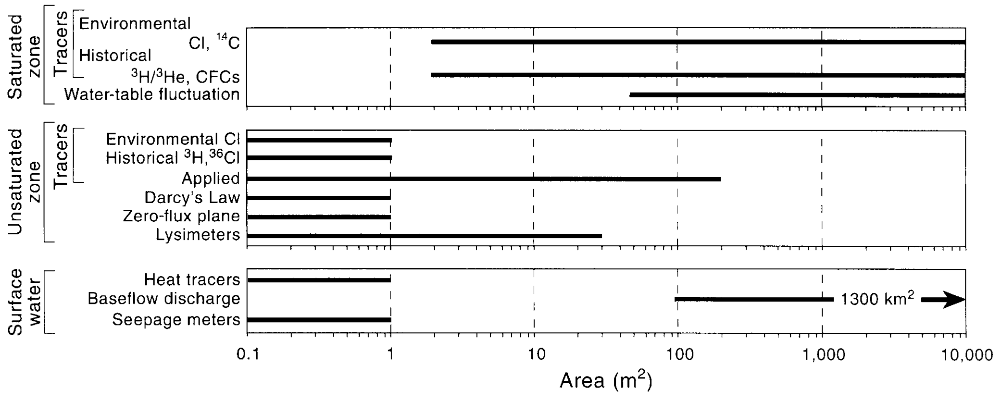

Fig. 9 Spatial scales represented by various techniques for estimating recharge. Point-scale estimates are represented by the range of 0 to 1 m

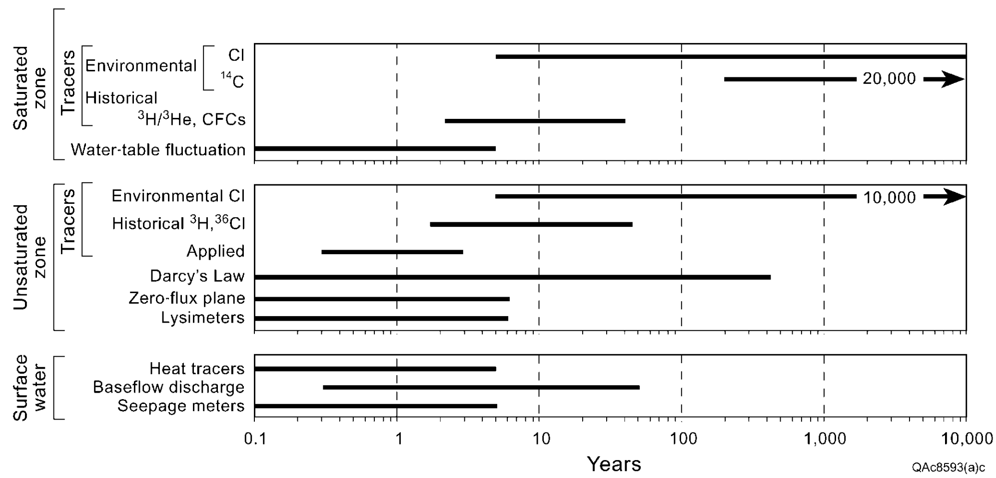

Fig. 10 Time periods represented by recharge rates estimated using various techniques. Time periods for unsaturated- and saturated-zone tracers may extend beyond the range shown

near land surface to demonstrate preferential flow (Fig. 9). Unstable wetting fronts have been reported in several field sites (Glass et al. 1988; Hendrickx and Dekker 1991; Hendrickx et al. 1993; Fig. 10). Important factors in the development of unstable flow in porous media include layering of sediment (Hillel and Baker 1988; Glass et al. 1989), air entrapment (Glass et al. 1990), and water repellency (Hendrickx and Dekker 1991; Dekker and Ritsema 1994).

The importance of preferential flow in assessing aquifer vulnerability to contamination depends on the type of contaminants. Preferential flow is much more important for contaminants that exceed health standards in the parts-per-billion range (e.g., pesticides) than the parts-per-million range (e.g., nitrate) (Steenhuis and Parlange 1991). Nitrate contamination requires movement of the bulk of the pore water, which is much greater than the generally smaller water volume transported along preferred pathways. Arrival of the first 1% of the chemical at the groundwater is more readily accommodated by preferential flow than is the transport of the bulk of the mass.

Estimation of recharge in areas of preferential flow is difficult. Preferential flow may be evidenced by multiple peaks of thermonuclear tracers such as ^{36}Cl and ^{3}H in the unsaturated zone. For calculating recharge rates, the tracer center of mass rather than the tracer peak should be used, because the latter does not conserve mass. Deep penetration of ^{3}H (~20 m) beneath clay-rich playas in the Southern High Plains, USA, is attributed to preferential flow along cracks, as shown by blue-dye tracer experiments at the surface (Scanlon and Goldsmith 1997). Migration of ^{3}H to a depth of 12 m in fractured chalk in Israel resulted from preferential flow along fractures (Nativ et al. 1995). Water fluxes as great as 110 mm/year were calculated from the ^{3}H data. Chloride concentrations in the unsaturated zone were insensitive to preferential flow. High ^{36}Cl/Cl ratios have been observed to depths of 440 m at Yucca Mountain (Liu et al. 1995; Flint et al. 2002, this volume), suggesting that preferential flow occurs along fractures. Recharge rates could not be calculated from these data because of uncertainties in flow-path lengths and average water contents along the flow paths. Estimation of recharge in areas of preferential flow may be improved by focusing on areas where flow from preferred pathways is integrated, such as in perched water in unsaturated systems or in groundwater. Chloride concentrations that are much lower in groundwater than in the unsaturated zone reflect the contribution of preferred pathways in structured zones at a site in Western Australia (Johnston 1987).

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Table 1 Appropriate techniques for estimating recharge in regions with arid, semiarid, and humid climates

Hydrologic zone	Technique	
	Arid and semiarid climates	Humid climate
Surface water	Channel water budget	Channel water budget
	Seepage meters	Seepage meters
	Heat tracers	Baseflow discharge
	Isotopic tracers	Isotopic tracers
	Watershed modeling	Watershed modeling
Unsaturated zone	Lysimeters	Lysimeters
	Zero-flux plane	Zero-flux plane
	Darcy's law	Darcy's law
	Tracers [historical (^{36}Cl, ^{3}H), environmental (Cl)]	Tracers (applied)
	Numerical modeling	Numerical modeling
Saturated zone	–	Water-table fluctuations
	–	Darcy's law
	Tracers [historical (CFCs, $^{3}H/^{3}He$), environmental (Cl, ^{14}C)]	Tracers [historical (CFCs, $^{3}H/^{3}He$)]
	Numerical modeling	Numerical modeling

General Approach to Choosing a Technique for Recharge Estimation

The goal of the recharge study plays an important part in determining appropriate techniques for quantifying recharge because the objective determines the space/time scales required. Evaluation of water resources generally requires techniques that provide regional estimates of recharge, whereas spatial variability of recharge and preferential flow are critical to contaminant transport or determination of aquifer vulnerability to contamination. Estimates of future recharge generally require numerical-modeling approaches. Background information on climate; geomorphology, including topography; vegetation, irrigation, soil type, and permeability; physiographic setting; subsurface geology; and hydrology (water-table depth, gaining vs. losing streams) can be used to develop a conceptual model of recharge at a site and to delineate recharge sources. Sources and mechanisms of recharge may also dictate the techniques to be used to quantify recharge. Surface-water sources require techniques such as channel water budgets, heat tracers, and water-table fluctuations. Climatic regions, e.g., arid vs. humid, have fundamental differences in recharge that may require different approaches (Table 1). Surface-water and saturated-zone techniques are more widely used in humid regions, whereas unsaturated-zone techniques are widely used in arid and semiarid regions. Watershed-modeling approaches may be more accurate in humid regions, where perennial surface-water flow can be used for model calibration. Although historical tracers can be used in the unsaturated zone in humid regions, their use is limited because of generally thin unsaturated zones and the ease of using such tracers in the saturated zone. Water-table fluctuations and Darcy's law could also be used in arid and semiarid regions where water tables are shallow.

The space and time scales of the various techniques also affect the choice of technique used. Surface-water and groundwater approaches provide regional estimates of recharge, whereas unsaturated-zone techniques generally provide estimates at points or small scales (Fig. 7). However, point estimates may represent much larger areas. Time scales of different approaches are variable (Fig. 8). Surface-water approaches provide recharge estimates on the time scale of events or longer by summing up events, whereas unsaturated-zone and saturated-zone tracers provide recharge estimates over long periods (≤ thousands of years), which may be important for studies related to disposal of radioactive waste. Long-term recharge estimates provided by tracers are not always advantageous, however, because they do not provide detailed information on variations in recharge over time.

The estimated recharge rate at a site may determine the most appropriate procedures for quantifying recharge because different techniques measure recharge over different ranges (Fig. 6). Initial estimates of recharge rates can be obtained by analogy to other sites having similar climatic and geomorphic settings or by numerical simulations that use existing climatic data. Water-table depth may also be important because some techniques (e.g., $^{3}H/^{3}He$, CFCs) are used only in areas of relatively shallow (≤10 m) water tables.

The required accuracy and reliability of recharge estimates influence the choice of technique. Saturated-zone techniques generally provide recharge estimates that are more reliable because they estimate actual recharge, whereas surface-water and unsaturated-zone techniques estimate potential recharge. With the exception of lysimeters, water-budget approaches are generally less accurate in semiarid and arid regions than in humid regions, because in dry areas recharge constitutes a smaller fraction of the water budget and the recharge term accumulates the errors in all the other terms of the water-budget equation (Gee and Hillel 1988). Techniques that require hydraulic-conductivity data, such as Darcy methods and unsaturated- and saturated-zone models, are inherently inaccurate because hydraulic conductivity can vary over several orders of magnitude. Various sources of uncer-

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tainty include those related to measurement of hydraulic conductivity, applicability of data at the measurement scale (laboratory vs. field scale) to the scale of recharge calculation, and spatial variability in hydraulic conductivity. Uncertainties in hydraulic conductivity are even greater in unsaturated systems than in saturated systems because of nonlinear relationships between hydraulic conductivity and water content. These uncertainties in hydraulic conductivity could readily result in order-of-magnitude uncertainties in recharge estimates. Uncertainties in recharge estimates based on tracer data include those associated with measurement of tracer concentrations, estimated inputs of tracers, and assumptions about tracer transport processes. These uncertainties are generally less than those associated with water-budget approaches or methods that use hydraulic-conductivity data.

Additional concerns that can influence the choice of technique include time and expense constraints. If recharge estimates have to be developed in a short time (months), then techniques based on long-term monitoring (several years) cannot be used. Tracer techniques may be more suitable because they generally require only one-time sampling and may represent long time periods. Cost of the various approaches may also be a consideration. Although sampling and analysis of chemical and isotopic tracers are usually considered to be expensive, one-time sampling is generally sufficient; therefore, the costs may be less than those associated with long-term monitoring that require monitoring equipment and continual collection and analysis of data.

The process of recharge estimation is iterative and involves continual refinement of recharge rates as additional data are collected. Multiple techniques should be used to estimate recharge because of uncertainties associated with each approach.

Summary

A wide variety of techniques is available for quantifying groundwater recharge; however, choosing appropriate techniques for a particular site is not straightforward. Attributes of the various techniques, including the range, space/time scales, and reliability of recharge estimates are described because they determine which approach can be used for different studies. The goal of the recharge study is important in choosing a technique because it affects the required space/time scales of the recharge estimates. Typical study goals include water-resource evaluation, which requires recharge estimates over large space and long time scales; and evaluation of aquifer vulnerability to contamination, which requires recharge estimates over small space scales, including preferential flow and variable time scales. Most techniques based on data from unsaturated-zone approaches provide recharge over relatively small spatial scales, whereas those based on data from surface-water and groundwater approaches provide information over a wide range in spatial scales. The range of recharge rates that can be estimated by different techniques varies and determines, in part, which technique to apply. The reliability of the recharge estimates is also quite variable. Surface-water and unsaturated-zone approaches usually provide estimates of potential recharge, whereas groundwater techniques generally provide information on actual recharge, because water has reached the water table. Uncertainties associated with techniques that require hydraulic-conductivity data are generally high because hydraulic conductivity can vary over several orders of magnitude. Water-budget approaches may also have fairly high uncertainties because errors in all terms accumulate in the recharge rates; however, these errors can be minimized by using small time steps. Tracer techniques may be more accurate than other techniques. Models play a very useful role in the recharge estimation process. They are used to evaluate conceptual models, to determine sensitivity of recharge estimates to various parameters, and to predict how future changes in climate and land use may affect recharge rates. Recharge estimation is an iterative process that includes refinement of estimates as additional data are gathered. A wide variety of approaches should be applied in estimating recharge in order to reduce uncertainties and increase confidence in recharge estimates.

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