American Petroleum Institute

STEP Strategies for Today's Environmental Partnership

Estimation of Infiltration and Recharge for Environmental Site Assessment

Health and Environmental Sciences Department
Publication Number 4643
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STEP
Strategies for Today's Environmental Partnership

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- To participate with government and others in creating responsible laws, regulations and standards to safeguard the community, workplace and environment.
- To promote these principles and practices by sharing experiences and offering assistance to others who produce, handle, use, transport or dispose of similar raw materials, petroleum products and wastes.

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ABSTRACT

Chemicals released to the vadose zone may present an environmental risk if they leach into groundwater. The rates of chemical leaching and migration to groundwater are strongly controlled by the diffuse recharge that occurs over large areas of the landscape. This report reviews important processes pertaining to diffuse recharge and presents a review of current physical and chemical methods (applied to the vadose zone and groundwater) to quantify diffuse recharge. Readily available estimates of diffuse recharge are compiled and organized according to major watersheds throughout the country.

The recommended approach to quantify recharge depends upon site-specific conditions, project budget, time constraints, and the nature of the project. In some cases, sufficiently accurate estimates of recharge are available in the technical literature. In other cases, field measurements are required. The methods selected from among the many available physical and/or chemical techniques must be appropriate for the site conditions. Physical methods are based on hydraulic or geophysical data collected in the soil, groundwater, or streamflow. Chemical methods rely primarily on natural and anthropogenic tracers found in the soil or groundwater. Mathematical models of soil and groundwater flow are also valuable recharge quantification tools. For projects with limited budget and time available, recharge can be determined from methods that use a one-time sampling of data, such as collecting soil cores, analyzing chemical tracers, or obtaining existing water-level or streamflow records. Where site-specific recharge must be known accurately and time is no factor, large soil lysimeters are the best choice.

Regardless of the method to obtain recharge, there is an inherent uncertainty in the estimate or calculation. Unfortunately, the degree of uncertainty is difficult to predict a priori and depends in part on the method, conditions such as water content and

site heterogeneity, as well as the skill of the analyst. When the results from two or more different recharge analyses reasonably agree, the recharge estimate can be applied with greater confidence.

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EXECUTIVE SUMMARY

This report reviews available resources and methods for obtaining data on diffuse natural groundwater recharge for site-specific assessments (e.g., Tier 2 and Tier 3 Risk Based Corrective Action analyses). Diffuse groundwater recharge is important because it affects the rate at which residual chemicals could be leached from the soil and migrate to groundwater. Accurate recharge estimates are essential for quantifying the risk to groundwater presented by residual chemicals in the vadose zone.

Groundwater contaminant fate and transport models such as the U.S. Environmental Protection Agency's Composite Model with Transformation Products (1995) and American Petroleum Institute's VADSAT (1995) generate results that are very sensitive to the input values for infiltration and recharge. These models will overestimate receptor well concentrations when inappropriately high recharge rate estimates (e.g., an arbitrary percentage of precipitation) are used as input.

The purpose of the study documented in this report was to determine which, if any, of the available approaches could provide reliable, cost-effective estimates of recharge. The specific objectives of the study were to (1) summarize the key concepts and soil physics principles related to diffuse recharge; (2) describe, compare, and contrast estimation techniques applicable for assessing petroleum hydrocarbon or salt-impacted sites; (3) compile site-specific recharge estimates from sites throughout the country; and (4) identify areas for further research.

A literature search revealed dozens of methods, both physical and chemical, to quantify recharge in humid and dry climates. For determining site-specific recharge estimates, techniques that rely on very local measurements are more appropriate. Such techniques include lysimeters, chemical tracers, the Darcy flux and plane of zero flux methods, one dimensional soil-water balances and soil-water models, and soil temperature methods based on near-surface soil temperature gradients.

The most accurate (and often most costly) approach to estimating recharge in any climate uses soil lysimeters to collect deep percolating soil water that eventually would reach the water table. In humid climates, reasonably accurate recharge rates can be obtained from water balance calculations in the vadose zone, provided that the period of accounting is weekly or more frequently. Vadose zone chemical tracers may provide more accurate estimates in dry climates for low to moderate cost.

SUMMARY OF RECHARGE ESTIMATES

Recharge estimates were gathered from the open literature and through requests for information from U.S. Geological Survey district offices throughout the country. These data were compiled to (1) identify key studies and sources of information on recharge estimates throughout the U.S., (2) understand which techniques are being applied in various hydrogeologic and climatic settings, (3) determine the frequency with which the various techniques are being applied, and (4) develop a database for future statistical analysis.

The recharge estimates are tabulated for watersheds throughout the country. Information for each recharge study area includes climatological data, site physiography, and general soil characteristics. The recharge estimates are organized according to major surface drainage basins within geographic regions (Appendix A, Table A-1) and estimation technique (Appendix A, Table A-2). As indicated in these tables, the most frequently applied methods to quantify recharge are the soil-water balance techniques and stream flow analyses. In addition, examination of the data reveals that within any climatic region individual studies produce a wide variation in the recharge estimates. This variation may be attributed to differences in the scales of investigations.

METHODS TO QUANTIFY DIFFUSE NATURAL RECHARGE

Physical methods described in this report include both direct and indirect methods. The only direct method of measuring recharge is lysimetry, which is costly and requires lengthy data collection periods. Indirect methods described in this report include:

- Soil-water balance
- Darcy flux
- Plane of zero flux
- Soil temperature
- Electromagnetic
- Groundwater basin outflow
- Water-level fluctuation
- Stream gauging

The soil-water balance method is one of the most widely used indirect estimation techniques. However, the accuracy of this method depends upon the accuracy of estimates of its component parameters (runoff, infiltration, evapotranspiration and storage), which sometimes are poorly known or exhibit significant variability at a site. The greatest uncertainty lies in estimating evapotranspiration. Data compiled in this report indicate that recharge estimates using the soil-water balance method can vary over two orders of magnitude over large areas. However, this method may be suitable for small sites in humid or temperate regions where parameters that rely on climatic data are known to have low variability. Several vadose zone field test methods and equations needed to measure or calculate the component parameters of the water balance equation are discussed.

The Darcy flux and plane of zero flux methods provide useful estimates when resources are available to collect a sufficient number of field measurements. These methods require measurements of vadose zone moisture content and hydraulic conductivity over the seasonal range of site-specific soil moisture conditions. The soil temperature, electromagnetic, groundwater basin outflow, water-level fluctuation, and streamflow methods provide regionally averaged estimates of diffuse recharge. These methods may be most useful when the regional hydrogeology (i.e., the location of recharge areas and aquifer boundaries, storage and outflows, etc.) is well understood.

Chemical methods for estimating diffuse recharge are subdivided into those requiring measurements in either the vadose or saturated zones. Where project resources permit, chemical methods may provide better estimates of long-term recharge because they reflect recharge conditions over long periods of time.

Vadose zone chemical tracer methods track the movement of stable and radioactive isotopes. Chemical methods described in this report include the chloride mass balance method and those using tritium, chlorine-36, and stable isotopes as tracers. Chemical tracer techniques in the saturated zone determine the age of groundwater, which in turn permits calculation of groundwater travel time. Where recharge to an aquifer occurs primarily by direct local recharge, the age of the groundwater is related to local recharge. Chemical tracers used in aquifers include tritium, chlorofluorocarbons, krypton-85, carbon-14, and chlorine-36.

Mathematical models (soil-water and groundwater) are best suited to predict recharge when the physical properties of the soil and groundwater are well characterized. The water balance models typically require site-specific climatic data for precipitation, temperature and solar radiation; soil characteristics data including porosity and moisture retention characteristics; or a limited set of soil characteristics parameters, including field capacity, wilting point, saturated moisture content, and organic matter content.

The soil-water balance model HELP (Schroeder *et al.*, 1994) was reviewed as a tool for estimating recharge rate. If recharge rates are low and the period of soil-water balance accounting is too long, then HELP (and other soil-water balance models) are likely to underestimate recharge because they only roughly approximate the physics of unsaturated flow. However, at one arid-climate field site, HELP-generated recharge estimates compared favorably to independent estimates using the Darcy flux method and the chloride mass balance technique.

CHOOSING AN APPROPRIATE RECHARGE ESTIMATION TECHNIQUE

No universally acceptable methods to compute diffuse recharge can be applied to all sites. The method selected will depend on the site geology, soil characteristics, depth to the water table, vegetative cover, and climatic conditions, along with factors such as time constraints, available budget, and the importance of recharge to the success of the project. Section 4 of this report provides a guide to the appropriate selection of recharge estimation techniques based on optimal site characteristics, cost and relative accuracy.

In most cases, a limited project budget requires use of a less sophisticated technique. In such cases, one must accept some uncertainty in a site-specific recharge estimate and must attempt to understand the degree of that uncertainty. However, no comprehensive uncertainty analysis exists for the techniques described in this report.

When time and budget are limited, one can refer to estimates contained in reports by the U.S. Geological Survey or state and local water resource or geological surveys. Where site-specific measurements are required but resources are limited, one may consider an approach using a one-time data collection such as a Darcy flux analysis based on laboratory or field measurements of the deep vadose zone hydraulic properties or chemical tracer sampling of the vadose zone. If resources are available, it is desirable to use both a physical and chemical method at the site.

FUTURE RESEARCH

This study revealed several areas in which further research is needed, as outlined below:

- The reliability of some of the key methods to quantify recharge, especially in dry climates, needs to be improved. One example where considerable improvement could be achieved is in critically evaluating assumptions in the widely used chloride mass balance method.

- Methods are needed to rapidly address the nature of spatial variability in recharge over large areas. In particular, methods for quantifying the contribution of flow through macropores are needed.

- A better understanding of recharge method uncertainty in various hydrogeologic and climatic settings is needed. Additional comparison studies of the low-cost, simpler estimation techniques with more rigorous measurement systems, under a variety of conditions, would provide useful uncertainty data on recharge estimates used in risk-based corrective action and other site-modeling efforts.

- A statistical analysis of the database compiled for this study may identify a correlation between precipitation and recharge for various physiographic provinces and climatic regions.

Section 1

Introduction

Section 1
INTRODUCTION

Chemicals released into the vadose zone, from either accidental spills or wastes managed on the land, may migrate to groundwater, depending upon the nature of the release, design of the waste management facility, properties of the chemical, vadose zone characteristics, and leaching potential. Typically, quantitative analyses are required to assess whether such contaminant releases or leachate pose a threat to human health and the environment. For example, in promulgating the 1990 Toxicity Characteristic (TC) Rule, the U.S. Environmental Protection Agency (EPA) used a computer code called EPACML (EPA's Composite Model for Landfills) to estimate the potential human exposure to chemicals inappropriately disposed of in municipal landfills. Such chemical fate and transport analyses are an integral component of risk assessments required at sites remediated under the Comprehensive Environmental Response, Compensation and Liability Act (CERCLA). Similar computations are required in designing land treatment facilities for petroleum-contaminated soils and in evaluating risk-based alternatives at sites of fuel releases into the soil.

There are many analytical and numerical methods available to determine the mass flux of chemicals migrating by liquid-phase advection through the vadose zone into an aquifer. In virtually all instances, *recharge* is a key data need that must be prescribed in these calculations. Unfortunately, because of the difficulty in obtaining recharge values at a specific site, many analysts simply estimate the recharge rate based on their professional judgment, or they use model-embedded default parameters that are generally conservative for their purpose. A probabilistic approach such as a Monte Carlo analysis is sometimes used in risk assessments to address uncertainty in parameters. The probability of occurrence of a particular outcome is developed by making dozens or hundreds of calculations, each with a different set of parameters chosen at random and often without a sound basis for assessing the reasonableness of the parameters or the combinations of parameters randomly chosen. Regardless of the computational approach selected, sensitivity analyses of parameters commonly reveal that the mass flux of contaminants entering the aquifer is most strongly dependent on the recharge rate.

The objective of this report is to provide analysts a sound technical basis for selecting recharge values used in mass transport calculations, such as those required in risk-based corrective action (RBCA) assessments. The scope of the study is twofold:

- Identify relevant techniques to quantify areally distributed, diffuse, natural recharge
- Compile existing data on diffuse recharge throughout the United States

The approach to achieve these objectives was based on a search of existing literature contained in the in-house library of Daniel B. Stephens & Associates, Inc., information from prior personal research and publications on recharge, computer database searches followed by retrieval of selected documents from the University of New Mexico library, and responses to requests for information from district offices of the U.S. Geological Survey throughout the country.

Section 2 of this report summarizes the concepts and terminology associated with the analysis of recharge. The report then presents a review of methods to quantify recharge in Section 3 and highlights results of previous studies to quantify recharge within subregions of the United States in Section 4. Also, included in Section 5 is a brief discussion of considerations for selecting recharge esturation techniques. Section 6 addresses technical issues for future research. A comprehensive glossary of terms used in this report is also provided.

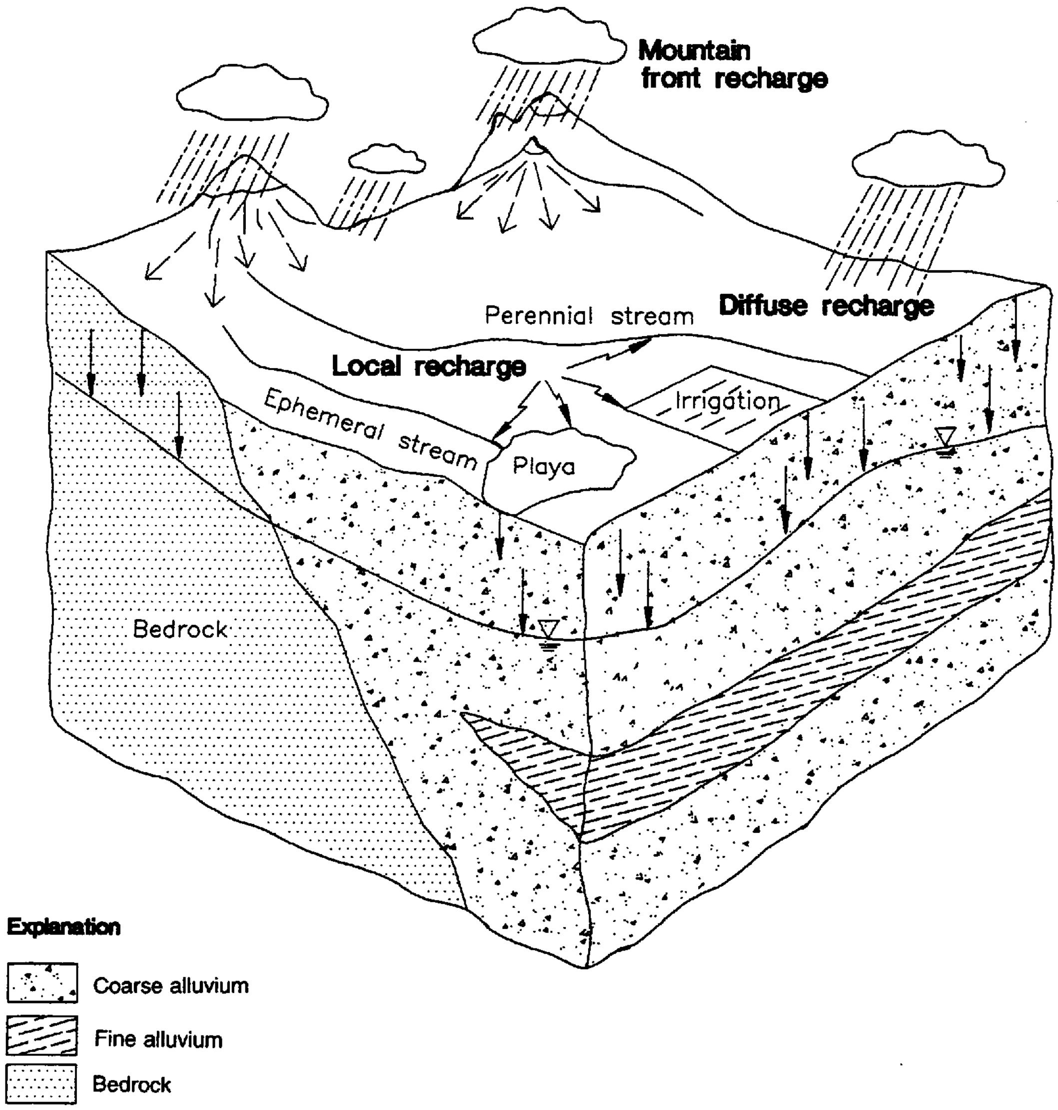
Section 2

Recharge Concepts and Terminology

Section 2
RECHARGE CONCEPTS AND TERMINOLOGY

Recharge is simply the addition of water to an aquifer. Natural recharge to groundwater commonly occurs as diffuse recharge, localized recharge, and recharge from mountain fronts (Figure 2-1). These different types of groundwater recharge are distinguished by the source of the water and by the path the water takes to enter the saturated zones within a groundwater basin. Diffuse recharge is natural recharge derived from precipitation that falls on large portions of the landscape and percolates downward through the vadose zone to the aquifer. Sometimes diffuse recharge is called deep percolation or recharge by direct precipitation. Diffuse recharge probably dominates in humid climates and is the topic of this report. Localized natural recharge occurs mainly where there is prolonged ponding within a basin, such as along a losing stream channel or a playa. In comparison to diffuse recharge, local recharge is probably the most important source of natural recharge in arid and semiarid lands. Mountain front recharge typically involves complex processes of unsaturated and saturated flow in fractured rocks, as well as infiltration along channels flowing across alluvial fans. On a large scale, mountain front recharge through fractured bedrock is primarily a diffuse recharge process, whereas infiltration from mountain streams is considered a local recharge process.

Diffuse, and to a lesser extent, local recharge are relatively large-scale processes within a groundwater basin. That is, the aquifer surface area over which the recharge occurs is usually a significant part of the basin area, perhaps on the order of one to hundreds of square kilometers. Diffuse and local recharge are also affected by processes that occur at much smaller scales, such as preferential flow in macropores, unstable flow, and the effects of spatial variability in media properties. Every process, whether at the large or small scale, ultimately involves water movement through the vadose zone. Flow in variably saturated,



Source: Stephens, 1995 (with permission)

Figure 2-1. Conceptual model of diffuse and localized recharge heterogeneous porous and fractured media further involves complex interacting processes such as flow from fractures into the matrix and vapor phase transport. To fully understand recharge processes, therefore, an appreciation for both geology and soil physics is necessary.

SUMMARY OF SOIL PHYSICS PRINCIPLES

Before beginning the discussion of recharge, it is helpful to review some of the important and relevant principles of soil physics. The review will highlight some of the common terms and parameters including water content, pressure head, soil-water characteristic curves, and hydraulic conductivity. Methods of measuring these parameters are discussed later in the report when we describe the recharge methods that require these data. We also will discuss the important processes of infiltration and redistribution which lead to recharge by water migration through the vadose zone.

Water Content

Water content is most often defined as the volume of water in a bulk volume of soil or rock. Sometimes, water content is expressed as ratio of the mass of water per dry mass of soil. In either case, water content is a dimensionless parameter, and is typically expressed as a percent. However, it is essential to recognize that the water content on a mass basis is numerically less than that on a volume basis by a factor that is the ratio of the dry bulk density to the water density. For a typical soil, the difference could be 50 to 70%. For hydrologic problems relevant to recharge, the water content is assumed to be the volumetric water content, unless otherwise indicated.

Hydraulic Head and Pressure Head

The total hydraulic head in the soil is a measure of the potential energy per unit weight of the soil water. This parameter is important because spatial differences in it determine the direction water will migrate. The change in hydraulic head with *direction is called* the hydraulic gradient. Hydraulic head includes two additive components: pressure head and elevation head. All these terms have units of length, such as centimeters, meters, or feet of water.

Pressure head reflects the energy status over a representative pore volume which includes primarily the effects of hydraulic pressure, capillary forces and adsorptive forces. The hydraulic pressure is attributed to saturated conditions which may develop in the vadose zone or to fluid pressure below the water table. The hydraulic pressure is neglected in partially saturated zones. Capillary forces occur in partially saturated media and are caused by interfacial tension between the air and water held with the soil pores. Adsorptive forces, applicable to unsaturated media, are caused by the attraction of the polar water molecules for solid surfaces. Capillary and adsorptive forces are difficult and impractical to separate, but the former is dominant in wet or sandy soils and the latter is more important in dry or clayey soils. Capillary and adsorbed forces are neglected if the soil is fully saturated and the soil is under a positive hydrostatic pressure.

Pressure head is determined with respect to the energy status at the surface of a pure water reservoir. The pressure head at this reference state is zero. Under partially saturated conditions as well as in the tension-saturated region of the capillary fringe above the water table, the pressure head is less than zero. In fact, in very dry soils pressure head may be as low as negative several tens of meters. Note that fluid gauge pressure (e.g., in units of Pascals, bars, or psi) divided by the specific weight of water (consistent units) gives the pressure head in units of length.

The elevation head is the height above an arbitrary datum of the point where pressure head is measured. The space derivative of the elevation head in the vertical direction is always negative one (–1), when the vertical coordinate axis is positive upward. The gravitational gradient is always present and is independent of the water content or the pressure head.

Sometimes pressure head is expressed as a potential energy per unit volume, and is simply called the soil-water potential or matric potential. Soil-water potential has units of pressure, such as Pascals, bars, or psi. Because of its pressure units and negative value relative to atmospheric pressure, soil-water potential is also referred to as soil suction.

Soil-Water Retention Curve

There is a relationship between the soil-water content and the pressure head which is called the soil-water retention curve. This important relationship describes the ability of the soil to absorb and release water. At the water table, the pressure head is zero and the water content is at its maximum, the porosity. As the soil dries from the maximum water content, both the water content and pressure head decrease (Figure 2-2A). The relationship between these two parameters is a characteristic of a soil. Coarse textured soils such as sand and gravel drain readily and hold much less water at a given pressure head than fine-textured silt and clay soils with large surface areas which are reluctant to give up water due to the stronger adsorptive/capillary forces.

The soil-water retention curve is slightly complicated by hysteresis, in that the curve describing drying from a fully saturated condition differs from the curve describing wetting from some low-water-content initial condition. This relationship is shown in Figure 2-2B. Hysteresis is especially important to consider in modeling periodic infiltration and exfiltration (evaporation) in uniform-textured soils. The effect of hysteresis in predictive models is to slow the downward migration of infiltration.

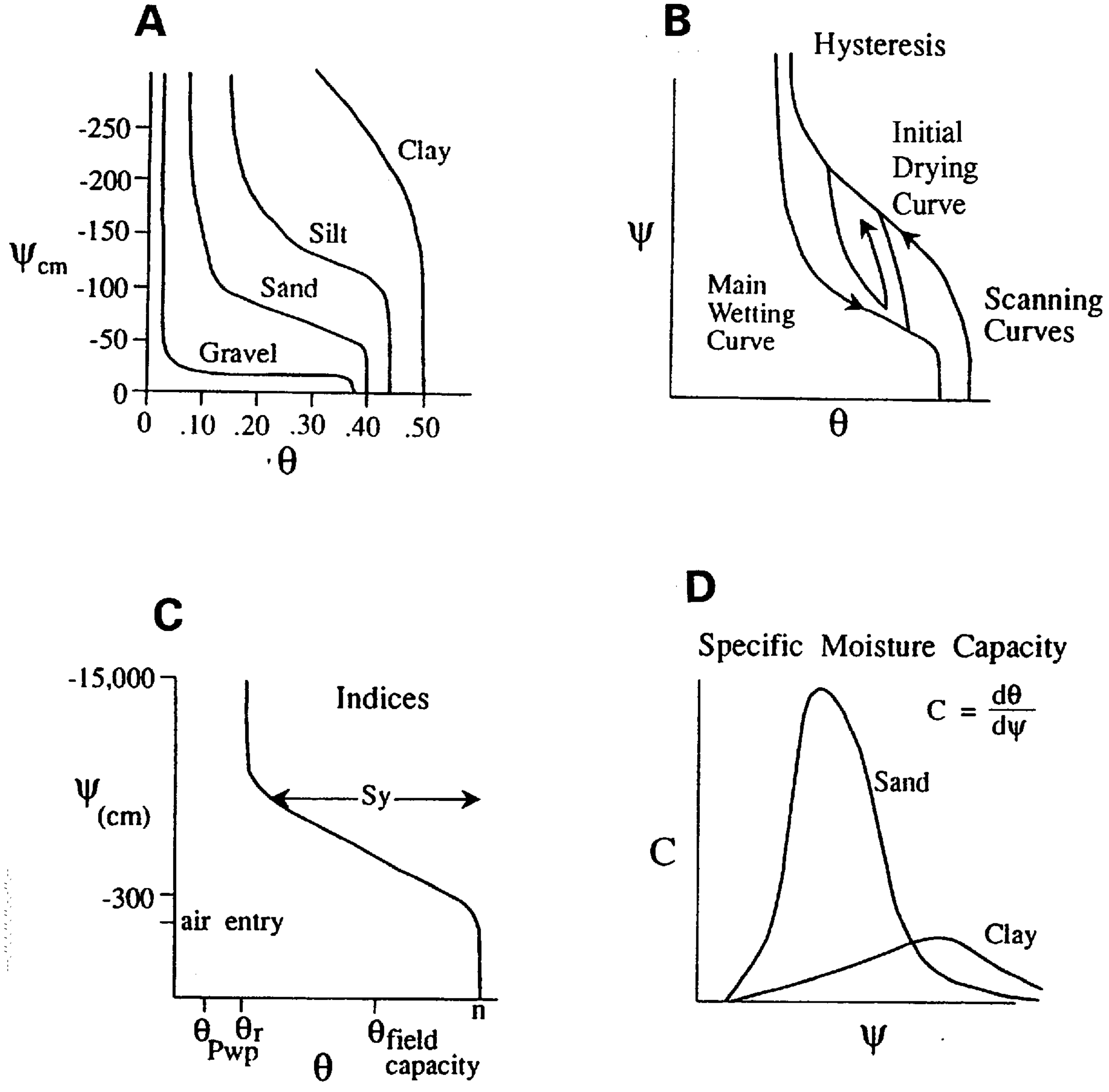


Figure 2-2. Soil-water retention: Pressure head, ψ, versus volumetric water content, θ, (A) for different soil textures, (B) showing the effect of hysteresis on a specific soil, (C) illustrating index parameters, and (D) expressed as specific moisture capacity.

On the drying curve, the pressure head at which air begins to enter the pores is called the air-entry value (Figure 2-2C). Field capacity is an imprecisely defined term which describes the water content remaining in the soil after a few days of drainage following a thorough irrigation; it has been incorrectly used to describe the water content at which water flow ceases. Field capacity is so poorly defined that its use should be avoided if possible in most quantitative analyses. The asymptotic water content approached as the soil dries is called the residual water content (Figure 2-2C). On the wetting curve, as the soil approaches saturation, air may become entrapped, so that the water content at zero pressure head may be less than the porosity (Figure 2-2B).

In essence, the soil-water retention curve describes the water storage in the vadose zone. For example, it is easy to see from the curves in Figures 2-2A, 2-2B, and 2-2C that incremental changes in pressure head yield changes in the water content. The change in water content per unit change in pressure head is called the specific water capacity, $C(\Psi)$ (Figure 2-2D). Most numerical models of saturated and unsaturated flow use specific water capacity in calculations of infiltration and recharge.

Hydraulic Conductivity

The hydraulic conductivity describes the ease with which water may be permitted to flow through the soil. When the pores are completely filled with water, the hydraulic conductivity is at its maximum value, the saturated hydraulic conductivity. As the soil drains, the largest pores begin to dewater first. During the dewatering process, the cross-sectional area of water in the pore decreases. Additionally, because the water in the pores occupies progressively finer pore spaces, the path of water transport becomes more tortuous. As a consequence of the diminishing cross-sectional area and the increasing tortuosity, the hydraulic conductivity decreases as the soil drains.

When the soil re-wets, there may be air trapped in the pores that prevents the soil from reaching full saturation. The satiated hydraulic conductivity, or maximum hydraulic conductivity achievable in the field, refers to the condition in which the soil is essentially flooded, but the field-measured hydraulic conductivity may be less than the saturated hydraulic conductivity. Such conditions are important to recognize during ponded infiltration or fluctuating water tables, for example.

Figure 2-3A illustrates the relationship between hydraulic conductivity and pressure head for a sand and for a clay loam. Note that the hydraulic conductivity of the sand near zero pressure head is much greater than for the clay loam. But as the pressure head decreases and the soil drains, the hydraulic conductivity of the clay loam decreases more gradually than that of the sand. In fact, the unsaturated hydraulic conductivity of the clay loam may exceed that of the sand at low pressure head; the same apparently counter-intuitive relationship is possible for any coarse and fine soil.

Figure 2-3B illustrates the dependence of hydraulic conductivity on percent water saturation. Note here that at all saturations, the hydraulic conductivity of the clay loam is less than that of sand. For reference, relative hydraulic conductivity (the ratio [dimensionless] of the unsaturated hydraulic conductivity to the saturated hydraulic conductivity) is shown in Figures 2-3C and 2-3D in terms of pressure head and water content, respectively. Most numerical models of saturated and unsaturated flow require hydraulic conductivity input as shown in Figure 2-3A to simulate infiltration and recharge.

Bear in mind the extreme variability of the hydraulic conductivity for a particular soil over a range of water contents likely to occur in the field, from near saturation to air dry (Figure 2-3). It would not be uncommon for a particular soil depth to exhibit a million-fold change in hydraulic conductivity as a consequence of normal wetting

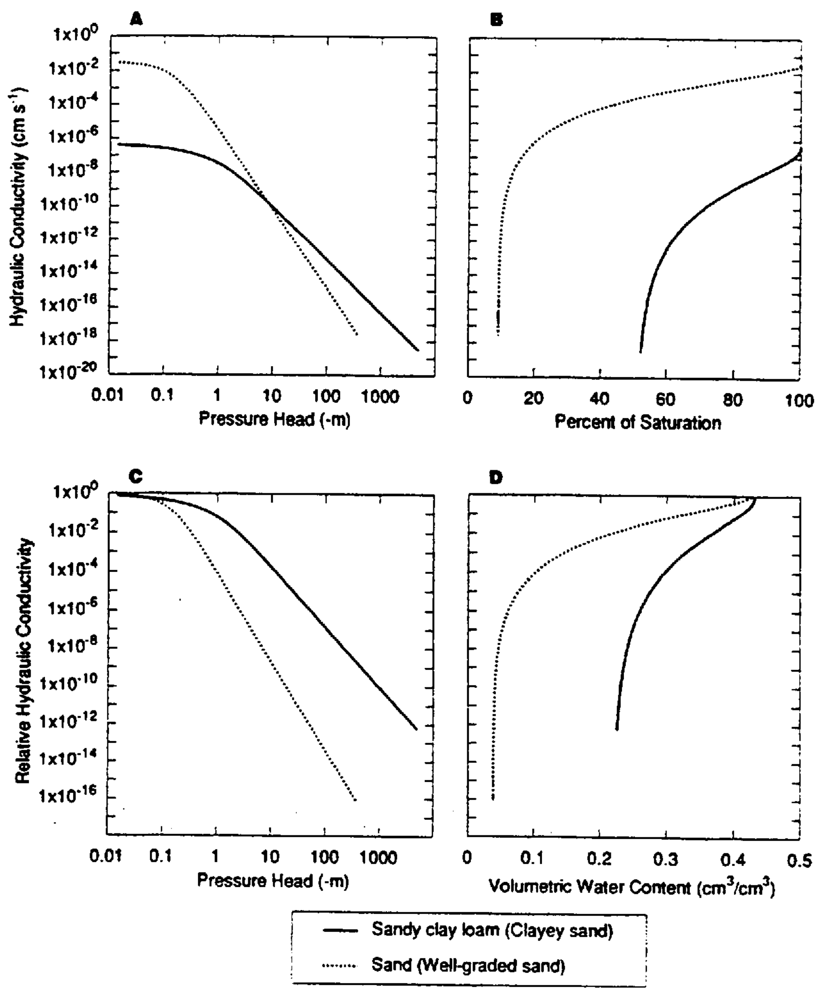


Figure 2-3. Unsaturated hydraulic properties of two soils

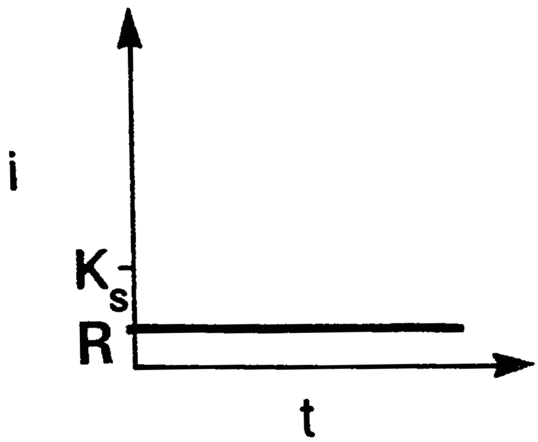
and drying in the field. Below the very dry surface layer, there is virtually always *sufficient moisture* for the hydraulic conductivity to exceed zero, even though the soil may appear to be dry. This concept, and the omnipresent gravitational gradient, are very important in addressing recharge, especially in areas of low precipitation.

<u>Infiltration, Redistribution and Recharge</u>

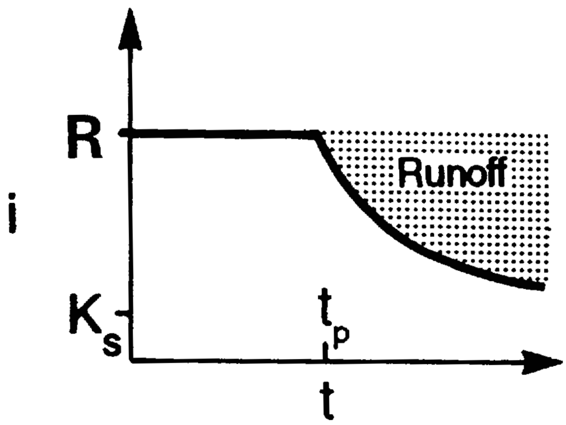
Infiltration is the volume of water that crosses the soil surface to enter the vadose zone. The infiltration rate is the time rate of change of infiltration per unit cross-sectional area of the soil. Infiltration rate is controlled by both the soil properties and the rate of water application to the soil.

If there are ponded conditions on the soil surface, the infiltration rate is greater for a more permeable soil, that is, for a soil with the greater saturated hydraulic conductivity. For a particular soil texture, the infiltration rate will be greater for an initially drier soil than for a wetter one. This is because the pressure head of the soil is lower in the dry soil, and consequently, the hydraulic gradient across the soil surface is greater when the initial water content is lower.

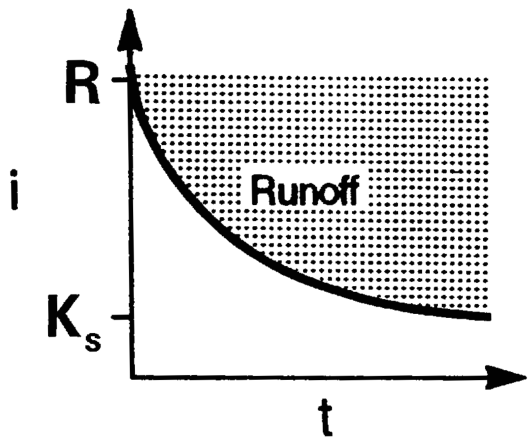
Figure 2-4 illustrates a series of infiltration rate curves. Figure 2-4A illustrates that in many field situations, the infiltration rate, i, is controlled by the rate of water application, R. That is, the soil may be so permeable (large saturated hydraulic conductivity, K_s) relative to the rainfall rate, R, that all the rainfall penetrates the soil without ponding. When ponding occurs at time t_p, there is potential for surface water runoff, depending upon the topography. After ponding occurs (Figure 2-4B, 2-4C), the infiltration rate gradually decreases to the saturated hydraulic conductivity because the hydraulic gradient between the surface and wetting front decreases as the wetting front penetrates deeper into the soil.



A All rainfall infiltrates; profile is unsaturated



B All rainfall infiltrates before time of ponding, t_p; runoff occurs after t_p; saturated zone develops



C Ponding and runoff occur at $t = 0$; profile is satiated behind the wetting front

Figure 2-4. Infiltration rate (i) dependence on rain intensity, R

For a given soil texture, increasing the depth of ponding also generally increases the infiltration rate. However, when the water table is very deep relative to the depth of ponding, the infiltration rate will approach a constant value that is independent of the ponding depth; the magnitude of this steady, ponded infiltration rate for deep water table conditions approaches the saturated, or satiated, hydraulic conductivity, because as the water table depth increases, the hydraulic head gradient approaches one.

The water content profile during infiltration is shown in Figure 2-5. The depth where the water gradient is steepest is called the wetting front. When infiltration ceases at the soil surface, the upper part of the soil profile will dry because of evaporation or water drainage to lower parts of the profile (Figure 2-5). Even after infiltration ceases at the surface, water already in the profile may move deeper. The deep migration is driven by the hydraulic gradient, that is, by the pressure head gradient between the wet and dry soil and by the downward, and ever present, gravitational gradient. Redistribution describes the process of simultaneous wetting and drying of the soil profile following infiltration (Figure 2-5). When water reaches the water table during a sustained infiltration event, or during redistribution, recharge occurs.

CONTAMINANT TRANSPORT

Contaminants in the vadose zone soils may be transported by a variety of mechanisms, including transport of volatile chemicals in the vapor phase, transport as a non-aqueous phase liquid (e.g., oil phase), transport of chemicals sorbed on migrating colloids, and transport as a dissolved phase. Our primary interest here is on chemical transport which occurs as water moves through the vadose zone to recharge an aquifer. Percolating water in the vadose zone is therefore most relevant to colloidal and dissolved phase transport.

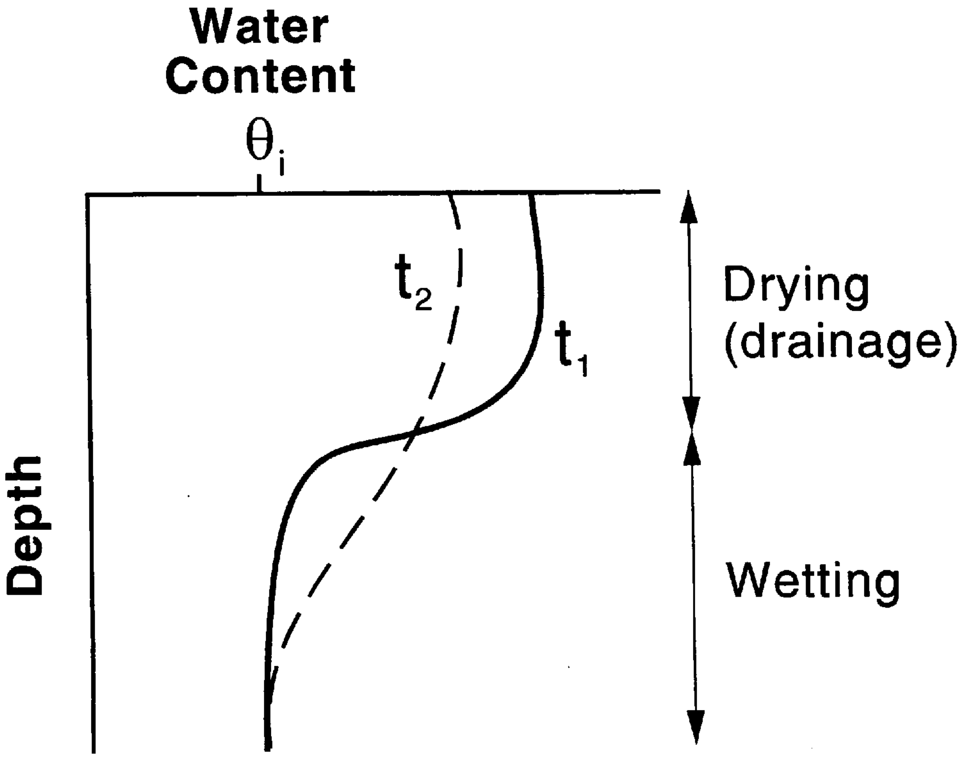


Figure 2-5. Water content profiles during infiltration (time t_1) and redistribution (time t_2). θ_i is the initial water content.

Little is known about colloidal transport in the vadose zone at present, except that this mechanism probably operates on highly sorbed particles, such as radionuclides, to transport the chemicals much farther and more rapidly than would be expected otherwise.

Chemicals in the vadose zone soils may dissolve in percolating vadose zone water and leach chemicals to the aquifer. The one-dimensional transport equation for predicting the concentration of dissolved chemicals in the vadose zone or aquifers is:

$$D \frac{\partial^2 C}{\partial x^2} - v\frac{\partial C}{\partial x} = \frac{\partial C}{\partial t} \qquad \text{(Equation 2-1)}$$

where C is concentration, D is hydrodynamic dispersion coefficient, v is mean pore water velocity, t is time and x the space coordinate.

Equation 2-1 is called the advection-dispersion equation. In this simple form, this equation does not include sorption or decay. However, it does allow for the hydrodynamic dispersion of chemicals as they mix in the pore fluids. Hydrodynamic dispersion is a consequence of the combined effects of molecular diffusion, as well as a mechanical dispersion caused by the complex pore-scale velocity distributions and tortuous flow paths. Dissolved contaminants are carried through porous media in a process called advection and at an average rate equivalent to the mean pore-water velocity. It is therefore no surprise that the most important parameter in the advection-dispersion equation for risk assessment purposes is the mean pore-water velocity. The mean pore-water velocity is usually the parameter that exhibits the greatest sensitivity in flow and transport models used for risk assessment.

In the vadose zone, the mean pore-water velocity is calculated as the Darcy velocity, q, divided by the effective mean water content θ_e:

$$v = \frac{q}{\theta_e} \qquad \text{(Equation 2-2)}$$

The effective water content may be slightly less than the measured field water content due to immobile water in pores that do not conduct flow. The Darcy velocity in the deep vadose zone, just above the water table, is the recharge rate. Therefore, measurements of recharge rate, which are the subject of this report, are highly relevant to contaminant transport. Because values for effective water content are much less than one, contaminant migration rates in the vadose zone are likely to be at least a few times to perhaps a few tens of times greater than the recharge rate.

DIFFUSE RECHARGE CONCEPTS

The vadose zone includes the geologic media between the land surface and the top of the regional water table (Figure 2-6). The process of diffuse natural recharge begins with infiltration into the vadose zone of moisture originating at the land surface as rainfall, snowmelt, or overland flow. The rate of water movement through the vadose zone is dependent upon the intensity and duration of the associated precipitation event, the hydraulic properties of the vadose zone, such as hydraulic conductivity and specific moisture capacity, and the spatial heterogeneity of the vadose zone. Although water moves primarily downward, significant lateral water movement in the vadose zone may occur due to topographic effects, heterogeneity, and anisotropy of the porous or fractured media.

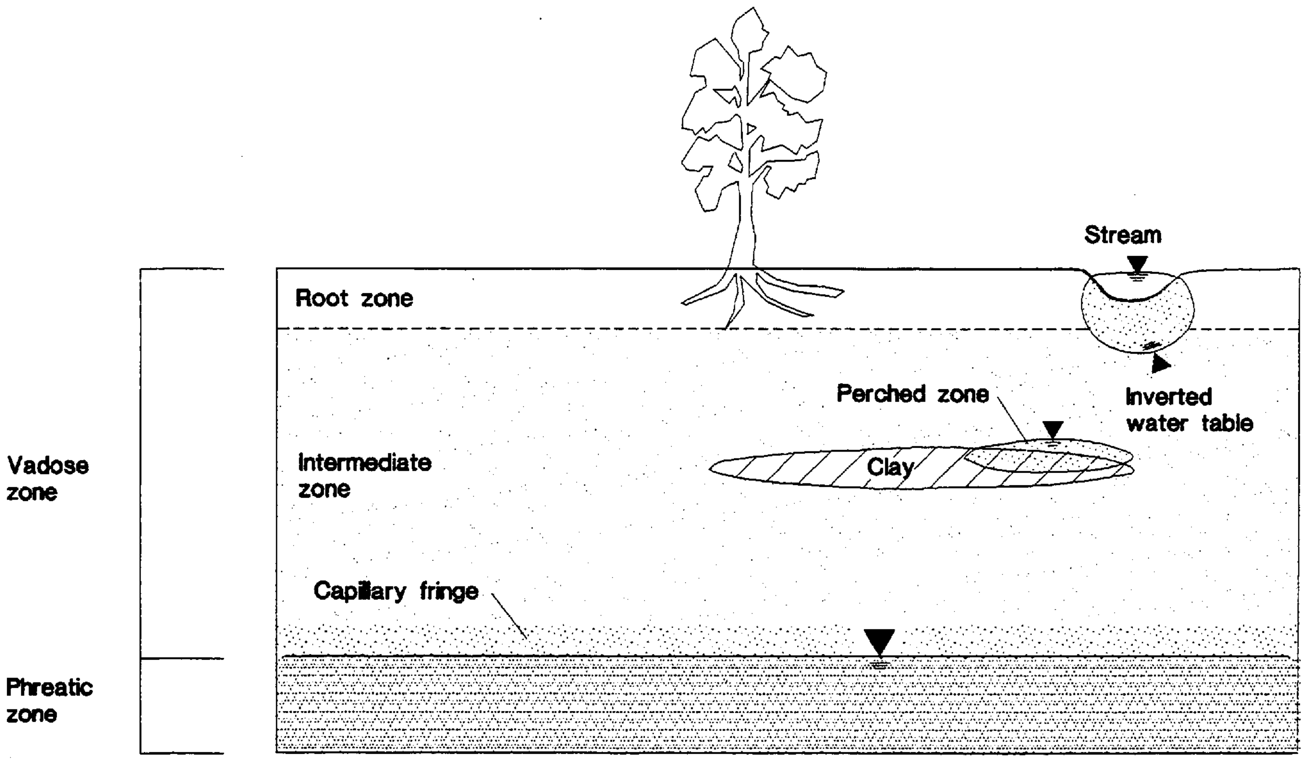


Figure 2-6. Conceptual model of the vadose zone

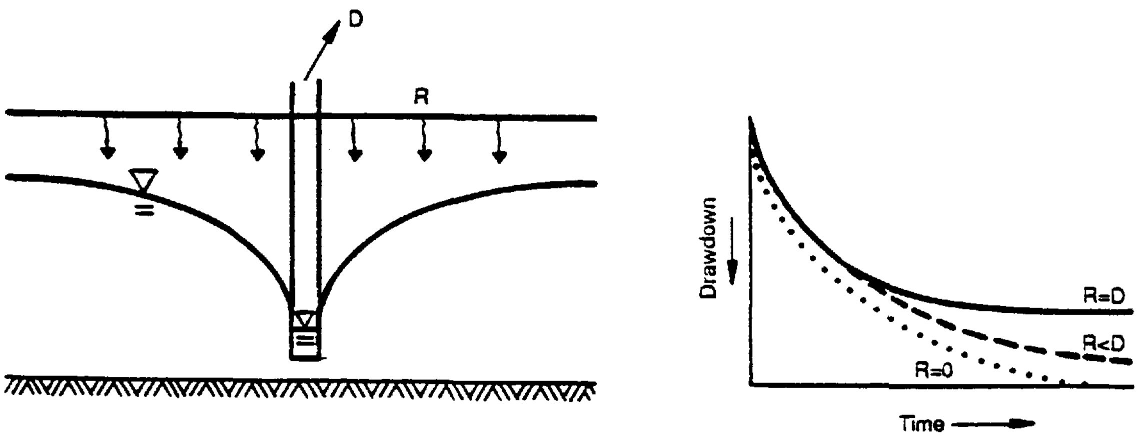
The effects of capillarity in fine-grained, dry soils may induce significant lateral movement of moisture through heterogeneous vadose zone materials. Capillarity effects are also responsible for creating barriers that substantially impede downward flow and slow recharge. Such capillary barriers are formed when uniform, coarse layers underlie fine-textured layers. In this circumstance, water in the fine-textured material cannot displace air in the pore spaces of the underlying coarse layer until the fluid pressure above the fine/coarse interface exceeds the water-entry value of the coarse layer.

Flow through the vadose zone during an infiltration event may occur under both saturated and unsaturated conditions. Saturated flow is more likely in humid climates, with moderate- to low-permeability soils and shallow water table conditions, following a series of storms of high intensity or prolonged duration. Unsaturated flow is expected to occur in drier climates, where soils are highly permeable, where storms are infrequent, and where the water table is deep. Rapid flow through the vadose zone may occur where water is ponded above fractures or macropores that create continuous high-permeability pathways for very localized saturated flow. Where this occurs, water and potentially contaminants would drain rapidly from the surface in preferential flow pathways that may lead directly to the water table. To be realistic, conceptual and quantitative models of groundwater recharge should incorporate soil heterogeneities and locations of preferential pathways that could affect the quantity and distribution of groundwater recharge.

The process of recharge is quite complex in that the infiltrated precipitation that falls on the earth's surface and is potentially available for groundwater recharge is subject to the effects of climate, vegetation, topography, soils, and variations in vadose zone materials. When infiltration is continuous, then recharge occurs in a process wherein the vadose zone becomes progressively wetter with depth and eventually the wetted soil reaches nearly a constant water content throughout. In most events, however, infiltration is sufficient to wet only a portion of the vadose zone. Following such events, during the redistribution stage, the infiltration pulse continues to move vertically downward in response to gravitational and capillary forces. At later times and at increasing depths, the water content bulge may be dampened such that an interval of high or increased water content is no longer discernible. The bulge of the water content profile during redistribution also becomes less distinct as the frequency of infiltration events increases. Where the water content is constant with depth, the rate of flow through this part of the vadose zone is constant. Although no transient, propagating water pulses are distinguishable, downward flow can still be significant and can still result in a large downward soil-water flux.

Recharge occurs when water exits the vadose zone and crosses the water table. Upon reaching the water table, the recharge may cause the water table to rise. The amount and rate of water table rise depends primarily upon the specific yield of the aquifer materials and the quantity and duration of the recharge. However, if there are sources of groundwater discharge such as pumping or evapotranspiration by phreatophytes, discharge may exceed recharge and the net result will be a water table decline (Figure 2-7). If recharge and discharge are equal during a time period, then there will be no change in the water table elevation. In the absence of other recharge sources, one must consider that the steady position of the water table is sustained by a constant source of diffuse recharge through the vadose zone. Where there is a slope on a water table that maintains a steady position over time, some recharge must be occurring in order to sustain groundwater outflow.

This report focuses primarily on methods to determine the volumetric flow rate or specific discharge through the vadose zone that becomes recharge. Specific discharge, also called the Darcy velocity, quantifies the magnitude and direction of groundwater flow through a unit cross-sectional area of aquifer or soil material. It



Source: Stephens, 1995 (with permission)

Figure 2-7. Water level decline in a well pumping at a rate *D* from an aquifer which receives within the cone of depression recharge, *R*.

is important to note that the recharge rates from the deep vadose zone will almost always be less than water infiltration rates across the soil surface, in part due to losses by evapotranspiration. It is also important to recognize that the diffuse natural recharge rate is not the same as the mean velocity of pore-water migration, as illustrated in Equation 2-2.

For additional information on the physics of flow through the vadose zone, refer to the texts by Hillel (1980a, 1980b), and for further broad discussions about recharge processes from a hydrogeological perspective, refer to the text by Stephens (1995).

Section 3

Methods to Quantify Diffuse Natural Recharge

Section 3
METHODS TO QUANTIFY DIFFUSE NATURAL RECHARGE

The various physical and chemical techniques for quantifying diffuse natural recharge require data collected from the vadose zone, the aquifer, or from gaining streams. Most of these methods are considered to be indirect methods, in that recharge is determined by making measurements of, or calculations derived from, physical or chemical information that is a manifestation of the recharge process. Such techniques include physically based methods applied to the vadose zone such as the soil-water balance, the plane of zero flux, and other methods which reflect soil-water content including soil temperature and electrical conductance. The groundwater basin-outflow method and the use of streamflow data are also physically based methods to calculate recharge from groundwater flux rates in aquifers and from stream-aquifer interactions. The only direct physical method for measuring the recharge flux is by lysimetry, or soil pore-water sampling, which directly captures percolating water at depth and measures the volume collected over time. Chemical methods also provide indirect means of calculating recharge by tracking water movement through both the vadose zone and groundwater aquifers. The most promising chemical methods involve the use of natural environmental tracers, solute balances, stable isotopes, and applied tracers. Numerical models simulating saturated and unsaturated flow can also be used to indirectly calculate the recharge component of the model's water balance.

The following sections describe each physical, chemical, and numerical method currently used to estimate or quantify diffuse recharge. The discussions address the physical basis for the required measurements or calculations, the interpretation and application of the data, and the method advantages, disadvantages and uncertainties.

PHYSICAL METHODS FOR DETERMINING DIFFUSE RECHARGE

The following paragraphs describe physical methods to quantify diffuse recharge and provide discussion of the conditions under which the method is best applied, advantages, disadvantages, and uncertainties. The methods discussed include soil-water balance, lysimeter measurements, the Darcy flux and plane of zero flux methods, soil temperature methods, electromagnetic methods, the basin outflow method, and methods incorporating water-level fluctuations and stream gauging.

Soil-Water Balance

A water balance is an equation of water mass conservation for a particular volume or region. In hydrology, one can derive a water balance for a surface water body, a watershed, an aquifer system, or a portion of the vadose zone. These regions are clearly linked together, as the output from one becomes the input to another.

The general equation for the soil-water balance is derived by considering the mechanisms by which water can enter, exit, or be stored in a defined volume of the vadose zone. For most problems, the inflow across the upper boundary of the vadose zone is infiltration, while outflow from the upper boundary is evaporation and transpiration, and outflow from the lower boundary is groundwater recharge. Net inflow (inflow minus outflow) must equal the change in soil-water stored in the vadose zone. In the soil-water balance equation, for a discrete time interval, we add flows that contribute water to the vadose zone, subtract discharges and water losses, and equate this value to changes in the amount of water stored in the soil volume. The soil-water balance equation can be presented in terms of the recharge component as:

$$R = I - E - T - \Delta S \qquad \text{(Equation 3-1)}$$

where R = deep percolation or recharge (L^3T^{-1})

I = infiltration (L^3T^{-1})

E = evaporation (L^3T^{-1})

T = transpiration (L^3T^{-1})

ΔS = the net change in soil-water storage (L^3T^{-1})

Usually, evaporation and transpiration are considered together as evapotranspiration or ET. The maximum ET, called the potential evapotranspiration (PET), occurs when the surface is fully vegetated and there is no limit to the water available for the vegetation.

Recharge in Equation 3-1 is typically computed for a horizontal area across which downward flow occurs. This horizontal area is problem-dependent, but is usually a unit cross-sectional area in plan view. The standard approach is to divide both sides of the equation by the area, and express the recharge rate and other water balance components as fluxes or specific discharges having units of LT^{-1}.

The analysis of recharge using the water balance method applied to the upper vadose zone is usually based on the monthly variation in precipitation and the monthly calculated evapotranspiration. By the conventional analysis, recharge is then predicted to occur when mean monthly precipitation exceeds actual evapotranspiration and there is no net decrease in monthly soil-water storage. However, Rushton and Ward (1979) concluded for a cool, humid climate that daily periods of accounting are required; otherwise recharge would be underestimated. From field observations at sites in England, they noted that for observed recharge patterns to be represented adequately, they needed to allow recharge during times when the calculated monthly water balance indicated there was a soil-water deficit.

In dry climates, where potential evapotranspiration is calculated on a monthly or *annual basis*, the potential evapotranspiration nearly always exceeds the mean annual precipitation, yet it has been well documented that recharge occurs under these conditions, especially in sandy and poorly vegetated soils (Stephens, 1994). The poor predictions are often attributable to the period of water balance accounting.

The importance of the period of water balance accounting is critical in dry climates where significant recharge occurs very infrequently, sometimes only annually or every several years. By averaging precipitation and evapotranspiration measured at discrete times (e.g., daily or monthly average) and specific locations and integrating averages over a basin or watershed, the water balance method applied in an area of low precipitation will erroneously predict that the long-term recharge is negligible. Actually, the recharge process in dry climates is episodic, localized and tends to occur during wet, cool seasons when precipitation actually exceeds the evaporative demand for at least short periods of time. Because these recharge periods may be shorter than the period used in the water balance method, the water balance method may underestimate recharge.

Additionally, the reliability of recharge estimates calculated as the residual of a soil-water balance obviously depends on the accuracy with which each of the other water balance components can be measured. Although a soil-water balance can be a fairly accurate and practical method when applied in humid or temperate climates (e.g., Hansen, 1991; Lyford and Cohen, 1988; Terry *et al.*, 1979), it is unlikely to be successful in arid and semiarid settings where measured precipitation is nearly always equal to or less than potential evapotranspiration, and the uncertainties in long-term calculations of both parameters are high. For example, when precipitation and evapotranspiration values are nearly equal, Gee and Hillel (1988) estimate that uncertainties in recharge of 200 percent or more can result from a mere 5 percent error in precipitation combined with a 10 percent error in evapotranspiration.

The following subsections briefly describe each of the soil-water balance components and present some of the methods to obtain them. A brief summary of evaluations of selected field methods to measure ponded infiltration rate, hydraulic conductivity, pressure head and water content are summarized in Table 3-1. For more detailed discussions, the soil physics texts by Hillel (1980a, 1980b) and Marshall and Holmes (1992) are excellent.

Infiltration. Infiltration is usually quantified in one of three ways: by the residual from a surface-water balance analysis, by field measurements, or by calculation based on soil hydraulic properties, such as unsaturated hydraulic conductivity and hydraulic head gradient. In a surface-water balance, infiltration may be computed by adding together measurements of precipitation, applied irrigation, and surface run-on, and subtracting from this sum the surface runoff, interception of water on plant canopies, direct evaporation, and increases in surface-water storage. Methods to quantify each of these components of a surface-water balance can be found in standard engineering hydrology texts (e.g., Linsley *et al.*, 1992) and reference books (e.g., Wilson *et al.*, 1995a). The reliability of infiltration values calculated as the residual of a surface-water balance suffers the same drawbacks as those associated with the soil-water balance discussed above. It is also important to recognize that, in addition to vertical flow, lateral inflow and outflow may occur within the region of the vadose zone where the water balance is to be computed. These components are usually so small that they are neglected, as was done in Equation 3-1. However, in some cases lateral water movement is significant, especially in heterogeneous or anisotropic soils and in areas of topographic variability.

Table 3-1. Summary of Selected Vadose Zone Field Test Methods
Page 1 of 2

Characteristic	Single Ring Infiltrometer	Air-Entry Permeameter	Borehole Permeameter	Tension Infiltrometer/ Disc Permeameter	Instantaneous Profile (IP) Test
Parameters Measured	Ponded infiltration rate	Vertical K_{fs}	Multidimensional K_{fs}	Multidimensional K_{fs} and $K(\psi)$, sorptivity, pore geometry	Vertical K_{fs}, $K(\psi)$, $K(\theta)$
Range of Use	$K_{fs} = 10^{-2}$ to 10^{-6} cm/s	$K_{fs} = 10^{-1}$ to 10^{-9} cm/s	$K_{fs} = 10^{-1}$ to 10^{-8} cm/s	$K_{fs} = 10^{-2}$ to 10^{-6} cm/s; tensions up to 15 cm of water	$K_{fs} = 10^{-2}$ to 10^{-6} cm/s
Time to Complete	<4 h at $K_{fs} = 10^{-5}$ cm/s	<4 h at $K_{fs} = 10^{-5}$ cm/s	<4 h at $K_{fs} = 10^{-5}$ cm/s	20 minutes at 10^{-2} cm/s 8 hours at 10^{-6} cm/s	3 days for 10^{-2} cm/s 3 weeks for 10^{-6} cm/s
Depth of Test	Surface/shallow	Surface to ~0.5 m	Any	Surface/shallow	Surface to 5 m
Suitable Soils	Coarse sands and finer	Unconsolidated coarse sands and finer	Any	Any, including fractured and macro-porous soils. Disc permeameter not suitable for wet soils.	Unconsolidated sand and silt.
Relative Accuracy	Low	High	Variable depending on solution	High	High
Relative Cost	Low	Low to moderate	Low-moderate	Low	Moderate to high
Disadvantages	Lateral flow affects accuracy; surface crust may reduce infiltration	Difficult to drive ring in stony soil; difficult to identify wetting front in wet soil	Analytical solution affects accuracy of results; borehole smearing can create hydraulic barrier	Small-scale test. Analytical solution for disc permeameter limits its use to relatively dry soils.	Long test times. Requires considerable installation and monitoring effort. Not suitable for highly layered soils. Soil profile must be freely draining.
Advantages	Simple, rapid; estimates K_{fs} from infiltration; can increase size to reduce lateral flow effects. Provides large-scale test results.	Portable and easy to operate; accurate results.	Simple solutions good approximations for sands; other solutions account for capillarity	Portable and easy to operate. Provides accurate results for a variety of soils, including rock.	Provides accurate, large-scale results.

Table 3-1. Summary of Selected Vadose Zone Field Test Methods
Page 2 of 2

Characteristic	Tensiometer	Psychrometer	Neutron Moisture Logging	Time Domain Reflectometry (TDR)
Parameters Measured	Soil matric potential	Water vapor activity (relative humidity)	Soil moisture content.	Soil moisture content.
Range of Use	Potentials = 0 to 0.8 bars	Potentials = 0.9 to 70 bars	0 to 100% saturation.	0 to 100% saturation.
Time to Complete	Equilibrates within 1 hour	Typically equilibrates within 1 hour	Typically less than 2 minutes per reading.	Less than 1 minute per reading.
Depth of Test	Typically <6 m	Any	Any	Any
Suitable Soils	Any moist to wet soils.	Any damp to dry soils.	Any	Any
Relative Accuracy	High	High	Moderate; typically ±1% by volume.	Moderate; typically ±2% by volume.
Relative Cost	Low	Low	Moderate	Moderate to high
Disadvantages	Difficult to install. Requires considerable maintenance. May not recover if too dry.	Requires readout device on data logger. Temperature sensitive. Requires careful calibration.	Contains radioactive source. Requires special handling and license. Measures moisture within a variable size "sphere."	Highly sensitive to soil salinity and water chemistry.
Advantages	Provides easy, accurate, measurements within the range of operation.	Provides easy, accurate, measurements within the range of operation.	Widely used and accepted; very portable.	Provides repeatable, measurements to a constant soil volume. Can be automated.

The most common field methods used to measure surface infiltration rates in the vadose zone include infiltrometers and rainfall simulators. Infiltrometers measure infiltration under ponded conditions (often artificially contrived), while a rainfall simulator is useful to determine surface infiltration under non-ponded conditions. A review of infiltrometer test methods is provided in the ASTM Standard Guide D 5126-90 (ASTM, 1995). Please note that infiltrometers do not explicitly measure saturated hydraulic conductivity unless the hydraulic head gradient is also determined or can be reasonably estimated to be unity (one) within the infiltrometer sample.

One type of infiltrometer, called a single-ring infiltrometer, is simply a metal cylinder pushed 5 to 10 cm into the soil. Alternatively, infiltrometers can be constructed as infiltration basins, wherein water is ponded within a rectangular area bounded by low berms constructed by compacting native soil or a soil-bentonite mixture. Water is ponded in the infiltrometer, and the steady volumetric rate of water added to maintain a constant head is measured.

Although infiltrometers can be made in many sizes, most researchers recommend using a cylinder large enough to minimize or avoid the effects of laterally diverging flow at the base of the cylinder. Bouwer (1995) suggests using a fairly large cylinder of 2 m or more in diameter, or a bermed, rectangular, infiltration basin of at least 2x2 m for infiltration tests. Large infiltrometers yield reasonably accurate estimates for surface infiltration rates under the imposed hydraulic head.

Infiltrometers are simple and cost-effective to construct and operate, and are most suitable for materials with field saturated hydraulic conductivities within the 10^{-2} to 10^{-6} cm/s range. Tests can typically be completed within a few days to within an hour, depending on the soil.

The lower limit of the range of permeabilities suitable for infiltrometer testing is influenced by several factors. For low-permeability materials (e.g., silt- and clay-rich soils) where the flow rate of water into the soil is slow and the test duration long, evaporation of water from the infiltrometer can exceed the infiltration rate. In this case, the open top of the infiltrometer can be covered or otherwise sealed to minimize evaporation. This is the concept behind the ASTM sealed double ring infiltrometer (ASTM D 3385-88). In clay-rich soils subject to swelling, the volume of water taken up by swelling clays can be a significant portion of the total volume of water in the infiltrometer. This can lead to conservatively high estimates for infiltration rate. Some investigators (Chen and Yamamoto, 1986) accounted for the effects of swelling by measuring the increase in thickness of the clay and applying a correction factor during data reduction.

Conversely, where dispersed clay particles are disturbed and during filling of the infiltrometer, the dislodged particles can collect within the pore spaces in the upper surface of the soil and create a low-permeability crust. This effect can be minimized by installing splash guards within the infiltrometer or by covering the soil surface with coarse sand prior to infiltration. Furthermore, since the swelling and dispersion of clays is very sensitive to the chemistry of the infiltrating water, it is important to determine the electrolyte concentration and the sodium adsorption ratio of the infiltrating water (see Dane and Klute, 1977, for further discussion).

A variation of a large-scale infiltrometer test can be achieved by using a rainfall simulator, which emulates precipitation as rainfall. Rainfall simulators can quantify infiltration over areas of from 1 to 10 m^2 and can provide highly accurate results under a wide range of typical field conditions. Their primary advantage is that they can be used to simulate actual precipitation rates and duration. Many workers have used this approach to simulate controlled duration and intensity storms (e.g., Zegelin and White, 1982).

Infiltration can also be calculated from hydraulic properties of the soil using *mathematical expressions* that describe the infiltration process. The type of equation chosen depends upon the nature of the process, such as transient or steady infiltration and ponded or non-ponded infiltration. Perhaps the most well known mathematical model for infiltration was developed by Philip (1957). Philip's equation predicts infiltration from unsaturated soil properties and the field water content for the case in which a constant water content is maintained at the soil surface. A simpler expression to calculate the infiltration rate is Darcy's equation:

$$q = -K(\theta)\nabla h \qquad \text{(Equation 3-2)}$$

Here, the infiltration rate, designated as q (LT^{-1}), is calculated from field measurements of the hydraulic gradient, ∇h (LL^{-1}), near the soil surface and the hydraulic conductivity, K (LT^{-1}), at the field water content, θ, behind the wetting front. The following discussion summarizes some of the methods for determining the parameters in Equation 3-2.

The saturated and unsaturated hydraulic conductivity of vadose zone materials can be measured either by field methods or by laboratory methods applied to soil core samples:

- *Field methods:* Measurements of saturated hydraulic conductivity can be made using air-entry permeameters (AEPs) and borehole permeameters (BHPs). Unsaturated hydraulic conductivity can be determined by the instantaneous profile or internal drainage method and tension infiltrometers and disc permeameters. As described in a later section, the hydraulic gradient is computed from *in situ* measurements of pressure head using tensiometers (soil-water capillary tension) and psychrometers (soil relative humidity).

- *Laboratory methods:* Saturated hydraulic conductivity can be determined using constant-head (ASTM D 2434-68) and falling-head permeameters. Unsaturated hydraulic conductivity relationships can be determined directly using several laboratory methods, as summarized by Stephens (1993). Hydraulic conductivity can also be calculated from laboratory measurements of the moisture retention characteristic curves.

The field methods for determining hydraulic conductivity and the hydraulic gradient are discussed in the following paragraphs; further discussion regarding laboratory methods is included in a later segment of this section which describes recharge by the Darcy flux method.

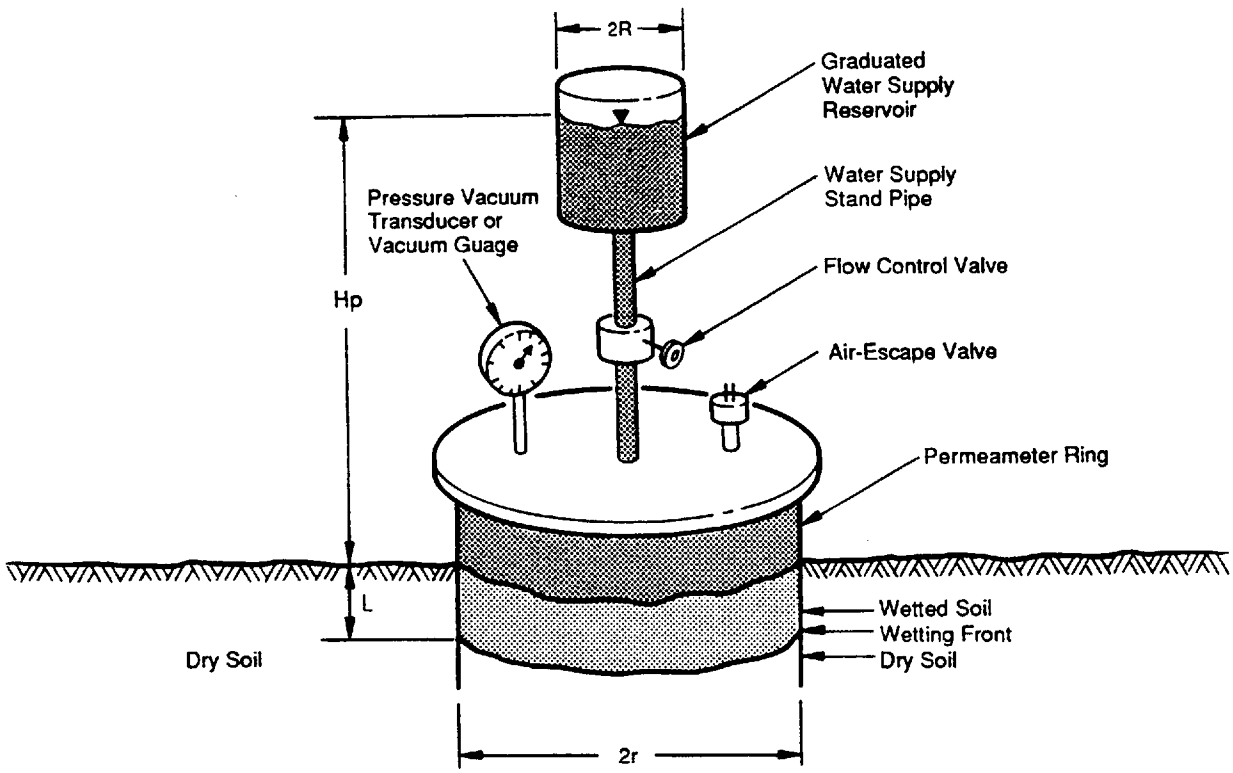
Field Methods for Measuring Saturated Hydraulic Conductivity. The test methods summarized herein determine field-saturated, or satiated, hydraulic conductivity (K_{fs}). This value is less than saturated hydraulic conductivity (K_s), owing to entrapped air that commonly occurs during ponded infiltration. Entrapped air that develops in a field permeameter reduces the area through which water flow occurs and accordingly reduces the hydraulic conductivity measured in the field by as much as a factor of two or more. The field values may be adjusted for entrapped air by empirical methods or the soil tested can be flooded with carbon dioxide gas to minimize entrapped air. In many cases, however, the field-saturated hydraulic conductivity is preferred in recharge calculations.

Air-entry permeameter. As indicated by Bouwer (1978), the AEP is the simplest and quickest technique for measuring saturated hydraulic conductivity in the vadose zone. This method uses measured values for infiltration rate under an imposed hydraulic gradient and air-entry pressure to determine saturated hydraulic conductivity.

The device consists of a metal cylinder with a sealed top about 30 cm in diameter that is connected to a graduated, overhead water supply reservoir and a vacuum gauge (Figure 3-1). The cylinder is driven 15 to 25 cm into the soil and both cylinder and reservoir are filled with water. The testing is conducted in two stages: infiltration, and drainage. During the first stage of testing, water is allowed to infiltrate into the surface materials under positive head, and the rate of decline in the reservoir water level is used to measure infiltration rate. At the start of the second stage, the water supply reservoir valve is closed and the infiltrated water is allowed to drain under tension. The tension within the permeameter above the soil surface increases as drainage occurs. Tension increases until the air-entry pressure of the soil is reached.

The solution equation (Figure 3-1) is based on a formulation of the Darcy equation, wherein half the air-entry pressure head is assumed to be equivalent to the pressure head along the wetting front, Ψ_f. The pressure head along the wetting front, the depth of the wetting front, and the height of the water level in the reservoir provide estimates for the hydraulic gradient, and the rate of fall of water in the reservoir during the first stage are used to determine field-saturated hydraulic conductivity.

The AEP method provides a good measure of vertical field-saturated hydraulic conductivity typically in less than 4 hours with moderate costs. When applied at the surface, the AEP provides results for the upper 2 feet or so of vadose zone material. However, the method can also measure the hydraulic conductivity of deeper layers by placing the device in pits or trenches excavated to depths of up to about 8 to 10 feet. The method can be applied to a wide range of soil types ($10^{-9} < K < 10^{-1}$ cm/s) including sand, silt and clay. Although AEP tests have been successfully completed in lithified rock, they are most commonly used for testing unconsolidated materials. AEP tests are generally not suitable for soils with roots, worm burrows, or macropores, nor are they appropriate for gravelly soils where emplacement of the ring is difficult.



$$K_{FS} = \frac{L\left(\frac{dH}{dt}\right)\left(\frac{R}{r}\right)^2}{H + L - 0.5\,P_a} = \text{Vertical, Field Saturated Hydraulic Conductivity } \left[\frac{L}{T}\right]$$

L	=	*Depth to wetting front [L]*
H	=	*Height of water in reservoir at end of infiltration stage [L]*
$\frac{dH}{dt}$	=	*Rate of fall of water in reservoir just before end of infiltration* $\left[\frac{L}{T}\right]$
R	=	*Radius of reservoir [L]*
r	=	*Radius of permeameter ring [L]*
P_a	=	*Air-entry head pressure [L]*

Source: Modified from Havlena and Stephens, 1991 (with permission)

Figure 3-1. Air-Entry Permeameter

AEPs provide an expedient means of accurately measuring the vertical, field-saturated hydraulic conductivity of near-surface materials. The AEP method is the only commonly used field method that measures the hydraulic gradient, and hence, the actual saturated hydraulic conductivity in a vertical orientation. Several assumptions are included within the use of AEPs (ASTM D5126-90): (1) the movement of infiltrating water is essentially one-dimensional downward, constrained by the permeameter ring, (2) soil gas within the pore spaces does not offer any impedance to the downward movement of water, (3) the wetting front is distinct and easily identified, (4) dispersion of clays and subsequent clogging of the surface layer of finer soils is insignificant, and (5) the soil is non-swelling, or the combined effects of swelling and dispersion can be minimized (e.g., by tightly packing coarse sand within the permeameter ring above the soil surface). Thorough reviews of the application of an AEP can be found in Bouwer (1995) and in the ASTM Standard Guide D 5126-90.

Borehole permeameter. A second method of measuring field-saturated hydraulic conductivity in the vadose zone is with a BHP. The test is conducted within the lower section of an open or screened borehole. Water is allowed to flow into the borehole, and the flow rate is regulated and monitored so that the water level in the borehole is maintained at a constant, known depth above the bottom of the hole. Field-saturated hydraulic conductivity is determined from the steady-state rate of inflow of water into the borehole, the borehole geometry, and the depth of ponding within the borehole.

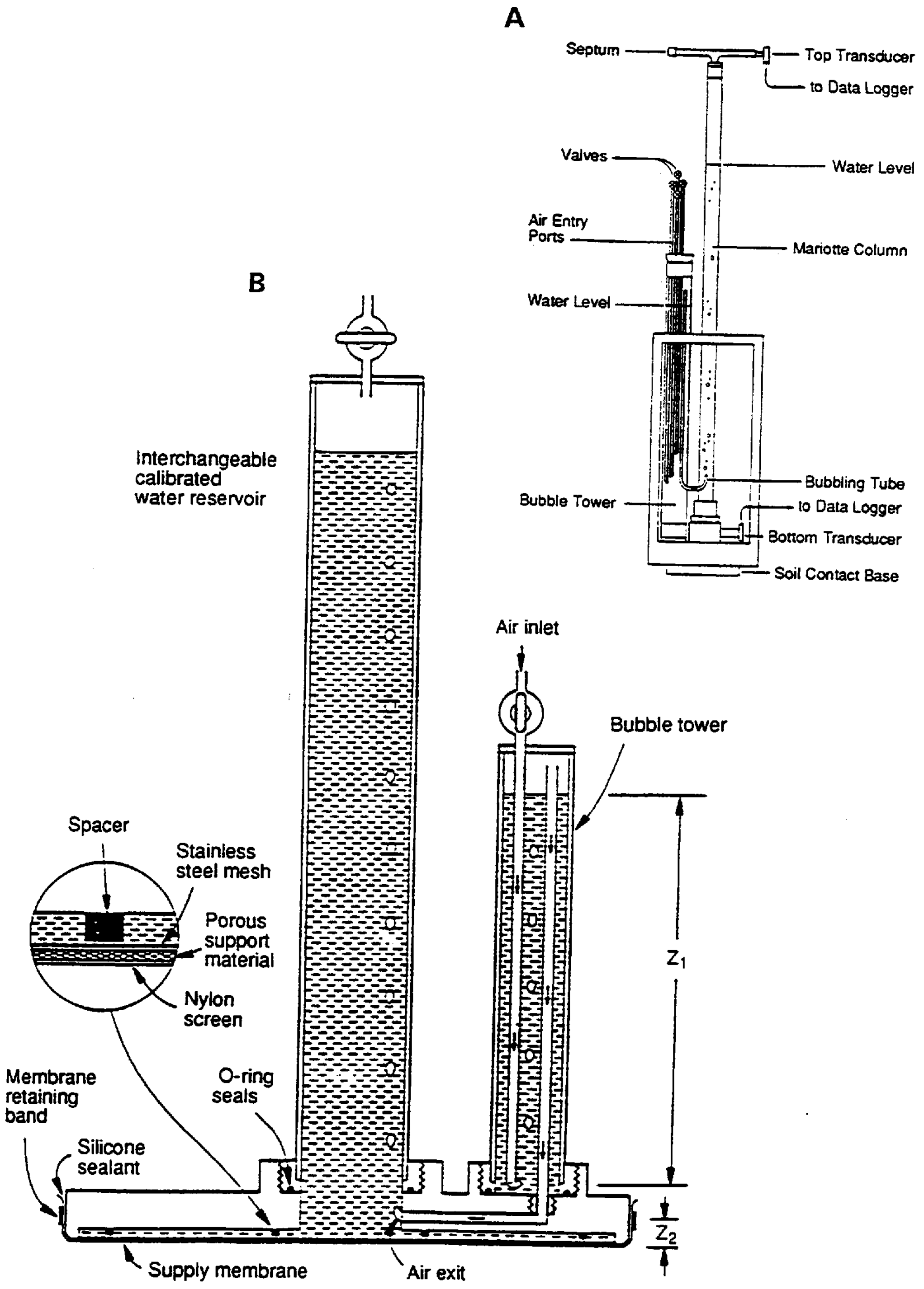
Several solutions are available for different conditions and applications. Less rigorous solutions (e.g., Glover, 1953; U.S. Bureau of Reclamation, 1978) ignore capillary effects and are best suited for moist or coarse-grained materials where capillary effects are minimal. The solutions which consider unsaturated flow away from the borehole (Stephens *et al.*, 1987; Philip, 1985) are thought to be most accurate and provide a good measure of K_{fs} in relatively dry or fine-grained materials with significant capillary effects. Other solutions for BHP tests which require more than one test within each borehole have recently been developed (Reynolds and Elrick, 1985). These tests also consider unsaturated flow away from the borehole, and can also provide estimates for matric flux potential and sorptivity. For each BHP test, a constant infiltration rate can be reached in an hour or less for sandy materials and in several hours (up to 36 to 40 hours) in finer-textured materials for a low to moderate cost.

The method is suitable for a wide range of fine to coarse materials within the $10^{-8} < K < 10^{-1}$ cm/s range. BHP tests are the only currently available tests which can measure K_{fs} at depth in the vadose zone and can be performed at any depth or at multiple depths as the borehole is drilled to obtain a profile of hydraulic conductivity.

Borehole permeameters provide an accurate means to measure the saturated hydraulic conductivity of materials at virtually any depth within the vadose zone. The more rigorous BHP tests which incorporate capillarity provide the most accurate results. Unlike the AEP, because the flow of water from the permeameter is not constrained, BHP tests measure an effective hydraulic conductivity which includes horizontal as well as vertical components. A good review of the BHP method is provided in ASTM Standard Guide D 5126-90.

Field Methods for Measuring Unsaturated Hydraulic Conductivity. The two most frequently used test methods for measuring unsaturated hydraulic conductivity in the field are the tension infiltrometer and the instantaneous profile test.

Tension infiltrometer/disc permeameter. The tension infiltrometer, and the very similar disc permeameter (Figure 3-2), have recently been developed by Perroux



Source: *Ankeny et al.*, 1988 (with permission)

Figure 3-2. A. Tension Infiltrometer
B. Disc Permeameter

and White (1988) and Ankeny *et al.* (1988; 1989) as methods to rapidly determine saturated and unsaturated hydraulic conductivity and various unsaturated hydraulic parameters. Although disc permeameters and tension infiltrometers differ somewhat in the approach to compute conductivity, both infiltrometers are similar in design and operation. Both incorporate a Mariotte siphon head control device and a disc-shaped membrane through which water is allowed to infiltrate under tension into the underlying soil. Although they both may be used to measure saturated hydraulic conductivity, their primary application is to determine unsaturated hydraulic conductivity at low water tensions. However, their ability to operate under both positive and negative heads allows tension infiltrometers and disc permeameters to be used for determination of the relative contribution of macropores to the overall saturated hydraulic conductivity. This is an important feature of the tension infiltrometer and disc permeameter which greatly extends their application to include soils with macropores.

The standard solution for unsaturated hydraulic conductivity by the disc permeameter method is based on the Wooding (1968) solution for three-dimensional flow from a shallow circular pond or disc. This solution requires measuring transient infiltration rate. In the analysis, another unsaturated soil property, the sorptivity, is also obtained. This solution requires that the soils are initially dry; consequently, the method is not applicable for soils that are initially at or near saturation.

The solution for tension infiltrometers (Ankeny, *et al.*, 1988) is also based on the Wooding equation, but uses the simultaneous solution of four equations for four unknowns in order to determine the hydraulic conductivity. Two of the four unknowns are the hydraulic conductivities at two different tensions (the desired values). The other two unknowns are the values for another unsaturated soil property, matric flux potential, at the two tensions. This solution, therefore, requires that two tests be conducted at the same location, each at a different tension.

Tension infiltrometers and disc permeameters provide a very rapid, reasonably accurate way to measure saturated hydraulic conductivity and several unsaturated parameters under many common field conditions. They are most effective for measuring materials having saturated hydraulic conductivities in the 10^{-6} to 10^{-2} cm/s range. However, tension infiltrometers usually are not practical for measuring unsaturated conductivities that are much less than one or two orders of magnitude less than the saturated hydraulic conductivity. The equipment is extremely portable, and the testing procedures are straightforward. Tension infiltrometer and disc permeameter tests measure an effective hydraulic conductivity which includes horizontal as well as vertical components. In addition to saturated and unsaturated hydraulic conductivity, tension infiltrometers and disc permeameters also measure unsaturated flow parameters, including sorptivity, matric flux potential, and pore geometric parameters. White *et al.* (1992) provide an overview of the disc permeameter and tension infiltrometer methods.

Instantaneous Profile Test. The instantaneous profile (IP) test was first proposed by Watson (1966) as a field method for determining the unsaturated hydraulic conductivity. The method is based on determining the rate of drainage and hydraulic gradient in the soil profile following a thorough wetting under ponded conditions. The wetting is usually accomplished by berming an area several meters in diameter. After infiltration, the surface is covered with impermeable material to prevent evaporation. Hydraulic conductivity is calculated from:

$$K(\theta) = \frac{\int_o^D (\partial\theta/\partial t)\, dz}{\frac{dh}{dz}} \qquad \text{(Equation 3-3)}$$

where h is hydraulic head (L), z is vertical space coordinate, D is depth, θ is water content, and t is time. The integral represents the rate of internal drainage and has units of LT^{-1}.

The drainage rate is calculated from *in situ* measurements over time of the water content profile. Water content is readily measured *in situ* by the neutron moderation method. One or more neutron probe access tubes are installed within the bermed area to a depth of a few meters or so. Two-inch-diameter aluminum is the preferred material for neutron probe access tubes, but plastic and steel casing are sometimes used. Neutron logging, a widely accepted method for monitoring soil moisture content *in situ*, employs a 10- to 100-millicurie americium-beryllium neutron source and a detector within a cylindrical probe that is moved within the access tube. Hydrogen atoms within the water molecules in the soil water slow, or "thermalize," the neutrons. The number of thermalized neutrons are then detected by the probe. The number of thermalized neutrons is compared with a standard count and the ratio is used to determine *in situ* moisture content.

To calculate hydraulic conductivity at field water content, $K(\theta)$, the hydraulic gradient, dh/dz, must also be quantified by *in situ* measurement or by estimation. Measurements of the hydraulic gradient, using a nest of tensiometers for example, are described separately in a subsequent section. In lieu of measurements, however, it is sometimes reasonable to assume that the hydraulic gradient is simply unity, inasmuch as gravity is the dominant contributor to the gradient in this case.

IP tests produce accurate results for vertically oriented unsaturated hydraulic conductivity of near-surface soils. Tests can be conducted in most materials, including sand, silt, and clays. However, the time required for low-conductivity materials can often exceed several months. Because of the relatively large size of the test plots, the results of IP tests represent a scale much larger than any of the other test methods. The larger scale is important for most investigations where large-scale unsaturated hydraulic conductivity data are required. Inasmuch as the draining soil becomes more unsaturated with time, the minimum unsaturated hydraulic conductivity quantified will depend on the length of the test. Typically, tests are terminated before achieving the conductivity at the *in situ* or field water content, because of the long time for complete drainage.

<u>Calculating Unsaturated Hydraulic Conductivity</u>. Unsaturated hydraulic conductivity may also be calculated from the moisture retention curves. One method, developed by Brooks and Corey (1964), is a graphical procedure that operates on moisture content-pressure head, θ-Ψ, data to obtain two parameters: the pore-size distribution index, λ, and the critical pressure or bubbling pressure head, Ψ_b. The Brooks-Corey (1964) soil water retention curve is described by:

$$\theta = (n - \theta_r)\left(\frac{\Psi_b}{\Psi}\right)^{\lambda} + \theta_r \qquad \text{(Equation 3-4)}$$

Once these two parameters, λ and Ψ_b, and saturated hydraulic conductivity are known, the unsaturated hydraulic conductivity is calculated from:

$$K(\psi) = K_s \left(\frac{\psi_b}{\psi}\right)^{2 + 3\lambda} \quad \text{(Equation 3-5)}$$

A more popular method to calculate unsaturated hydraulic conductivity is based on a statistical model of the porous medium which is characterized by the soil-water retention curve. This is a computer-based technique which fits a two- or three-parameter equation to measured water retention data (van Genuchten *et al.*, 1991). These fitting parameters (α,N,m) and measured saturated hydraulic conductivity are applied in the following equation to compute unsaturated hydraulic conductivity.

$$K(\psi) = K_s \frac{(m[1 + (\alpha\psi)^N]^{-1})^2}{[1 + (\alpha\psi)^N]^{m/2}} \quad \text{(Equation 3-6)}$$

where $m = 1 - \frac{1}{n}$

The calculated unsaturated hydraulic conductivity methods are regarded as good alternatives to the more tedious field or laboratory measurement techniques. In some cases, there is considerable uncertainty in the predictions, especially in the dry range. Improved accuracy is obtained if one or more measurements of unsaturated hydraulic conductivity are used as matching factors. Stephens and Rehfeldt (1985) found that the calculation technique by van Genuchten compared very favorably with other field and laboratory methods. The computer program SOIL (EL-Kadi, 1985) incorporates the van Genuchten and Brooks-Corey methods for estimating unsaturated hydraulic conductivity from moisture retention data.

In the laboratory, the soil-water retention curve is determined on core samples using a hanging water column apparatus and pressure plate apparati (see, e.g., Klute, 1986). In the hanging column, the core sample is placed on the porous ceramic plate of the Büchner funnel and saturated. The Büchner funnel is connected to a flexible water-filled tube and burette with a stopcock. With the stopcock closed, the burette is lowered, so that the core is under a pressure head equal to the distance between the center of the core and the water level in the burette. The stopcock is opened, and water flows into the burette until equilibrium is reached. The amount of flow is measured, and the distance between the center of the core and new water level in the burette is recorded. The water content of the core at the applied pressure head represents one point on the water retention curve. The process is repeated stepwise to obtain the full curve.

Pressure plate apparati consist of a rigid vessel fitted with a porous plate on the base. Soil core samples are placed on the plate, and with the vessel lid closed, pressure is applied. Water flows out of the sample until the pressure head of the soil is in equilibrium with the applied pressure. The water content of the core at the applied pressure head becomes one point on the soil-water retention curve. The process is repeated stepwise over increasing pressures to characterize the entire curve.

Several empirical techniques are available for estimating the water retention characteristic curves based on particle-size distribution and other easily obtained data such as bulk density and percentage of organic matter present. These techniques include relationships developed by Rawls and Brakensiek (1985), Haverkamp and Parlange (1986), and others compiled by van Genuchten *et al.* (1992). Each of these methods is typically suitable for a different range of soil types. However, the method of Rawls and Brakensiek (1985) is useful for a wide range of soils, where sand fractions range from 5% to 70% and clay fractions range from 5% to 60%. The Rawls and Brakensiek method uses regression equations to solve for the Brooks-Corey (1964), soil-water retention parameters (ψ_b, λ, θ_r) as a function of percent sand, percent clay, and soil porosity:

$$\psi_b = exp[5.340 + 0.185C - 2.484n - 0.002C^2 - 0.044Sn - 0.617Cn + 0.001S^2n^2 - 0.009C^2n^2 - 0.00001S^2C + 0.009C^2S - 0.0007S^2n + 0.000005C^2S - 0.500n^2C] \quad \text{(Equation 3-7)}$$

$$\lambda = exp[-0.784 + 0.018S - 1.062n - 0.00005S^2 - 0.003C^2 + 1.111n^2 - 0.031Sn + 0.0003S^2n^2 - 0.006C^2n^2 - 0.000002S^2C + 0.008C^2n - 0.007n^2C] \quad \text{(Equation 3-8)}$$

$$\theta_r = -0.018 + 0.0009S + 0.005C + 0.029n - 0.0002C^2 - 0.001Sn - 0.0002C^2n^2 + 0.0003C^2n - 0.002n^2C \quad \text{(Equation 3-9)}$$

where C = Percent clay (5<C<60)

S = Percent sand (5<S<70)

n = Porosity (volume fraction)

λ = Pore-size index

ψ_b = Bubbling pressure (cm)

θ_r = Residual water content (volume fraction)

By estimating these parameters from more readily obtained particle size curves, one can determine both the soil-water retention curve and the unsaturated hydraulic conductivity.

In general, most of the estimation techniques produce reasonable, order-of-magnitude results, which are useful for many purposes. However, because of the sensitivity of Darcy-based analyses to the unsaturated flow parameters, all empirical estimation techniques should be used with caution.

Kool and Parker (1988) developed a hybrid field method to estimate the moisture retention characteristics and unsaturated hydraulic conductivity relationship. Their method uses moisture content and pressure head data obtained by neutron logging and tensiometers during a field test. The field data are analyzed using a non-linear least squares inverse algorithm which incorporates the van Genuchten model. This method appears to provide good results and extends the van Genuchten model to include hysteresis.

Determining Hydraulic Gradient. The vertical hydraulic gradient, (*dh/dz* in Equation 3-3 and ∇h in Equation 3-2) determines the vertical driving force under which water is moving, in both magnitude and direction. To determine hydraulic gradient requires measurement of the soil pressure head profile. Pressure head is measured *in situ* at different depths using tensiometers or psychrometers.

The tensiometer is the most widely used instrument for measuring pressure head. The tensiometer consists of a sealed tube filled with water, the lower end of which is attached to a porous cup. A pressure gage is attached to the upper end. The porous cup is kept saturated and must remain in hydraulic contact between soil water and water in the tensiometer. The principle is that water flows out of the porous cup (usually ceramic) in response to soil moisture conditions until the vacuum that builds up in the tensiometer equals the soil-water pressure head. Because the pressure head data are used for measuring hydraulic gradient, it is important to determine the pressure head with sufficient precision and sensitivity to allow a reasonable estimate of gradient. For increased accuracy, it is best to use a mercury manometer or a pressure transducer, as opposed to a vacuum gauge, to monitor the vacuum in the tensiometer. Tensiometers are useful to measure pressure head to as low as about –800 cm H_2O, and they also measure pressure head under fully saturated conditions. Tensiometers fail in freezing conditions, although some success may be achieved if antifreeze is used to fill the tensiometer.

Good reviews of this method are found in the ASTM Standard Guide D 3404-91, in Klute (1986), and in Yeh and Guzman (1995).

Psychrometers are most frequently used to measure pressure head in dry, thick vadose zone materials such as those found in the arid and semiarid west, since their range of operation extends from about −1000 to −70,000 cm H_2O. Psychrometers function by measuring the water vapor potential (relative humidity) in the subsurface atmosphere. Assuming the water vapor is in equilibrium with the pore liquid, the psychrometer provides a means for indirectly estimating the soil water matric potential. Because vapor pressure is a function of ambient temperature, the water potential measured is extremely sensitive to temperature and care must be taken to minimize air currents and temperature fluctuations. Good reviews of the theory and application of psychrometers are provided in Rasmussen and Rhodes (1995) and Klute (1986).

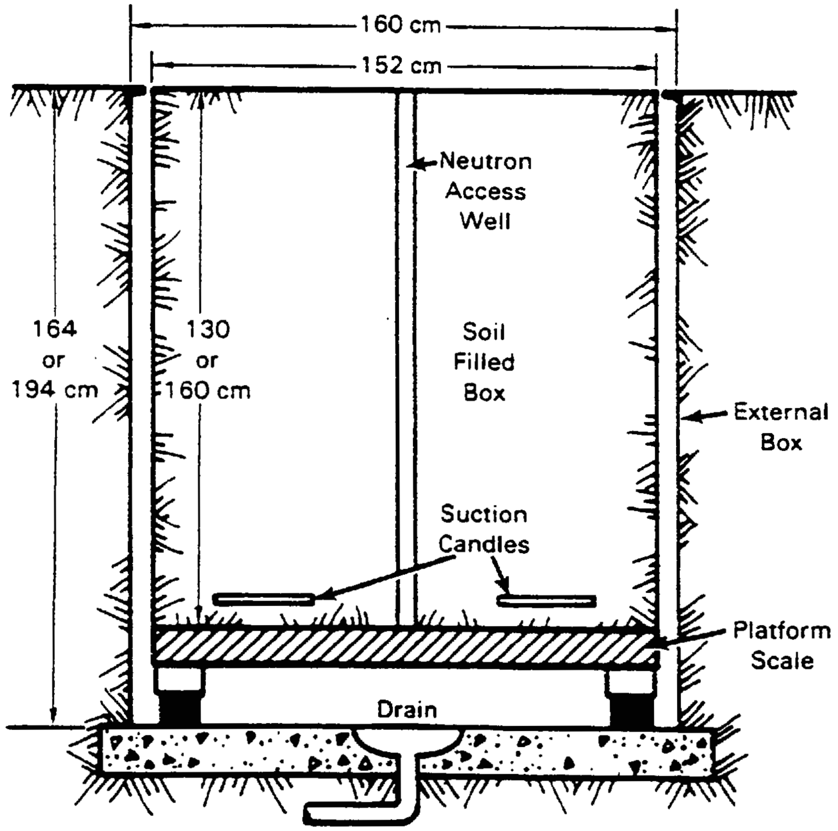
Determining Evaporation and Transpiration. Evaporation refers to the water lost from the vadose zone by vapor phase transport from the soil directly to the atmosphere. Transpiration is the water depleted from the vadose zone by plant root uptake. For most practical problems, it is both difficult and unnecessary to separate these two processes, so the two are combined and called evapotranspiration. Normally, the single largest outflow component from the vadose zone occurs through evapotranspiration. Research on evapotranspiration measurement is extensive, beginning over 400 years ago (Sosebee, 1976), and development of new quantification methods continues today. There are two different approaches to determine evapotranspiration: by measurement of changes in soil-water content and by estimation of climatic or meteorological parameters that correlate to evapotranspiration.

Lysimeters provide the most accurate method to directly measure changes in soil-water content due to water losses by evapotranspiration, and in fact, most evapotranspiration estimation methods have been verified by comparing the predictions to lysimeter data. Unfortunately, the lysimeters used for evapotranspiration studies are cumbersome, expensive to construct, and require rather long periods of data collection. Natural, non-irrigated lysimeters yield dependable values for actual evapotranspiration rates only when monitored over long periods between major rains, for example, seasonally or annually (Van Bavel, 1961). The accuracy of lysimeters is achieved only when several installation requirements are met. These are:

- The lysimeter exposure is representative of surrounding field conditions.
- The soil profile in the lysimeter has a moisture content, moisture tension, thermal conditions and root distribution representative of undisturbed conditions.
- The moisture stored in the lysimeter soil can be accurately measured.

There are in general three types of lysimeters used in evapotranspiration measurements: weighing, non-weighing and floating. All three types share the same concept. A small monolith of soil with vegetation is placed in a container and is returned to its original position in the landscape. Instrumentation is emplaced to allow measurements of precipitation, soil-water storage, and deep drainage. From these components one can compute evapotranspiration using the water balance equation (Equation 3-1).

The following procedures for the three types of lysimeters are used. In a weighing lysimeter, the lysimeter is placed on a scale (Figure 3-3) having a capacity sufficient to determine the mass of a soil monolith with a diameter of one to a several meters and a depth of about one meter. Large weighing lysimeters are expensive to build,



Source: Kirkham *et al.*, 1984 (with permission)

Figure 3-3. Cross-sectional view of weighing lysimeter.

difficult to move, and must be weighed *in situ*, but are considered the only accurate, practicable method of obtaining actual evapotranspiration rates, particularly in humid climates (Van Bavel, 1961). Precision weighing lysimeters can provide measurements of evapotranspiration that are reliable to within less than a millimeter of water. Bare soil evaporation can be determined using micro-lysimeters, a technique using short soil core samples that are easily removed from the soil for weighing over periods of 1 to 2 days (Boast and Robertson, 1982). The weighing micro-lysimeter technique is labor-intensive and time consuming, and its applicability to longer periods of time and varied soil conditions has not been demonstrated.

In a non-weighing lysimeter, evapotranspiration from the soil monolith is determined by measuring the rate of water supply to the monolith container that is necessary to maintain a constant depth to water in the base of the container. However, this procedure yields representative results only when duplicating shallow water table conditions.

In a floating lysimeter, the soil monolith is placed on a liquid-filled pillow so that water gains and losses can be obtained by measuring fluid pressure in the pillow through a manometer tube. This technique offers the same level of accuracy as the weighing lysimeter, so long as the installation requirements described in the previous paragraph are met and the sensitivity of the weighing apparatus is sufficient.

The approach of estimating evapotranspiration based on climatic measurements offers a sound alternative to lysimetric methods. Rosenberg *et al.* (1983) present an excellent detailed discussion of evapotranspiration, including micrometeorological methods to estimate this component of the soil-water balance. Micrometeorology entails measuring climatic variables at a given field location. These climatic variables include incoming or net radiation, air temperature, relative humidity, and wind speed above a bare soil surface, or above or within a crop or vegetated surface. These data are used to provide estimates of potential evapotranspiration. Among the climatological methods, some are based on air temperature (e.g., Thornthwaite, 1948; Blaney and Criddle, 1950), and others are derived from solar radiation measurements (e.g., Jensen and Haise, 1963) or incorporate both energy supply data and turbulent transfer of water vapor away from the surface (e.g., Penman, 1948). When the actual evapotranspiration rate is not limited by the amount of soil moisture present, the rate is primarily dependent upon such meteorological factors.

Micrometeorological measurements have been used to estimate actual evapotranspiration to within about 10 percent of actual values (Gee and Hillel, 1988). There are two primary advantages that micrometeorology offers over lysimetry:

- The climatic variables measured in micrometeorology apply over a wider scale than the spot sampling techniques employed with lysimeters.
- Where average climatological data are available, micrometeorological techniques can be used for prediction purposes.

Methods for taking micrometeorological measurements above a vegetated surface to determine actual evapotranspiration have been developed over the past 50 years or so. Two such methods are the Bowen ratio method (e.g., Tanner, 1960) and the eddy-correlation method (Swinbank, 1951).

The Bowen ratio method is based upon a simplified energy-budget equation (e.g., Marshall and Holmes, 1992; Hanks and Ashcroft, 1980), which accounts for energy inputs and losses at the land surface. Part of the net radiation received at the land surface is transformed into heat which warms the soil, plants, and atmosphere. Most importantly, the major portion of incident radiation is absorbed as latent heat during evaporation and transpiration. The energy-budget equation is thus represented by:

$$R_n - G - H - \lambda E = 0 \qquad \text{(Equation 3-10)}$$

where R_n = the net radiation at the surface (watts/m ^{2})

G = the rate at which heat is stored in the soil, water and vegetation (watts/m ^{2})

H = the sensible-heat flux which heats the air above the land surface (watts/m^2)

λ = the latent heat of vaporization (joules/g)

E = the evapotranspiration rate (g/s-m^2)

The energy-budget equation can be rearranged to:

$$\lambda E = (R_n - G) / (1 + \beta) \qquad \text{(Equation 3-11)}$$

where β is the Bowen ratio, $H/\lambda E$, or the ratio of sensible heat to latent heat. Further, it can be shown that the Bowen ratio may be determined by:

$$\beta = \gamma(T_1 - T_2) / (e_1 - e_2) \qquad \text{(Equation 3-12)}$$

where γ is the psychometric constant, e is the vapor pressure in air, T is air temperature, and the subscripts mean that the vapor pressure and temperature are measured at the same two elevations (Montieth and Unsworth, 1990). Therefore, by combining Equations 3-11 and 3-12, the evapotranspiration rate can be determined.

The data required include R_n, G, T and e. R_n can be conveniently measured with a net radiometer, and G can be obtained by installing a soil heat-flux plate just below the soil surface. Bowen-Ratio weather stations, which measure these parameters, are commercially available (Campbell Scientific, Logan, Utah). Measurements of temperature and vapor pressure are typically obtained at 1 m and 3 m above the surface, and the psychrometric constant, γ, which is equal to the specific heat capacity of air divided by the latent heat of vaporization, can be obtained from reference book constants. Vapor pressure and temperature measurements for calculating the Bowen ratio are usually obtained using weighted averages from periods of one-half to two hours (Montieth and Unsworth, 1990) over a 24-hour interval.

Application of the energy budget/Bowen ratio method is not successful for periods of less than 24 hours because the sensible heat flux cannot be measured with sufficient accuracy. The method is satisfactory for periods greater than or equal to 24 hours (Marshall and Holmes, 1992), and works best when applied in humid environments (Hanks and Ashcroft, 1980). Because the value for β is less reliable than values for R_n and G, some uncertainty is introduced when β is greater than 0.1, as typically occurs in arid environments.

The eddy correlation method, another micrometeorological technique, is based on the principle that water vapor flux across the land surface can be measured by correlating the vertical variations of wind speed, w, with vapor density, q_v (Tanner *et al.*, 1985):

$$\overline{\lambda E} = \lambda(\overline{w' \times q_v'}) \qquad \text{(Equation 3-13)}$$

where the overbars represent time averages and the primes represent instantaneous deviations about the time averages. The data collection requirements include an anemometer and hygrometer which are connected to a data logger.

The eddy correlation method can also provide measurements of sensible-heat flux, *H*:

$$\overline{H} = C_p \rho_a \ (\overline{w' \times T'}) \qquad \text{(Equation 3-14)}$$

where C_p is the specific heat and ρ_a is the density of air. Temperature fluctuations can be measured with a thermocouple connected to a data logger. The so-called eddy correlation-energy budget method (e.g., Czarnecki, 1990) is used to determine the actual evaporative flux when field instrumentation accounts for net radiation and heat conduction into the ground (as in the Bowen ratio method), and when sensible-heat-flux, *H*, is determined by the eddy correlation technique.

In an example of the application of the eddy correlation method, Tanner *et al.* (1985) placed the micrometeorological instruments 1.1 meters above a crop canopy to record data every 10 seconds on a data logger for computation of half-hour average evaporative and sensible-heat fluxes over a period of about 6 days. In another application of the eddy correlation method to evaporation from a playa lake, Czarnecki (1990) concluded from his comparison of the many techniques applied to compute evapotranspiration that the eddy correlation method was the most reliable because the results were based on the most direct measurements. Although it is generally agreed that the technique has potential for use over dry lands and in arid regions where the Bowen ratio is susceptible to large errors, the expense and logistical complexity of the instrumentation currently limit its practical application.

Another approach to measure evapotranspiration is to place a canopy over the plant and measure air flow rate and water content of the inflowing and outflowing air. Sebenik and Thomas (1967) applied this technique by constructing a plastic tent over a tree. Stannard (1990) developed a portable hemispherical chamber, containing fans and a psychrometer, which fits over vegetation, such as grasses and small shrubs, to rapidly measure evapotranspiration over a period of less than two minutes. This technique has been successfully used to rapidly and directly measure evapotranspiration rates in small-scale studies such as along an arroyo system in New Mexico (Constantz *et al.*, 1994). However, because of the obvious logistical difficulties associated with measuring evapotranspiration over extensive areas, estimates of evapotranspiration based on micrometeorology are usually preferred.

While actual evapotranspiration is the quantity we seek, it is important to recognize that the climatological methods discussed above calculate the potential evapotranspiration, that is, the amount of evapotranspiration that would occur from a short green crop that fully shades the ground, exerts negligible resistance to the flow, and is not limited by water availability. Potential evapotranspiration, which reflects the maximum evaporative demand, closely approximates evaporation from an areally extensive, open body of water, but potential evapotranspiration cannot exceed lake evaporation under the same meteorological conditions, primarily because the actual availability of water from a vegetated surface is always less than from open water. Potential evapotranspiration can be estimated from lake evaporation data which can be calculated from the U.S. Weather Bureau, Class A evaporation pan measurements, available for many regions (Linsley *et al.*, 1975). It has been established that the potential evapotranspiration rate is approximately 0.7 times the pan evaporation rate, but this pan coefficient can vary from about 0.40 to 0.85, depending upon wind speed, fetch of green crop, and relative humidity (McWhorter and Sunada, 1977).

Note that the actual evapotranspiration in a landscape is rarely equal to potential evapotranspiration, except in very humid climates where the water table is within about 1 meter of land surface, or immediately after the profile is thoroughly wetted by an infiltration event. In fact, evapotranspiration is less than or equal to potential evapotranspiration even when the crop is adequately watered. To compute the actual evapotranspiration from potential evapotranspiration when the water supply is limited requires an additional calculation based upon plant type, water availability, and vegetation coverage on the landscape:

$$ET = K_c \, PET \qquad \text{(Equation 3-15)}$$

where ET is evapotranspiration, K_c is a crop coefficient, and PET is potential evapotranspiration. The crop coefficient is usually obtained by establishing an experimental relationship between evapotranspiration (measured with lysimeters) and potential evapotranspiration (calculated by a specific method) for some brief period. The dependence of the crop coefficient upon available water (AW) may be described by a formulation developed by Jensen *et al.* (1970):

$$K_c = K_{co} \frac{\ln\left(\dfrac{100\,AW}{AW_{max}} + 1\right)}{\ln 101} \qquad \text{(Equation 3-16)}$$

where K_{co} is the crop coefficient for a field where water is not limiting and

$$AW = (\theta - WP)D \qquad \text{(Equation 3-17)}$$

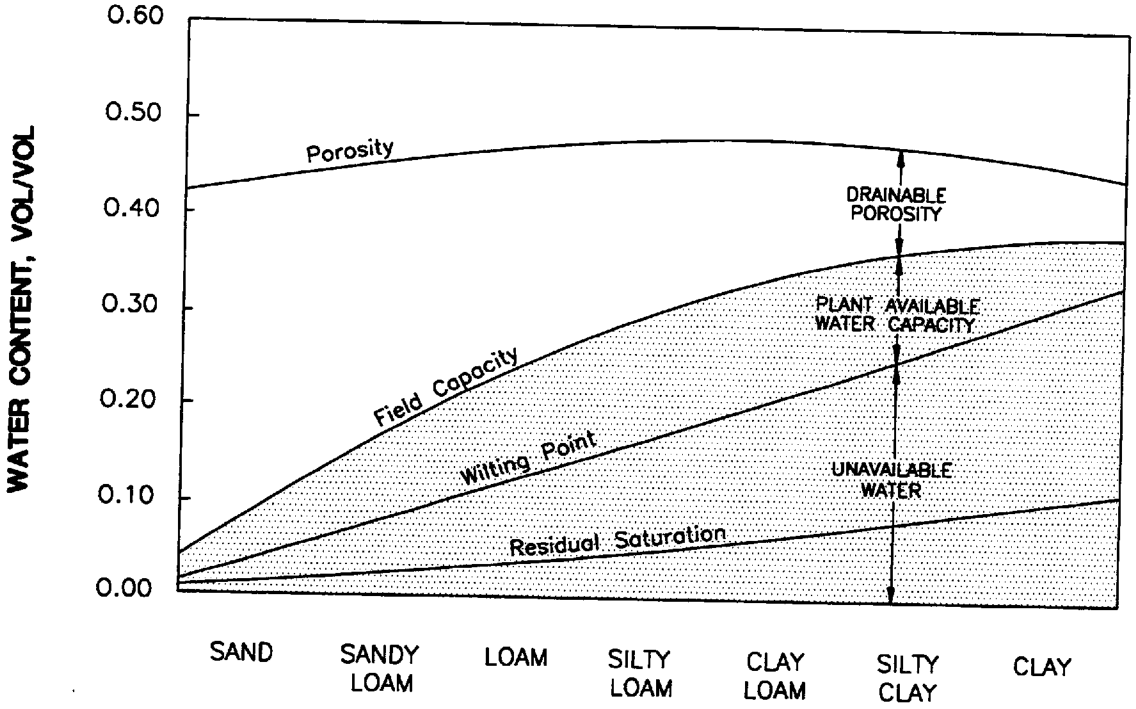
$$AW_{max} = (FC - WP)D \qquad \text{(Equation 3-18)}$$

where θ is the field water content, *FC* is the water content at the so-called field capacity, *WP* is the water content at the permanent wilting point, and *D* is the rooting depth.

Figure 3-4 shows typical values for porosity, field water content, field capacity, permanent wilting point, and available water for different soil textures. Equation 3-13 can be used to calculate a crop coefficient under circumstances of limited water supply. Because of their importance in agriculture, crop coefficients have been experimentally determined for many irrigated crops (Doorenbos and Pruitt, 1975). Values for crop coefficients vary widely from approximately 0.2 to 1.0 or more depending upon the type of crop, the crop's development stage, the time of year, climatic conditions such as relative humidity and wind, and latitudinal location. Values of plant-water use coefficients for selected native vegetation, presented in Table 3-2 (from McWhorter and Sunada, 1977), generally vary from 0.5 to 0.9, except for phreatophytes, which are 1.0 at all times.

Soil-Water Storage

The remaining component of the soil-water balance equation is the change in soil-water storage. Quantification of water storage changes requires repeated measurements of water content within the soil-water budget volume. Over a year or several years the water content change is usually small, and for some sites and



Source: Schroeder, *et al.*, 1994 (public domain)

Figure 3-4. Relation among moisture retention parameters and soil texture class.

Table 3-2. Estimated crop coefficients K_{co} for native vegetation with AW_{max} [From McWhorter and Sunada, 1977 (with permission)]

Vegetation	K_{co} NOV to MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT
Sagebrush-grass	0.50	0.60	0.80	0.80	0.80	0.71	0.53	0.50
Piñon-juniper	0.65	0.70	0.80	0.80	0.80	0.80	0.69	0.65
Mixed mountain shrub	0.60	0.67	0.81	0.85	0.82	0.74	0.65	0.60
Coniferous forest	0.70	0.71	0.80	0.80	0.80	0.79	0.75	0.71
Aspen forest	0.60	0.67	0.85	0.90	0.86	0.75	0.65	0.60
Rockland & miscellaneous	0.50	0.60	0.65	0.65	0.65	0.60	0.50	0.50
Phreatophytes	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00

climatic conditions, particularly arid and semiarid environments, the long-term change is negligible. Where calculations are made on an annual basis, this component of the water balance is usually ignored (Allison *et al.*, 1994).

Water storage in the vadose zone is simply the volume of water stored in the soil or rock to a particular depth. Because it is the change in water content that is required in the water balance, the method to measure water content is usually a geophysical method such as neutron probe logging. The use of weighing lysimeters, described in the previous section, also provides a means for calculating changes in soil-water storage. Neutron probe logging, and other methods described in the standard references mentioned previously, affords a means to repetitively and non-destructively measure water content changes at the same depths. Neutron logging is best applied in soils having textures of sand or finer sizes, and in monitoring water content following infiltration into soil that has a low initial *in situ* water content. When properly calibrated to laboratory measurements of initial water content, the neutron probe has been successful in determining relative changes in the *in situ* water content over time to within about 1 to 3 percent. It is important to note, however, that in uniform coarse-textured soil, considerable seepage can occur under conditions where the moisture content increases by only a percent or two. It is therefore possible that, under such circumstances, an increase in recharge could remain undetected by the neutron logging method.

From the discrete water content measurements obtained from neutron logging, the rate of change in the volume of water in storage over the time period of the water balance is calculated as:

$$S = \frac{1}{\Delta t}\int_0^D \theta \, dz \times Area \qquad \text{(Equation 3-19)}$$

The rate of change in water storage is simply calculated by determining the difference in total water storage at the beginning and end of the water budget period and dividing this by the time between monitoring events. In weighing soil lysimeters, the change in water storage within the monolith can be simply obtained from the change in mass divided by the water density.

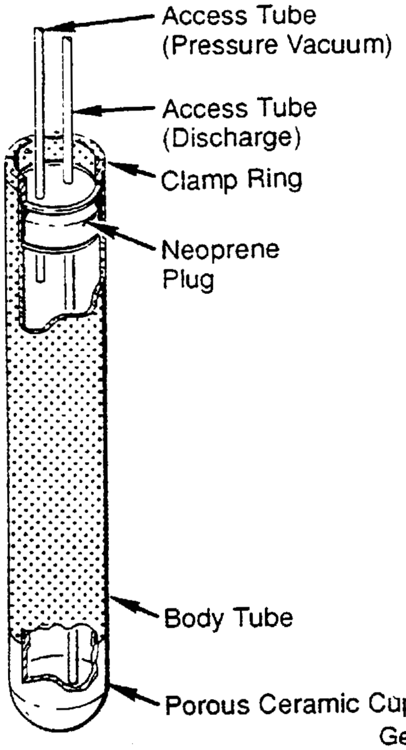
Lysimeter Measurements

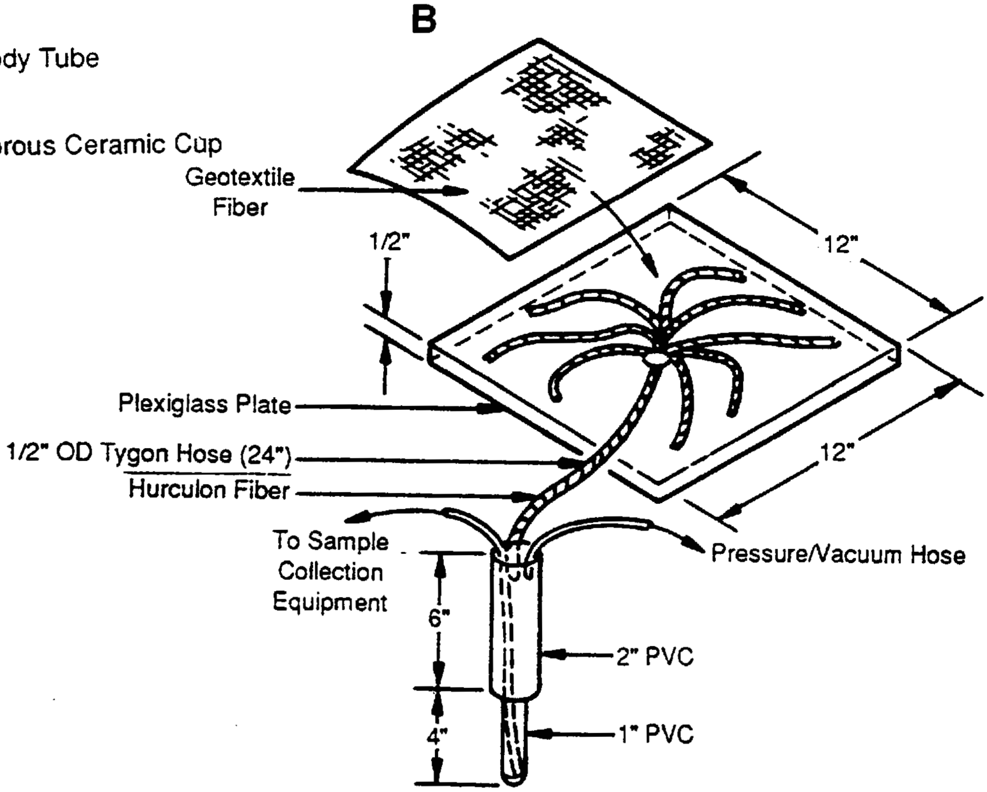
Lysimeters offer the only means of directly collecting deep drainage from the vadose zone. For the purpose of estimating recharge, lysimeters are used with the premise that the volume of water percolating below the root zone or zone of evapotranspiration can be extracted and measured over time, thereby directly quantifying the downward infiltration flux that would eventually become recharge. There are two basic categories of soil lysimeters useful for recharge: free-drainage samplers and the weighing and floating lysimeters already discussed in the evapotranspiration section.

A free-drainage sampler consists of a passive collection reservoir or chamber installed in the vadose zone to collect free-draining water from macropores which are intermittently saturated because of intense rainfall or flooding at the surface. Unsaturated flow percolating through the vadose may also be intercepted by free-drainage samplers, but no water will be collected until a sufficient head builds up to exceed atmospheric pressure or the suction applied by the sampler. Free-drainage samplers utilize a variety of passive collection devices, including metal pans (pan lysimeter), sand-filled funnels, glass blocks (glass block lysimeter), plastic troughs made of PVC cut lengthwise (trough lysimeter), corrugated steel pipe caissons (caisson lysimeter), or a lined, graded trench (trench lysimeter). Freely draining water collected in these vessels, or water accumulating at the impermeable base of the lysimeter, is then retrieved by pumping or applying a vacuum to a suction line within a collection bottle. The volume of water accumulated at the base of the lysimeter can be obtained by measuring water content change with a neutron probe, by measuring the volume of water recovered by a vacuum pump or pore liquid sampler, or by using piezometers to measure the depth of saturation.

Figure 3-5 illustrates two types of common devices to remove water from the base of a lysimeter: a porous-cup sampler (porous-cup suction lysimeter) and a vacuum plate sampler. The vacuum plate samplers are typically installed near the base of a horizontal trench. Vacuum plate samplers are available in diameters up to about 11 inches. A thorough review of porous suction samplers, including selection, installation and operation is provided in Wilson *et al.* (1995b).

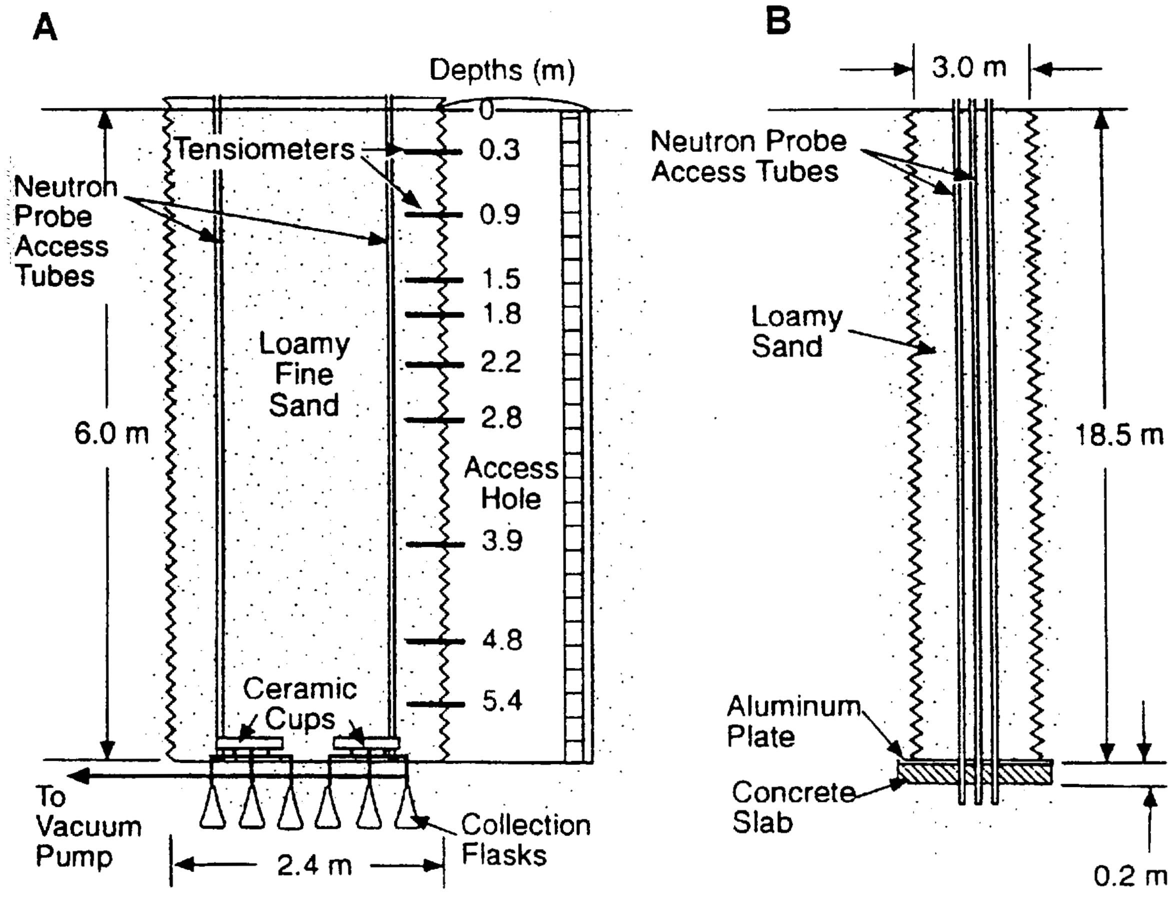
Figure 3-6 illustrates two different soil lysimeter designs in use at arid sites in Las Cruces, New Mexico and Hanford, Washington. The 18.5-m-deep lysimeter at the Hanford, Washington site, possibly the deepest one anywhere in the world, has been in place since 1972 (Gee *et al.*, 1994).





Source: Hornby *et al.*, 1986 (with permission)

Figure 3-5. A. Porous-cup sampler
B. Vacuum plate sampler



Source: Gee *et al.*, 1994 (with permission)

Figure 3-6. Schematic of soil lysimeters at two western desert sites
A. Las Cruces, New Mexico
B. Hanford, Washington

The principal advantage of soil lysimeters is that they are direct and precise. Gee and Hillel (1988) indicate that the deep drainage collected in a weighing soil lysimeter can be determined with a precision of about 1 mm. For deep water table conditions, one can capture the infiltrated water in a lysimeter constructed below the effective depth of evapotranspiration and infer that this water would ultimately reach the water table. At arid sites, the approach of using lysimeter data and direct water balance measurement techniques is considered more reliable than other estimation methods such as micrometeorology (Gee and Hillel, 1988; Gee *et al.*, 1994).

Although the soil lysimeter is a direct method, this approach still produces a recharge estimate that is subject to some uncertainty. For instance, the weighing, pillow, trench, and caisson lysimeters contain disturbed soil. And with the other types of lysimeters, soil disturbance during installation is unavoidable. It is impossible to completely preserve the *in situ* water content, pore geometry, stratification and macropore structures that can strongly influence soil-water movement. Months, years, or even decades may be required to completely reestablish vegetation with the same canopy and rooting characteristics as the surrounding native soils. Furthermore, unless the lysimeters are completed below the rooting depth, the calculated water flux will provide only an upper bound for actual recharge to the water table below. Additionally, the lysimeter data will represent recharge estimates only over the period of measurement, and therefore, depending on the prevailing precipitation and climatic conditions during this time, the estimates may not represent long-term behavior. In addition to the uncertainty of this method, disadvantages to consider in lysimetry include the considerable expense for construction, the disturbance of site facilities, operations and soils, and the long-term commitment to monitoring required to obtain representative data.

Darcy Flux Method

Darcy flux calculations also comprise a relatively direct physical means to calculate recharge from the vadose zone based on separate measurements of hydraulic gradient and hydraulic conductivity (Equation 3-2). In a previous section of this chapter, we presented an overview of some of the methods to obtain hydraulic conductivity and hydraulic gradient; consequently, they will not be repeated here.

The method usually is based on *in situ* measurements of pressure head or water content. These measurements are used to determine the hydraulic conductivity of the field soil at the *in situ* pressure head or water content. Pressure head may be measured with tensiometers in moist soil and with psychrometers in dry soil. Water content may be determined from core samples or neutron probe. The hydraulic conductivity at the field pressure head or water content is obtained from the K-Ψ curve for the soil. This curve is derived by separate field or laboratory tests, as described previously. The frequency of monitoring should be based on local conditions. For parts of the year, daily measurements may be required in humid climates, whereas monthly data should suffice in dry periods in areas of low rainfall.

Tensiometers or psychrometers placed at depth intervals within the vadose zone are also useful to compute the hydraulic gradient in Darcy's equation. However, below the root zone, where the pressure head is nearly constant with depth, it is a good assumption that the hydraulic gradient is approximately unity and flow is downward. Where this assumption is valid, measurements of hydraulic gradient are not required. A unit gradient is most likely to occur after prolonged periods of redistribution, or below the depth where evapotranspiration or thermal gradients may be significant. Consequently, where this assumption is reasonable and where vapor phase transport downward is negligible, the recharge rate is equal to the vertical, unsaturated hydraulic conductivity at the *in situ* moisture content or pressure head.

Under a unit gradient, the precision in the recharge rate computed by Darcy velocity *calculations is* comparable to that for the unsaturated hydraulic conductivity. Even with considerable care applied in the conductivity analysis, it is possible that errors in the recharge rate by this method could range from a factor of less than two or three in wet soils to more than an order of magnitude in dry soils. Where a unit gradient does not exist, the errors associated with this method could be even greater. In these situations, it is best to directly measure the hydraulic gradient using tensiometers or psychrometers.

Good examples of the Darcy flux method are illustrated in the field studies by Nnyamah and Black (1977) applied at a Douglas fir stand in British Columbia, by Sophocleous and Perry (1985) in Kansas, and by Stephens and Knowlton (1986) at a sparsely vegetated sandy site in New Mexico. Stephens and Knowlton had good success using the Darcy flux method to estimate recharge. Their analysis included tensiometric measurement of the hydraulic head gradient, which was found to be slightly greater than unity during much of the test period. However, the results using an assumed unit gradient were acceptably close to the results that incorporated the measured head gradient. Knowlton *et al.* (1992) report good agreement between Darcy-flux-based recharge estimates and chemistry-based estimates.

<u>Plane of Zero Flux Method</u>

The plane of zero flux (or zero flux plane) method relies on locating a depth in the soil profile where the hydraulic gradient is zero. A zero flux plane develops during redistribution of a pulse of infiltrated water and is usually present at times when evapotranspiration exceeds rainfall. Above this plane or surface, soil-water movement is upward, and below this plane, water moves downward. Where a zero flux plane is present, it can be used to separate water content changes due to evapotranspiration and drainage. Any change in water content below the plane of zero liquid flux reflects drainage that eventually becomes recharge. To determine the water flux below the zero flux plane, the following equation is applied to volumetric water content profiles measured over different time periods:

$$q = \frac{1}{\Delta t} \int_{D_o(t)}^{D} \theta dz \qquad \text{(Equation 3-20)}$$

Recharge is calculated by summing the water content changes over the depth interval, D to D_o, below the zero flux plane. For accurate results, the monitored soil column should extend from the zero flux plane downward to a depth where the moisture content does not fluctuate appreciably. Implementing the method requires instrumentation of soil-water potential sensors such as tensiometers to locate where the hydraulic gradient is zero. Alternatively, the zero flux plane can sometimes be inferred from water content data collected by neutron probe, frequency domain reflectometry probe, time domain reflectometry probes, or other *in situ* water content measuring devices. The recharge flux for the time interval between measurements is then represented by the integrated change in moisture content in the monitored section of the soil profile.

The zero flux plane method was applied by Dreiss and Anderson (1985) to quantify deep water percolation beneath a land treatment facility. The study used three replicate sets of tensiometers and neutron probes to measure pressure head and moisture content. Each set contained four or five tensiometers and one neutron access tube. The study concluded that the zero flux plane method provided reasonable estimates of cumulative seasonal recharge when weekly measurements of moisture content and pressure head were used. The total error for the recharge estimate was calculated to be within 15 percent of the total moisture input (total precipitation plus change in storage). Allison *et al.* (1994) point out that the zero flux plane method breaks down when the hydraulic gradient is downward throughout the profile, and this occurs at times when there is significant recharge. During these periods, the Darcy flux method can be applied.

Soil Temperature Methods

The soil temperature profile depends on the geothermal gradient, the period and amplitude of the atmospheric temperature changes, the thermal properties of the vadose zone, and the water flux through the vadose zone. Three different techniques utilizing temperature gradient data have been developed to provide estimates of the soil-water flux or recharge.

Bredehoeft and Papadopulos (1965) developed a type-curve method to compute the flux by fitting steady-state temperature profile data from the saturated zone to the theoretical type curves. This approach has been applied by Cartwright (1970, 1979) and by Boyle and Saleem (1979). Applications of the method have been primarily limited to relatively deep aquifers where the temperature increases with depth (i.e., the temperature gradient is upward). Results have shown that recharge fluxes estimated with type curves agree well with those obtained from hydraulic data and basin water balance calculations.

Stallman (1965) developed an analytical solution for computing the steady downward flux from sinusoidally varying surface temperatures. Taniguchi and Sharma (1993) built upon Stallman's analysis and computed recharge from *in situ* measured changes in temperature. After applying the method to forested sites in western Australia, they concluded that their temperature difference method produced reasonable results when the annual recharge rate is low (less than approximately 200 mm/yr) and the temperature is measured at depths less than several meters.

A third temperature method for estimating soil-water flux was developed by Wierenga *et al.* (1970) based on a steady-state soil-heat balance. When water infiltrates to a particular depth, the mean temperature of the soil within this depth will change. By *in situ* measurement of the initial and final soil temperature and the temperature of the infiltrating water, the amount of infiltrated water is obtained. Because infiltration-induced temperature changes are not readily detected below about 2 meters, the method is best suited to determining soil-water flux near the surface rather than recharge (Taniguchi and Sharma, 1993).

Although the instrumentation requirements for implementing temperature methods are quite simple, the methods have not been widely used to date. One reason for this may be the difficulty and uncertainty associated with determining thermal properties of the media.

Electromagnetic Methods

In the recharge-electrical conductivity model, groundwater recharge is related to soil texture, water content, and soil-water conductivity. In principle, electrical conductivity generally increases as the clay fraction decreases; greater recharge is likely in coarse soils free of clay. For a given soil texture, the maximum recharge should occur where the water content is greatest, that is, where the electrical conductivity is greatest. However, changes in soil salinity may influence the electrical conductivity measurement. As salinity of a soil decreases, the electrical conductivity decreases, and this effect on conductivity is similar to that induced by coarsening soil texture and decreasing water content. Therefore, it is difficult to identify areas of favorable recharge characteristics using electromagnetic methods where soil salinity is highly variable.

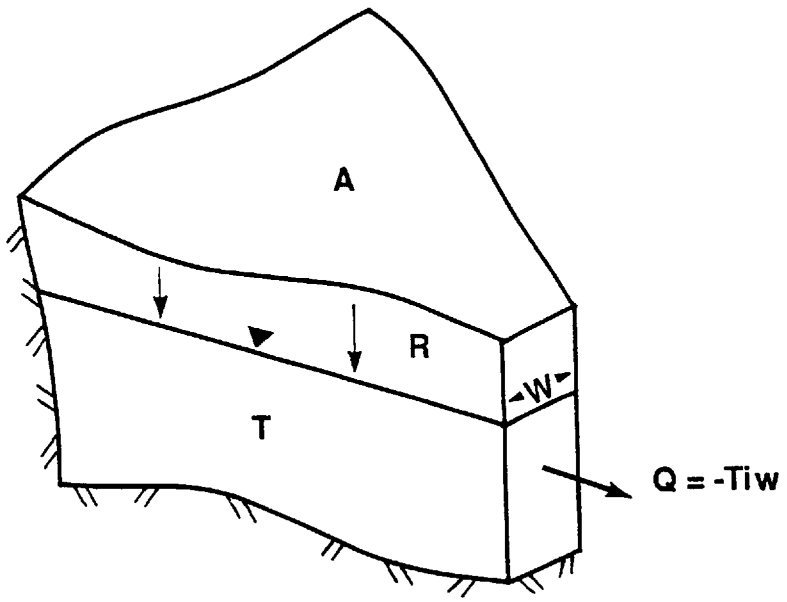
There have been few attempts to determine recharge rates from electrical conductivity measurements. Cook *et al.* (1992) applied frequency domain and time domain electromagnetic measurements, as well as the direct current resistivity method, to delineate recharge zones at a site in southeastern Australia. By comparing the geophysical analyses with independent estimates of recharge, they concluded that soil texture, specifically clay content, is the principal reason for the correlation between electrical conductivity and recharge. At this site in Australia, the soil-texture and solute effects dominated over the effect of water content, thereby causing an inverse relationship between recharge and electrical conductivity. Air-borne electromagnetic surveys were also conducted in this same area to identify recharge zones (Cook and Kilty, 1992) based on this relationship. From limited testing it appears that the electromagnetic methods are suited for reconnaissance-level investigations to identify recharge areas where more quantitative methods for recharge should be applied.

Groundwater Basin Outflow Method

Darcy's equation can be applied to aquifers to compute recharge by dividing the flow rate through an aquifer cross section by the land area contributing to recharge (e.g., Theis, 1937). In this conceptual model, vadose zone percolation is assumed to be the only source of recharge that is uniformly distributed over a basin having well-defined, impermeable lateral and lower boundaries. We also assume that at some downgradient location, the groundwater flow rate, Q, leaving this part of the basin (i.e., underflow out of the basin) is obtained from a form of Darcy's equation:

$$Q = -Tiw \qquad \text{(Equation 3-21)}$$

where T is the aquifer transmissivity, i is the hydraulic gradient of the aquifer (a negative quantity when the head loss is taken in the direction of flow), and w is the width of the aquifer where underflow is calculated (Figure 3-7). Then, the average recharge rate for the basin would be:



Notes:
A = Basin area
R = Areal recharge
Q = Steady-state ground-water flow out of basin
T = Aquifer transmissivity
i = Hydraulic gradient
w = Aquifer width perpendicular to flow direction

Source: Stephens, 1995 (with permission)

Figure 3-7. Diagram of the relationship between recharge and groundwater basin underflow.

$$R = \frac{Q}{A} + S_y \frac{\Delta h}{\Delta t} \qquad \text{(Equation 3-22)}$$

where Q is obtained from Equation 3-21, A is the upstream surface area of the watershed where recharge could occur, S_y is the specific yield, and $\Delta h/\Delta t$ is the average head change during the time interval. The method has the potential to work well if the basin boundaries and areas of recharge are well defined and the transmissivity can be determined with reasonable accuracy.

Theis (1937) employed a steady-state version of Equation 3-22 to calculate recharge to the Ogallala aquifer in eastern New Mexico and Texas. Maxey and Eakin (1951) applied this technique to 22 groundwater basins in Nevada which they assumed were in hydrodynamic equilibrium, such that the recharge simply equaled the outflow or discharge from the basin. They then correlated the recharge with the elevation and mean annual precipitation of the basin and developed relationships that they suggested could be useful to estimate recharge for other basins simply on the basis of basin elevation. However, Watson *et al.* (1976) indicated that geology, hydrologic characteristics of the consolidated and unconsolidated rocks, antecedent soil moisture, and vegetation strongly affect recharge, and that unless these are incorporated, considerable uncertainty in results from the Maxey-Eakin method would remain. More recently, Avon and Durbin (1994) concluded that the Maxey-Eakin method produced recharge estimates comparable to estimates obtained independently by water balance and groundwater model methods.

<u>Water-Level Fluctuations</u>

In undeveloped, unconfined aquifers not subject to tidal influences, water levels fluctuate primarily in response to recharge from precipitation, discharge by basin outflow, and evapotranspiration. As a consequence of diffuse recharge, the slope of the water table and the transmissivity remain nearly constant during normal seasonal water-level fluctuations in most aquifers, that is, except for periods of extreme precipitation or prolonged ponding. The annual cycle of change in saturated thickness of the aquifer due to natural recharge and discharge processes usually has an insignificant effect on transmissivity. Throughout the year, water levels decline during periods of little to no recharge and rise when recharge exceeds basin outflow. On average, however, the mean water level in undeveloped aquifers remains virtually unchanged year after year.

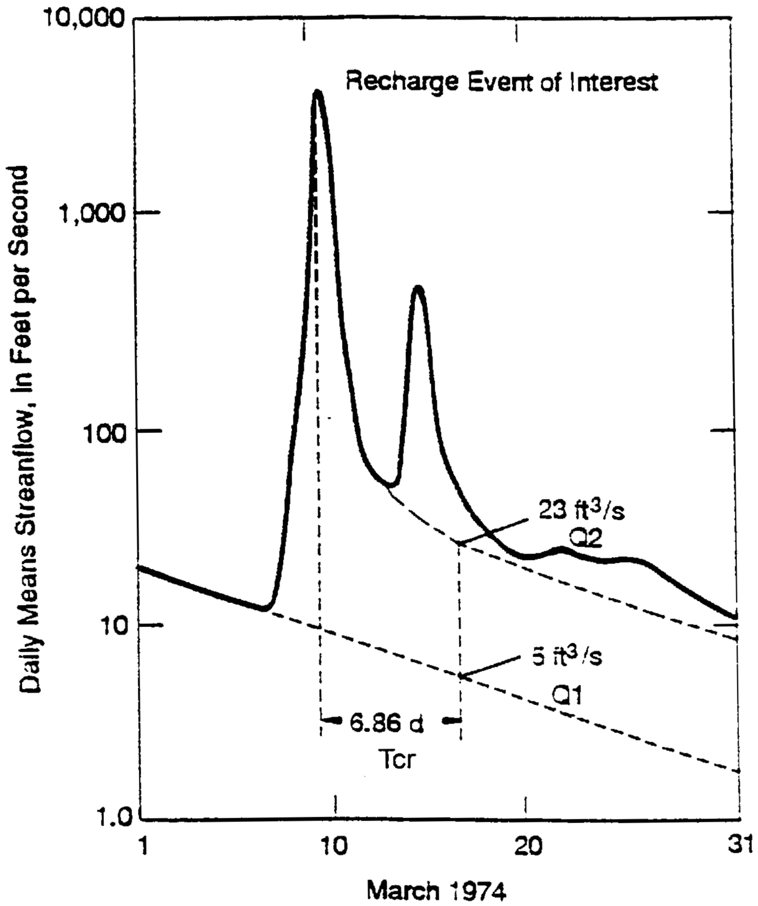
For this case, Theis (1937) suggested that the recharge rate could be determined during periods of no recharge by multiplying the annual rate of water-level decline by the specific yield, inasmuch as over the long term the annual basin outflow equals the annual inflow. Similarly, during recharge periods, the rate of water-table rise times the specific yield gives the recharge rate, provided that one measures the water-table rise from an extrapolation of the preceding water-table recession curve that represents the rate of water-level decline during the prior period of reduced or absent recharge.

Sophocleous (1991) suggested a simple modeling approach to obtain recharge from precipitation records, vadose zone water balance analysis, and water-level fluctuations in wells. In this analysis, called the hybrid water-level fluctuation method, one establishes field instrumentation for determining recharge from a soil-water balance at one or more locations. At these same locations within the basin, the water-level response to precipitation is determined from a monitor well hydrograph. Knowing recharge from the soil-water balance and water table response, one can obtain the aquifer specific yield, if basin outflow during the period is neglected. Multiple instrumented sites within a basin provide some basis for determining a mean specific yield or its spatial variability. Recharge throughout the basin at sites with monitor wells, as opposed to soil-water balance *instrumentation*, can be calculated simply from the water-level fluctuations and the estimate of specific yield. Sophocleous (1992) applied the hybrid water-level fluctuation method to the Great Bend Prairie of central Kansas to identify zones of similar recharge within the region.

Stream Gauging

Stream gauging data can be very valuable in quantifying recharge in humid climates where perennial streams are fed by ground water. Stream hydrographs are generally characterized by a series of peaks followed by recessions. The peaks usually represent surface runoff, interflow and bank storage, whereas the recession curve represents primarily discharge from the aquifer to the river. In ground-water basins that discharge to perennial streams, the ground-water recharge is approximately equal to the surface-water discharge minus direct surface-water runoff. The approach determines surface-water runoff from stream hydrographs. It furthermore assumes that the aquifer discharge to the stream (baseflow) is caused by diffuse recharge from rainfall over the surface-water drainage basin, and is valid for periods between major rainfall events, where no surface runoff directly contributes to streamflow.

There have been several approaches to quantify recharge from steamflow measurements (e.g., Meyboom, 1961; Rorabaugh, 1964; Daniel, 1976). Figure 3-8 illustrates a recent method to compute recharge from a runoff event, based on the upward displacement of the recession curve at some critical time, T_{cr}, following a flood peak. According to Rutledge and Daniel (1994), the critical time is calculated from:



Explanation

——— Daily Streamflow

---------- Extrapolated Ground-water Discharge

Procedure

1. Compute Recession index, (32 d/log cycle)
2. Compute Critical Time, Tc (0.2144xK, or 6.86 d)
3. Locate Time that is 6.86 Days after Peak
4. Extrapolate Pre-event Recession to Q1 (5 ft^3/s)
5. Extrapolate Post-event Recession to Q2 (23 ft^3/s)
6. Compute Total Recharge by Equation 18

$$\frac{2 \times (18\ ft^3/s) \times 32\ d}{2.3026} \times \frac{86400\ s}{1\ d} = 4.32 \times 10^7\ ft^3$$

Source: Rutledge and Daniel, 1994 (with permission)

Figure 3-8. Procedure for using the recession-curve-displacement method to estimate recharge in response to a recharge event.

$$T_{cr} = 0.2144K_i \qquad \text{(Equation 3-23)}$$

where K_i is the recession index, the time in days required for groundwater discharge to the stream to decline by one log cycle of flow. By computing the flows at time T_c from the recession curve preceding the flood peak, Q_1, and the recession curve following the flood peak, Q_2, the recharge value from the rainfall event is:

$$R = \frac{2(Q_2 - Q_1)}{2.3026} K_i \qquad \text{(Equation 3-24)}$$

The recharge rate is then determined by dividing the recharge volume by the surface area of the drainage basin. This method seems to offer the greatest potential for success where streams are not managed to control water storage and diversions, where the streams are well-connected to the aquifer, and where there is little recharge due to snowmelt. Rutledge and Daniel (1994) indicate that an automated version of this method greatly reduces the labor required to evaluate records and has been successfully applied to 15 streamflow gauging stations in the eastern United States.

Chiew *et al.* (1992) developed an integrated surface-water and groundwater model to compute recharge for a nonirrigated area in southeastern Australia that is a tributary to the Murry River. The surface-water component was HYDROLOG and the aquifer component was AQUIFEM-N. The model was calibrated against heads and streamflow, with recharge estimated as an output from the calibrated model.

CHEMICAL METHODS FOR DETERMINING DIFFUSE RECHARGE

Chemical methods provide indirect means of calculating recharge by tracking water movement through the vadose zone and ground water. The following paragraphs describe the various chemical and isotopic techniques available to evaluate recharge.

Chemical Methods in the Vadose Zone

Among the chemical methods for calculating recharge from the vadose zone, there are stable and radioactive isotopes, including tritium, chlorine-36, oxygen-18, and deuterium, together with chloride mass balance. One advantage of some of these methods is that the analysis may represent an integration of hydrologic events over decades or even tens of thousands of years. Another advantage is that the data are derived from one-time, *in situ* sampling, without need for field instrumentation for monitoring. These can be important considerations in selecting an appropriate method, especially if quantifying long-term natural recharge is the objective. The following discussion focuses on the chemical methods that are applied to the vadose zone. Later, we consider chemical methods applied to aquifers.

Tritium. Tritium (^3H), a radioactive isotope of hydrogen with a half-life of about 12.4 years, is well suited as a hydrologic tracer because it is part of the water molecule. During the atmospheric nuclear testing that began in the 1950s, the tritium in the atmosphere increased substantially over a relatively short time, culminating in the period 1963-1964 (Phillips *et al.*, 1988). This tritium pulse rapidly circulated worldwide, with primary deposition in the northern hemisphere where most of the testing occurred. After the vapor condensed and fell as precipitation, the record of tritium has been preserved in atmospheric water that infiltrated the soil profile. Recharge, or more precisely, net infiltration, is obtained from the depth to the center of mass of the tritium pulse, L, with the following equation:

$$R = \theta \frac{L}{\Delta t} \qquad \text{(Equation 3-25)}$$

where θ is the mean water content through depth, L, and Δt is the time increment. To analyze for tritium in the unsaturated zone, soil samples are typically collected using a core-barrel or split-spoon sampler. Care must be taken to eliminate moisture loss during sample handling. Tritium analyses are available through commercial or private university laboratories. The water for tritium analysis is typically obtained using a vacuum distillation process, and the tritium content is counted using standard liquid scintillation techniques.

Because tritium can move in the vapor as well as in the liquid phase, strong temperature gradients may influence the tritium peak through thermally-driven, vapor-phase migration (Knowlton *et al.*, 1992). Also bear in mind that the quantity calculated here approximates recharge only if the tritium has migrated below the root zone; otherwise this analysis represents a soil-water flux that may exceed the actual recharge. In fact, Tyler and Walker (1994) demonstrated that tracer methods applied within the root zone significantly overestimate the deep soil-water flux, because the roots induce nonconstant soil-water velocities within the root zone. They suggest that tracers, such as tritium, can be effective when the recharge exceeds 10 percent of the annual precipitation.

Chlorine-36. Another tracer of soil-water flux or recharge is chlorine-36 (^{36}Cl). This is a radioactive isotope with a half-life of about 300,000 years, produced as a byproduct of thermonuclear testing near the oceans in the 1950s (Bentley *et al.*, 1982). Chloride is very stable in the environment and enters the hydrologic cycle as the chloride ion dissolved in water and as a component of dust fallout. Because it is soluble and nonvolatile, chloride is an excellent tracer for liquid-phase transport;

however, it is also subject to the effects of anion exclusion which result in more rapid movement of the chloride anion relative to other conservative tracers.

Recharge can be determined from the depth of the chlorine-36 peak, in the same manner as that described above for tritium. The practical application of the chlorine-36 method at many sites is probably limited because it requires analysis by a tandem accelerator/mass spectrometer, which is a highly specialized instrument commercially available only at a few research institutions. Phillips *et al.* (1988) first applied the chlorine-36 method, along with the tritium technique, to study both liquid and vapor transport and to evaluate the suitability of these tracers for quantifying recharge in very dry desert soils. In two cases where both tracers were measured in the same profile, the tritium pulse had penetrated approximately 1 to 2 m deeper than the chlorine-36 peak, which was retained near the soil surface. The difference in the relative transport rates of tritium and chlorine-36 was attributed to low water content and fluctuating temperature gradients which enhance vapor movement of tritiated water relative to simple advection of the chlorine-36. A comparison of the recharge estimates calculated by the tritium peak, Darcy flux, chlorine-36, and chloride mass balance methods, indicated good agreement between the Darcy flux (0.70 cm/yr) and the tritium peak (0.84 cm/yr) calculations, and an underestimation of recharge by the chlorine-36 peak and chloride mass balance methods.

<u>Chloride Mass Balance</u>. The chloride mass balance method relies upon the slow accumulation in the soil profile of natural chloride that dissolves in precipitation and infiltrates. The concentration of chloride in precipitation typically decreases with increasing inland distance from the coasts. The expected chloride pattern in the soil profile, at least in areas of modest precipitation, is that chloride increases with increasing soil depth as water is extracted by the plant roots, but below the root zone, the chloride concentration is expected to be constant where the deep

percolation migrates toward the water table. The chloride concentration increases in proportion to the ratio of precipitation to recharge (Allison and Hughes, 1978):

$$R = P\frac{C_p}{C_s} \qquad \text{(Equation 3-26)}$$

where P is the average precipitation rate, C_p is the chloride concentration in precipitation and dry fallout, and C_s is the average soil chloride concentration below the root zone. This simple, one-dimensional form of the chloride transport equation assumes that the chloride-laden water moves as piston flow, that is, without liquid dispersion or macropore flow. This simplification is usually appropriate because adequate information for more comprehensive transport equations is typically not available.

The three fundamental assumptions of the chloride mass balance method are (1) all chloride originates from atmospheric deposition, (2) the only long-term sink for chloride is downward advection, and (3) chloride behaves conservatively during soil-water transport. Additionally, it is assumed that vadose zone flow is at steady-state. Chloride patterns that depart from this model may produce a bulge in the chloride concentration at some depth in the profile, and below this, the concentration decreases to approach a near constant value. Based on the first assumption mentioned, it should be noted that where chloride from other anthropogenic sources may be present in the soil profile, such as at oil production facilities, this technique would probably not be useful.

Phillips (1994) discussed possible explanations for the "chloride bulge" in desert soil profiles such as preferential or bypass flow in macropores, but suggested that the low concentration of chloride at depth most likely reflects greater recharge during a wetter paleoclimatic period when the indigenous plants were less efficient at capturing the soil moisture. An alternative hypothesis only recently advanced suggests that the chloride bulge is best explained by ultrafiltration, a process that causes chloride anions to migrate more slowly than the water (McCord *et al.*, 1994). If ultrafiltration does indeed affect migration of chloride anions in the vadose zone, many previous infiltration studies based on the chloride mass balance method may underestimate the recharge component. Additional research is required to investigate this phenomenon, especially in arid environments.

<u>Stable Isotopes</u>. An isotope is a variation of an element produced by differences in the number of neutrons in the nucleus of the element; hence isotopes of an element have different masses. The two stable, or non-radioactive, isotopes of hydrogen (^{1}H and ^{2}H or deuterium (D)) and the three stable isotopes of oxygen (^{16}O, ^{17}O and ^{18}O) form part of the water molecule, and analyses of their concentrations in natural waters have long been used to trace movement of water in the subsurface. It is well established that the isotopic composition of precipitation at a particular location will vary seasonally and with individual storms. The isotopic composition of precipitation will also vary between locations depending upon climate and elevation. Nevertheless, the composition of all precipitation generally falls on a straight line of a plot of δD versus $\delta^{18}O$ (where δ is the relative difference of the isotopic ratios in precipitation versus standard mean ocean water [SMOW] expressed in parts per thousand). This line is called the meteoric water line.

The stable isotope concentration of the precipitation can be modified subsequent to infiltration, and this signature of the soil water reveals important information about recharge. Evaporation of soil water leads to a fractionation of the stable isotopes deuterium and oxygen-18. When water evaporates, the heavier atoms tend to remain behind in the liquid phase, thus leading to an enrichment in the concentration of the heavier isotopes in the residual liquid, and lighter isotopes fractionate into the vapor phase. When the water vapor condenses, the condensed liquid is more concentrated in the heavier isotopes than the residual vapor. The isotopic enrichment and shape of the isotope profile in the soil depend upon the net infiltration and evaporation rates, soil-water status, and diffusive properties of the liquid and vapor. Based on these and other considerations, Barnes and Allison (1983) developed a theoretical model to predict how the stable isotopes should be distributed in the soil. Knowlton (1990) built on this theoretical analysis and developed an equation for recharge based on measured deuterium isotope concentrations:

$$R(\delta_D - \delta_D^{Rec}) = \left(\frac{hN_{sat}D^v}{\rho}\right)(\varepsilon_D + \eta_D)\left(\frac{d[\ln(hN_{sat}(\varepsilon_D + \eta_D))]}{dz}\right) - \left(D + \frac{hN_{sat}D^v}{\rho}\right)\frac{d\delta_D}{dz} \qquad \text{(Equation 3-27)}$$

where R = recharge rate (m/s)

δ_D = standardized isotope ratio at any depth (per mil)

δ_D^{Rec} = standardized isotope ratio of the recharge water at depth (per mil)

h = relative humidity

N_{sat} = saturated water vapor density in air (kg/m^3)

D^v = effective diffusivity of water vapor in air (m^2/s)

ρ = density of liquid water (kg/m^3)

ε_D = equilibrium enrichment factor

η_D = diffusion ratio excess

z = depth below land surface (m)

D = effective self-diffusion coefficient of water (m^2/s)

This method requires measurement of the distribution of deuterium or oxygen-18, and water content with depth in order to estimate the recharge rate. The remaining model parameters can be easily estimated or obtained from standard reference books. Soil samples can be collected during standard coring operations, stored in Mason® jars, and subjected to a vacuum distillation procedure to extract the soil water. The soil-water extraction process for stable isotope analysis is not a routine procedure offered by commercial laboratories; thus the application of the technique is limited. This method holds promise, but has only recently been developed and requires additional applications in the field.

Allison *et al.* (1984) used another technique to assess recharge with stable isotope measurements of pore liquids collected from the vadose zone. They assumed that if the rainfall events are uniform throughout the year and the evaporation rate is constant, then the isotopic enrichment increases linearly with the square root of the time since the last rainfall. From this, they developed a relationship between recharge and the magnitude of enrichment or shift between meteoric line and composition of deuterium in soil water. The method apparently has not been widely used thus far and requires further validation for a variety of soils. Nevertheless, it may be a useful and simple tool to estimate recharge in areas of low precipitation.

<u>Chemical Tracers in Aquifers</u>

Chemical tracers commonly occurring in aquifers that permit quantification of recharge include tritium (^{3}H), tritium/helium-3 ($^{3}H/^{3}He$), carbon-14 (^{14}C), chlorine-36 (^{36}Cl), and chlorofluorocarbons. In principle, these tracers determine the age of the groundwater, which in turn permits calculation of groundwater travel time:

$$R = vn_e = L\frac{n_e}{t_a} \qquad \text{(Equation 3-28)}$$

where v is the component of average linear velocity, n_e is the effective porosity, L is the distance along the flow path, and t_a is the travel time or age of the groundwater at the distance L.

In an unconfined aquifer analysis, the groundwater flow velocity is obtained from the apparent tracer age gradients in the direction of flow. Flow velocity can be determined from one date at one depth in the aquifer if the distance from the point of recharge entry into the aquifer to the sampling location is well established. In this analysis, which is most applicable to shallow, unconfined aquifers, the travel time through the vadose zone is considered negligible compared to that in the aquifer; however, where it is known, the travel time through the vadose zone can be added to the travel time through the aquifer. Alternatively, the age of groundwater from a sample collected at the water table would be useful to evaluate the travel time and mean velocity through the entire vadose zone at the field water content in order to estimate recharge.

Two or more nested wells located along the groundwater flow path can be sampled to obtain groundwater ages and calculate the flow velocity. The effective porosity of the aquifer is usually estimated based on the lithology of the aquifer materials, and has commonly been assumed to be 0.3 to 0.4 for unconsolidated sediments (Solomon *et al.*, 1993). The hydrogeology of the site must be well understood prior to applying this technique, because the method is based on the presumption that the direction of flow in the aquifer is known with reasonable confidence.

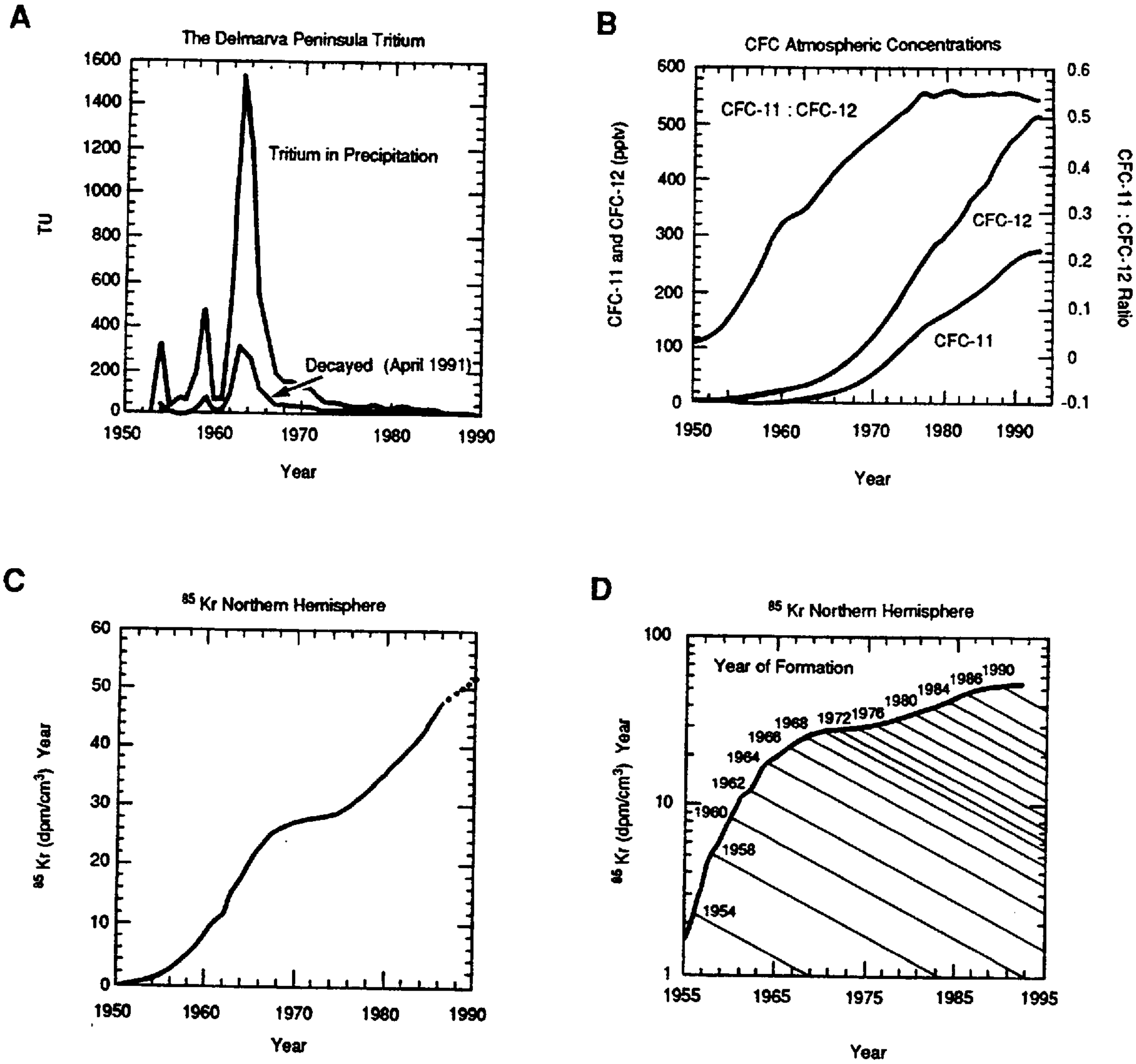
The application of chemical tracers for determining groundwater ages and recharge rates has several common approaches and difficulties. The simple models that calculate recharge from groundwater age assume that the tracers move in a piston displacement process which neglects liquid dispersion. In fact, however, mechanical mixing and diffusion within the porous media serve to decrease the input concentrations. If hydrodynamic dispersion is not taken into account by mathematical modeling, the calculated age will exceed the true age, but in permeable aquifers where advective transport dominates, dispersive effects are not significant, especially if the tracer input is relatively constant in time (Solomon and Sudicky, 1991). Also, the true age of a groundwater sample may be over- or under-estimated due to commingled water samples from different parts of the aquifer where the groundwater has different ages.

The following subsections briefly summarize some of the currently used methods for groundwater age dating.

Tritium. Tritium in aquifers is derived from both natural and anthropogenic sources. Tritium is produced naturally when cosmic rays interact in the upper atmosphere with nitrogen. Precipitation naturally contains approximately 5 tritium units (TU) (Mazor, 1991); however, precipitation that entered the soil prior to 1952 would have decayed by now to concentrations near or below the analytical detection limit. The thermonuclear testing that began in the 1950s generated peak concentrations in precipitation ranging from about a few hundred to about 10,000 TU. An example of measured tritium in precipitation for the Delmarva Peninsula on the eastern coast of the U.S. is shown in Figure 3-9A. The apparent age, t_a, of the water sample is:

$$t_a = - t_{1/2} \ln(\frac{A}{A_o}) \qquad \text{(Equation 3-29)}$$

where $t_{½}$ is the half-life in years, A is the activity of the sample at the time the precipitation or surface runoff entered the subsurface, and A_0 is the measured activity in the sample.



Source: Ezwurzel *et al.*, 1994 (with permission)

Figure 3-9. A. Average tritium concentration in precipitation on the Delmarva Peninsula.

B. Atmospheric concentration of CFC-11 nd CFC-12 in parts per trillion volume per volume air and the ratio CFC-11:CFC-12.

C. Krypton-85 specific activity (i.e., the ratio of ^{85}Kr to stable krypton in disintegrations per minute per cubic centimeter krypton) in the troposphere of the northern hemisphere between 40^{o} and 55^{o}N as a function of time.

D. Krypton-85 specific activity plotted on a logarithmic scale. Diagonal lines represent radioactive decay after ground water is isolated from the atmosphere.

Owing to the rather short half-life of tritium, its usefulness as a dating method will soon expire. Even where tritium is detectable, there is considerable uncertainty in the calculated age. The tritium distribution at the source of recharge is rarely known, but where it has been measured, tritium varies considerably with location, year, and season. This results in significant uncertainty in the initial tritium activity of the water (A) when it entered the subsurface. Additionally, the groundwater flow path taken by the tritium is almost always complex, and some samples may represent composite paths. For these reasons, tritium, where detectable, usually has only semiquantitative significance. That is, detectable tritium suggests that the groundwater sample contains at least some portion of water derived from precipitation that fell after 1952 (Mazor, 1991).

An analytical advance of the tritium method uses helium-3, the stable daughter of tritium decay. Tritium-bearing recharge will produce dissolved helium-3 that increases in concentration along the flow path as the tritium concentration diminishes. Combined measurements of tritium and helium-3 allow one to determine the age of the groundwater, provided that other sources of helium are negligible or can be measured to quantify the tritiogenic portion of the helium-3 (e.g., Ekwurzel *et al.*, 1994; Schlosser *et al.*, 1989). In most shallow groundwaters, subsurface sources of helium-3 are insignificant.

Measuring the ratio of tritium to helium-3 in soil water offers several advantages over using tritium data alone. First, because total tritium plus helium-3 concentrations are not affected by radioactive decay, the combined signal can be used as a stable tracer. This enables one to distinguish between the effects of dispersion and radioactive decay and also makes it possible to study the penetration of the bomb tritium peak into deeper layers of the subsurface even after complete decay of tritium to helium-3 has occurred. Additionally, the initial concentration of tritium in the atmosphere does not need to be estimated. As discussed above, the tritium concentration record in the atmosphere contains many spikes, at least prior to the conclusion of atmospheric testing, and the record is rarely known in detail at locations where recharge is to be calculated. Solomon and Sudicky (1991), who evaluated the effects of hydrodynamic dispersion and the nature of the tritium input function on the reliability of ages, concluded that the tritium/helium-3 method can be used to accurately date shallow groundwater with ages ranging from 0 to about 50 years.

Chlorofluorocarbons. Chlorofluorocarbons (CFCs), including CFC-11 (CCl_3F) and CFC-12 (CCl_3F_2), are chemically stable man-made volatile compounds that have been manufactured since the 1940s and 1930s, respectively, for use as aerosol can propellants, foaming agents in plastics, refrigerants, and solvents (Dunkle *et al.*, 1993). Their release into the atmosphere, documented worldwide, produced a steady increase in concentration in the atmosphere (Figure 3-9b). The atmospheric concentrations of CFCs in precipitation are governed by Henry's law. Because the CFCs are relatively stable in the atmosphere and subsurface, the CFC concentration in groundwater recharge should increase over time as a result of the increasing atmospheric production. The age of a groundwater sample analyzed for CFCs is determined simply by comparing the measured concentration in groundwater with a graph of the atmospheric water concentrations over approximately the past 50 years (e.g., Busenberg and Plummer, 1991).

A number of factors should be kept in mind when interpreting groundwater ages from CFCs. First, the temperature of the atmosphere must be determined, because temperature affects the CFC partitioning between the atmospheric gas and water phases. As the dissolved CFCs migrate through the vadose zone, there is opportunity for additional phase partitioning, depending upon the CFC partial pressure and gas-phase advection of CFCs. Also in the vadose zone, the CFCs may be sorbed by organic carbon in the soil. The same sorption processes may occur in the aquifer as well. Biodegradation may also reduce the CFC concentration in the aquifer, especially under anaerobic conditions (Busenberg and Plummer, 1992). Furthermore, contamination of the sample must always be guarded against. This is particularly important for older waters that have exceedingly low concentrations. Sample contamination for determining age may occur due to improper cleaning of sampling and analytical equipment, sample exposure to the atmosphere, or contamination of the aquifer by migration of CFCs released from industrial facilities. In spite of these potential concerns, Ekwurzel *et al.* (1994), in a field study on the mid-Atlantic coast, found that ages determined by the CFC method, tritium/helium-3 ratios, and the krypton-85 method, all agreed within about 2 years.

Krypton-85. Krypton-85, a radioactive noble gas with a half-life of 10.76 years, is produced in the atmosphere by the interaction of cosmic rays with krypton-84. However, nuclear weapons testing and reprocessing of nuclear fuel rods are by far the greater sources (Ekwurzel *et al.*, 1994). The atmospheric production of krypton-85 increased steadily since about 1950. After precipitation infiltrates and no longer contacts the atmosphere, the krypton undergoes decay, but owing to its inert characteristics, it does not interact chemically with the aquifer materials. Figure 3-9C illustrates the krypton activity in the northern atmosphere, and Figure 3-9D shows how to graphically determine the age of the water sample based on the sample collection date and the measured activity in groundwater.

Carbon-14. Carbon-14 (^{14}C) is a radioactive isotope of carbon produced in the upper atmosphere by cosmic ray interactions with nitrogen. The carbon-14 becomes part of the carbon dioxide molecule that dissolves in the water, enters the soil gas, and becomes part of the animal or plant tissue. When the water or soil gas are no longer free to exchange with the atmosphere, as when the animal or plant dies, the carbon-14 activity decreases at a rate controlled by its half-life, 5730 years. The concentration or activity ratio of carbon-14 to carbon-12, expressed as a percentage, is called the percent modern carbon (pmc). Except for input from thermonuclear testing in the 1950s and 1960s, this ratio has remained relatively constant in the atmosphere, varying by a factor of about two, over the last 100,000 years. Inasmuch as carbon-14 is detectable to about 1 pmc, the potential usefulness of the dating method in groundwater is about 20,000 years, based on Equation 3-29.

Unfortunately, there are a number of sources of uncertainty in carbon-14 dating. The most significant of these is due to the interactions of atmospheric carbon-14 with mineral carbon. Most of the carbon-14 in groundwater occurs in the bicarbonate ion. Minerals such as calcite and dolomite contain radiologically dead carbon that may be liberated along the groundwater flow path due to reactions such as the reduction of sulfate. Dilution by dead carbon makes the calculated age appear older than the true age, but methods to correct the carbon-14 activity for the geochemical effects have produced fairly reliable results. The principal difficulty seems to lie in the inability to reconstruct with confidence the geochemical and hydrological processes that have influenced the carbon-14 concentration in the sample. Another potential difficulty may arise if the sample age is post-1952, because the same nuclear weapons testing that produced tritium also generated carbon-14 concentrations of a few tens to about 200 pmc (Mazor, 1991). Therefore, care must be taken to use other methods, such as tritium, in combination with carbon-14 to assess whether the sample may be mixed with very young water. As with tritium, owing to the uncertainty in assessing the true age, carbon-14 is generally regarded as only a semiquantitative technique for recharge analysis.

Chlorine-36. Based on its long half-life, chlorine-36 (^{36}Cl) is potentially useful to date groundwater as old as about 2 million years. In many respects, chlorine-36 is an ideal tracer for dating old groundwater. Unlike carbon-14, chlorine-36 does not interact appreciably with most aquifers, although dead chloride can be dissolved from some salt-bearing natural formations to increase the apparent groundwater age. In clay-rich deposits the anion exclusion process may cause chloride ions to migrate slightly faster than the water. Chlorine-36 is readily detected in very small concentrations by a tandem accelerator/mass spectrometer. The ratio of chlorine-36 to stable chloride is used to determine the groundwater age from Equation 3-29. This method has been applied to date groundwater in Canada (Phillips *et al.*, 1986) and Australia (Bentley *et al.*, 1986).

MATHEMATICAL MODELS FOR ESTIMATING RECHARGE

Numerical models are best suited to predict recharge when the physical properties of the soil or groundwater are well characterized. Recharge is predicted as the output at the base of a model of the vadose zone, and it is calculated as the input to a calibrated groundwater flow model.

Soil-Water Models

Numerical models that are relevant to calculating deep percolation and recharge are water balance models and models based on Richards equation.

Water-Balance Models. The water balance models include codes such as HELP (Schroeder *et al.*, 1984), GLEAMS (Leonard *et al.*, 1989), and PRZM-2 (Mullins *et al.*, 1993). Additionally, Bauer and Vaccaro (1987) developed a soil-water balance model that has been applied to determine recharge for regional groundwater models that cover extensive areas of the Pacific Northwest (Bauer and Vaccaro, 1990) and Kansas (Hansen, 1991).

All these vadose-zone water-balance models partition precipitation into runoff and infiltration. Infiltration is further separated into components such as evapotranspiration, lateral drainage or interflow, soil-water storage, and deep percolation by applying deterministic and empirical equations that describe each of the processes. Potential and actual evapotranspiration are computed from historical regional climatic data (e.g., precipitation, temperature, solar radiation), from on-site weather measurements, or from default daily historical data for the nearest location stored in the program library. Other factors such as the vegetation cover and rooting characteristics also enter into the actual evapotranspiration analysis. Water that cannot be held in storage or extracted by the plants becomes available for deep percolation. Some models such as PRZM-2 take the deep percolation output from the water balance in the root zone and also route this through the deeper vadose zone using Richards equation for one-dimensional, unsaturated flow. For this routing, the water balance models typically require soil hydraulic conductivity, porosity, and moisture retention characteristics data or a limited set of soil characteristic parameters including field capacity, wilting point, saturated moisture content, and organic matter content.

The HELP model was developed by the U.S. Army Corps of Engineers to support landfill design. In this context, the HELP model simulates water movement across, into, through, and out of landfills using input data on weather at the site, landfill dimensions, and soil properties. However, in order to apply the HELP model to the determination of natural recharge, landfill materials could be replaced by native soils. Whether used to predict landfill seepage or natural recharge, deep percolation through the vadose zone is simulated according to the Darcy flux method using empirically derived values for unsaturated hydraulic conductivity estimated from a Brooks-Corey (1964) based relationship (Equation 3-5). This approach uses a limited set of soil properties, including total porosity and field capacity, and does not require laboratory characterization of the entire moisture retention characteristic curves.

Several of the water balance models (e.g., HELP and PRZM-2) incorporate the concept of field capacity (Figure 3-4). Field capacity for modeling purposes is typically defined as the water content the soil can hold against gravity (e.g., Phillips *et al.*, 1993). Percolation below any soil layer is allowed in the models only if the water content exceeds the field capacity. That is, if the water content is less than field capacity, then it is assumed that either the plants will consume all the water or the water will remain in the layer and no deep percolation can occur.

There is significant research, however, to demonstrate that water percolation can occur under gravity at water contents much less than field capacity (Stephens, 1985, 1994). Thus, at poorly vegetated sites where water content is almost always less than field capacity, recharge should not occur, but Stephens and Knowlton (1986) found that this is certainly not the case at a semiarid site in New Mexico. At an instrumented plot in a chalk deposit in England, Wellings (1984) concluded that the soil-water profile continuously changed over time, and no evidence supported the validity of the field capacity and available water concepts (Figure 3-4) as they relate to recharge. If recharge rates are low and the period of water balance accounting is too long, then water balance models are likely to underestimate recharge because of the approximate manner in which deep percolation is calculated.

In spite of the concern about the use of field capacity, the HELP model has recently been used to successfully predict deep percolation and recharge beneath a proposed landfill in southern New Mexico (Stephens and Coons, 1994). The simulated percolation rate compared very favorably with independent estimates of recharge at the site using the chloride mass balance method and a Darcy flux approach which used laboratory-determined values for unsaturated hydraulic conductivity following the van Genuchten relationship (Equation 3-6). The close agreement between the three approaches reported by Stephens and Coons (1994) lends credibility to the use of the HELP model to estimate recharge, and further supports the use of numerical models as a means of developing recharge estimates.

The water balance model developed by Eagleson (1979) improves significantly in the application of unsaturated flow physics and provides a probabilistic approach to recharge (Milly, 1994). The model considers uncertainty in storm characteristics, soil physical properties and one-dimensional vertical flow, evapotranspiration via Penman's equation, and groundwater flow out of the basin. Cumulative probability density functions for recharge by this method have been developed for a sub-humid site in Clinton, Massachusetts and for an arid site in Santa Paula, California (Eagleson, 1978). These functions provide the return period or probability of occurrence of recharge events of a particular magnitude.

Numerical Models Based on the Richards Equation. The Richards equation (Richards, 1931) is the governing equation for numerical models of unsaturated flow within the vadose zone:

$$\frac{\partial}{\partial z}K(\psi)\frac{\partial h}{\partial z} = C(\psi)\frac{\partial \psi}{\partial t} \qquad \text{(Equation 3-30)}$$

where $K(\psi)$ = Unsaturated hydraulic conductivity (LT^{-1})

ψ = Pressure head [L]

h = Total head (L)

$C(\psi)$ = Specific moisture capacity = $\frac{d\theta}{d\psi}\ L^{-1}$

z = Vertical coordinate

t = Time

The left-hand side of Equation 3-30 represents the net inflow into a fixed volume of the soil, and the right-hand side represents the net change in water storage in the volume. For heterogenous systems, material properties are specified for each layer. Richards equation is applicable to porous-media flow only and is not suitable for simulations of macropore and fracture flow, unless these features are so numerous at the scale of the model as to render the system equivalent to a homogeneous porous medium.

There are a large number of numerical models for simulating soil-water processes, including finite difference and finite element forms, based on one-, two-, or three-dimensional forms of Richards equation (Table 3-3). There are many other codes that include both flow and transport, such as VAM2D (Huyakorn *et al.*, 1989) and TRACR3D (Travis, 1984). To account for infiltration and evapotranspiration in these codes, in lieu of detailed meteorological information, the upper boundary of the model and/or the root zone is usually specified as a constant or time-varying flux or pressure head. In contrast to the water balance models, the numerical models allow the user to more realistically represent the physical properties of the porous medium, including complex geology with spatially varying hydraulic conductivity and water retention characteristics. When the lower boundary of the model is specified as the water table, the water flux out of the base of the model represents the groundwater recharge.

Large-scale applications of Richards equation-based numerical models to highly heterogeneous soils with variable hydraulic properties and flow characteristics can be extremely difficult and expensive. Such applications require complex discretization of the model domain and intensive determination of the hydraulic properties to be used as input variables. Computer simulations for these situations are prone to problems and require an enormous amount of CPU time. Because of this, Richards equation-based numerical models are generally applied to reasonably

Table 3-3. Available codes for single-phase (water) flow in the vadose zone

Code Name	Dimensions	Method	Reference	Code Type
2DSEEP	2	FEM	OECD, 1990	Unknown
3DSEEP	3	FEM	OECD, 1990	Unknown
AMOCO	3	FDM	Odeh, 1981	Proprietary
ANGEL	3	FEM	OECD, 1990	Unknown
BETA-II	3	FDM	Odeh, 1981	Proprietary
BRUTSAERT1	2	FDM	Oster, 1982	Public Domain
BRUTSAERT2	2	FDM	Brutsaert, 1971	Public Domain
CMG	3	FDM	Odeh, 1981	Proprietary
DELAAT	2	FEM	Oster, 1982	Unknown
FEMWATER	2	FEM	Yeh, 1987	Public Domain
FLUMP	2	FEM	Neuman and Witherspoon, 1971	Public Domain
GANDALF	2	FDM	Morrison, 1977	Public Domain
GPSIM	3	FDM	Odeh, 1981	Proprietary
GWHRT	3	FEM	Carlsson et al., 1983	Unknown
MOMOLS	1	FDM	Rojstocyer, 1981	Public Domain
PORES	3	FDM	Oster, 1982	Unknown
REEVES-DUGUID	2	FEM	Reeves and Duguid, 1975	Public Domain
SHELL	3	FDM	Odeh, 1981	Proprietary
SSC	3	FDM	Odeh, 1981	Proprietary
STGWT/ MOG WT	1	FDM	De Smedt and van Beker, 1974	Unknown
SUM2	2	FEM	Oster, 1982	Unknown
SUPERMOCK	2	FDM	Reed et al., 1976	Public Domain
TRIPM	2	FEM	Gureghian, 1981	Public Domain
TRUST/TNN	3	IFDM	Narasimhan, 1990	Public Domain
TS&E	3	Unknown	Oster, 1982	Proprietary
UNFLOW	2	FEM	Oster, 1982	Public Domain
UNSAT1	1	FEM	van Genuchten, 1978b	Public Domain
UNSAT1D	1	FDM	Oster, 1982	Public Domain
UNSAT2	2	FEM	Neuman et al., 1974	Unknown
UNSAT-H	1	FDM	Fayer and Jones, 1990	Public Domain
VERGE	3	FEM	Verge, 1976	Public Domain

FDM = finite difference method
FEM = finite element method
IFDM = integrated finite difference method

simple scenarios with spatially averaged values for the soil hydraulic properties and input parameters, often using overly-simplistic averaging schemes, which decrease the reliability of the model results.

A number of theoretical improvements and modeling techniques have been developed during the past several years which circumvent these limitations. The most notable among these are the stochastic methods of Yeh *et al.* (1985a,b,c), Mantoglou and Gelhar (1987a,b,c) and Mantoglou (1992). Yeh, Gelhar and Gutjahr (1985a,b,c) first proposed a stochastic approach for incorporating "effective" hydraulic properties and input parameters within numerical models of steady unsaturated flow. Their work was expanded by Mantoglou and Gelhar (1987a,b,c) to include non-steady flow and large-scale hysteresis of soil hydraulic properties. Although these approaches were based on sound mathematical derivations and received widespread acceptance within the scientific community, they both required smoothly variable soil properties and flow characteristics in space and in time. Further development by Mantoglou (1992) removed these restrictions, thereby greatly extending the utility of stochastic modeling to include most realistic field situations.

Subsequent model development and testing by Jensen and Mantoglou (1992) confirmed this theoretical approach. The results of a comparison of field data with the output of the stochastic modeling conducted by Jensen and Mantoglou (1992) shows that their stochastic approach predicts average system behavior better than models which use other schemes (e.g., geometric means) and provides a ". . . rational framework for modeling large-scale, unsaturated flow and estimating areal averages of hydrological processes in spatially variable soils."

Groundwater Models

Groundwater flow models can also predict recharge when other hydrologic information is known or assumed. In finite difference numerical models, for example, a grid of cells is laid over the domain of interest, and representative values of aquifer transmissivity and storage are assigned to the aquifer materials. The concept here is to calibrate the model to observed water levels (hydraulic heads) by adjusting the model parameters until model-predicted and observed heads reach suitable agreement. The flux input to the water table necessary to produce this agreement is then assumed to represent recharge. Examples of modeling to compute recharge include the two-dimensional groundwater flow model of the Ogallala aquifer in part of New Mexico (McAda, 1984) and the three-dimensional groundwater flow model at Wright-Patterson Air Force Base in southwestern Ohio (Dumouchelle *et al.*, 1993), among many dozens of others.

Inasmuch as there is always some uncertainty associated with aquifer properties and hydraulic heads, there is also uncertainty in the recharge estimate. The errors in model input parameters, which are frequently large, will be accumulated in a back-calibrated recharge estimate. Additionally, calibration by trial and error is quite tedious and does not readily permit quantification of the uncertainty in the recharge. However, automatic procedures, such as MODFLOWP (Hill, 1992), are now available to facilitate calibration and parameter estimation.

Numerous groundwater model studies have focused on recharge, for example:

- Theis (1937) applied a steady-state analytical solution for a sloping water table aquifer to predict recharge to the Ogallala aquifer by adjusting the recharge in the vertical slice model until a reasonable match was achieved between the shapes of the predicted water table and the observed water table.

- Su (1994) developed analytical solutions for certain boundary conditions to calculate transient recharge from precipitation, water-level hydrographs, and soil and aquifer properties; the analysis takes into account the sloping water table and groundwater drainage.

- Lohman (1971) presented a graphical method to calculate diffuse areal recharge where monitor wells were located in a five-spot pattern, much like a finite difference cell.

- Based on the one-dimensional groundwater flow equation applied to water table fluctuations for a project on Long Island, Jacob (1943) developed a predictive relationship between a weighted average annual precipitation and recharge.

- Simpson and Duckstein (1976) developed a discrete-state compartment, or mixing cell, model which partitions the aquifer into cells, within which conservation of mass is applied. Tracers are used to calibrate the model, including carbon-14 (Campana and Simpson, 1984), tritium (Campana and Mahin, 1985), and deuterium (Kirk and Campana, 1990).

- Where some information exists on recharge and its variability at certain locations within a groundwater basin, a stochastic inverse method may be useful to provide optimal estimates of the recharge throughout the entire basin (Graham and Tankersley, 1994). Graham and Neff (1994), who applied this analytical approach in conjunction with groundwater flow modeling of the Upper Floridian Aquifer in northeast Florida, indicated that some of the assumptions in the analytical method may need modification before the method is widely applicable.

Section 4

Recharge Estimates

Section 4
RECHARGE ESTIMATES

Because the process of infiltration is subject to complex interrelationships between climate, vegetation, topography, surface soils, and vadose zone materials, estimates of diffuse groundwater recharge vary significantly from region to region across the many different environmental regimes of the United States. Estimates of diffuse recharge for similar climatic regimes may also vary significantly because of variability in site-specific physical conditions and the estimation technique used.

A compilation of estimates of diffuse annual recharge for 18 separate geographic regions of the United States (Figure A-1) is presented in Appendix A. The data were compiled to (1) identify key studies and sources of information on recharge estimates throughout the U.S., (2) determine the frequency with which techniques are applied in various hydrogeological settings or climates, and (3) develop a database for future statistical analysis. The estimates, based on the methodologies discussed in Section 3, are presented by geographic region in Table A-1 and by estimation technique in Table A-2.

In Appendix A, recharge is expressed as a percentage of mean annual precipitation. Such an expression is often convenient and within a particular region may provide an indication of the relative importance of recharge in the overall water balance. However, when recharge events are infrequent, as they are in dry climates, all the annual recharge could occur within a period of only days or weeks. The actual recharge rate during the brief period of recharge would be significantly greater than the annual recharge rate. On the other hand, many storms occur during the year that do not lead to recharge. Consequently, expressing recharge as a percentage of precipitation may be inappropriate, especially in areas of low precipitation. Allison *et al.* (1994) suggested that an empirical relationship between precipitation and recharge would be most useful where mean annual recharge exceeds about 50 mm.

In the following subsections, the various estimation techniques commonly applied in recharge studies are compared on the basis of site characteristics, accuracy, cost, and other, method-specific criteria.

COMPARISON OF METHODS FOR ESTIMATION OF DIFFUSE RECHARGE

Of the numerous techniques described in Section 3 and compiled in Table A-2, most yield results that are problem- and scale-dependent. For example, for determining site-specific recharge estimates, techniques that rely on very local measurements are most appropriate. Such techniques include lysimeters, chemical tracers, the Darcy flux and plane of zero flux methods, one-dimensional soil-water balances and soil-water models, and soil temperature methods based on near-surface soil temperature gradients. These techniques will probably not provide data representative of other similar geographic regions, or even of other sites in the same basin or watershed, because of the spatial and temporal variability of factors controlling the recharge process. Conversely, other techniques based on ground-water models, regional water balances, aquifer temperature profiles, water-level fluctuations, streamflow data, and basin outflow calculations provide recharge estimates that are averaged over regional aquifers or basins and therefore may not be fully representative of a specific site within the study region. Additionally, because the recharge process differs between arid and humid environments (see Section 2), some techniques are best suited for arid or semiarid environments (for example, chemical methods in the vadose zone), whereas others are appropriate only in humid environments (for example, streamflow data).

Because of the variety of estimation techniques and conditions under which they are optimally applied, we have attempted a comparison of the various methods based on site conditions, relative cost and accuracy, and other method-specific criteria. This method comparison is summarized in Table 4-1 and discussed in more detail in the following subsections.

Soil-Water Balance

The techniques most widely used to estimate diffuse recharge are the soil-water balance method and the use of streamflow data (discussed in the following subsection). The soil-water balance method has been applied in a variety of climates, from the arid and semiarid areas of the western United States to the more humid regions of New England, the mid-Atlantic, and the Texas Gulf Coast. The approach is used for studies varying in scale from land plots and small watersheds to regional aquifers covering tens or hundreds of thousands of square kilometers.

From the compilation presented in Table A-2, one can see that recharge estimates derived by applying the soil-water balance method over a large study area vary significantly, sometimes up to two orders of magnitude. The technique is both sensitive to periods of accounting and subject to errors in averaging measurements over large time and space domains. The accuracy of the method depends on the accuracy of the component parameters, which are sometimes poorly known or exhibit significant geographic and temporal variability.

The greatest uncertainty lies in estimating evapotranspiration, but other major components of the water balance (precipitation, streamflow, runoff, changes in soil-water storage) can also introduce substantial error and uncertainty. As Gee and Hillel (1988) point out, because precipitation measurements are rarely more accurate than ±5% and evapotranspiration can be measured at best within ±10%,

Table 4-1. Comparison of Methods for Estimation of Diffuse Recharge
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Estimation Technique	Data Requirements	Optimal Site Characteristics	Relative Accuracy[1]	Relative Cost[2]	Comments
Soil-water balance	• Precipitation (P) • Surface runoff • Evaporation (E) • Transpiration (T) • Storage (S)	Humid/temperate climate (P≥ET); flat topography with measurable or negligible runoff; short, uniform vegetation; small scale	Low	High/low	Commonly used technique, not appropriate for arid climates where ET>>P; uncertainty varies by a factor of 3 to 10 or more. High cost if micrometeorological equipment pre-purchased; low if ET is calculated from PET.
Lysimetry	• Water volume	Applied under any site conditions; construction results in devegetation	High	High	Direct, precise measurement of deep drainage; precision ±1 mm/yr; long-term monitoring and maintenance required; when combined with soil-water balance, is very reliable for arid site
Darcy flux	• Hydraulic gradient • Unsaturated hydraulic conductivity	Applied under any site conditions	Low to moderate	Low to moderate	Results rely on measurement of unsaturated hydraulic conductivity; accurate with ± a factor of 10 or more
Plane of zero flux	• Soil water potential profile • Water content changes with time	Temperate, semiarid, or arid climates (ET>P); any soil type	Moderate	Moderate to high	Accuracy ±15% or ~20 mm/yr; requires weekly monitoring; fails during periods when rainfall exceeds K_{sat}
Soil temperature gradient [Bredehoeft and Papadopulos (1965) type-curve method]	• Steady-state temperature profile data from saturated zone	Deep aquifers with upward temperature gradient	Low	Low	Provides regionally averaged recharge estimate with accuracy similar to basin water balance

[1] Relative accuracy is a qualitative comparison of the accuracy of the method when applied under optimal site conditions and, like the relative cost tabulation, is based on the authors' professional judgment and experience.

[2] Relative cost is based on the authors' past experience and the following criteria: low-cost: less than $10,000; moderate cost: $10,000 to $50,000; high cost: greater than $50,000.

Table 4-1. Comparison of Methods for Estimation of Diffuse Recharge
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Estimation Technique	Data Requirements	Optimal Site Characteristics	Relative Accuracy[1]	Relative Cost[2]	Comments
Electromagnetic resistivity	• Electrical conductivity measurements • Independent recharge estimate for comparison	Fine-grained soils; varied vegetation and climate	Non-quantitative	Low to moderate	Provides reconnaissance-level, qualitative results that identify areas of recharge
Basin outflow	• Aquifer transmissivity • Aquifer hydraulic gradient basin boundaries • Upstream watershed surface area • Specific yield • Transient hydraulic head change	Any unconfined aquifer with a well-characterized flow regime, and well-defined recharge areas	Low	Low to high	Provides regionally averaged recharge estimate with accuracy similar to basin water balance; can often rely on existing data. Low cost provided data already exist; high cost if data collection required.
Water-level fluctuations	• Water table hydrograph • Specific yield	Any unconfined aquifer with a well-characterized flow regime, and well-defined recharge areas	Low	Low	Provides regionally averaged recharge estimate with accuracy similar to basin water balance; can often rely on existing data
Stream gauging	• Streamflow hydrograph	Humid/temperate climate; well-developed, unmanaged watershed with perennial streams; stream-connected shallow aquifer; minimal snowmelt	Moderate	Low	Avoids need to measure climatic parameters; provides regionally averaged recharge estimate for watershed with better accuracy than basin water balance

[1] Relative accuracy is a qualitative comparison of the accuracy of the method when applied under optimal site conditions and, like the relative cost tabulation, is based on the authors' professional judgment and experience.

[2] Relative cost is based on the authors' past experience and the following criteria: low-cost: less than \$10,000; moderate cost: \$10,000 to \$50,000; high cost: greater than \$50,000.

Table 4-1. Comparison of Methods for Estimation of Diffuse Recharge
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Estimation Technique	Data Requirements	Optimal Site Characteristics	Relative Accuracy[1]	Relative Cost[2]	Comments
Tritium profile	• Undisturbed soil profile below the root zone • ^{3}H input function • ^{3}H concentrations in soil moisture	Arid, semiarid and temperate climates where R>10% MAP; sediments of any texture and pedogenic carbonates	High	Low	Very accurate water tracer; conceptual model assumes piston flow (i.e., ignores preferential flow); subject to vapor transport, which causes overestimate of R where R ≤ 1 mm/yr
Chlorine-36 profile	• Undisturbed soil profile below the root zone • ^{36}Cl input function • ^{36}Cl concentrations in soil moisture	Arid, semiarid and temperate climates where R>10% MAP; sediments of any texture and pedogenic carbonates	High	Low to moderate	Fairly accurate water tracer, subject to anion exclusion and ultrafiltration; conceptual model assumes piston flow (i.e., ignores preferential flow); high cost of analysis ($500-$1000/sample)
Chloride mass balance	• Undisturbed soil profile • Meteoric chloride concentration • Chloride concentration in soil moisture • Mean annual precipitation	Arid, semiarid and temperate climates where R>10% MAP; sediments of any texture and pedogenic carbonates	High	Low	Conceptual model assumes (1) average rate of chloride deposition in precipitation is constant, and (2) piston flow; very inexpensive
Stable isotope profile	• Undisturbed soil profile • Water content profile • D and ^{18}O concentrations in soil moisture	Arid and semiarid climates where soil-water movement is in quasi-steady-state; sediments of any texture	Unknown	Low to moderate	Conceptual model assumes one-dimensional, vertical, quasi-steady-state soil-water movement; non-routine soil-water extraction process; requires further research to evaluate uncertainty

[1] Relative accuracy is a qualitative comparison of the accuracy of the method when applied under optimal site conditions and, like the relative cost tabulation, is based on the authors' professional judgment and experience.

[2] Relative cost is based on the authors' past experience and the following criteria: low-cost: less than $10,000; moderate cost: $10,000 to $50,000; high cost: greater than $50,000.

Table 4-1. Comparison of Methods for Estimation of Diffuse Recharge
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Estimation Technique	Data Requirements	Optimal Site Characteristics	Relative Accuracy[1]	Relative Cost[2]	Comments
Groundwater age dating	• Hydraulic gradient • Effective porosity • Distance to tracer peak • Apparent groundwater tracer age	Shallow, unconfined aquifer; vertical hydraulic gradients near the water table; applicable to any climate, soil texture and vegetation	High	Low	Groundwater age best determined by $^{3}H/^{3}He$, ^{36}Cl, and/or CFCs; requires thorough understanding of aquifer flow system and careful application; very consistent results. High relative accuracy if source of ^{3}H is known.
Soil-water balance models	• Precipitation • Surface runoff • Evapotranspiration • Soil-water storage • Soil hydraulic properties	Applicable to any conditions and any scale where vertical flow occurs	Low to moderate	Moderate to high	Relies on estimates of AET and unsaturated hydraulic conductivity; uncertainty varies by an order of magnitude or more
Soil-water models based on Richard's equation	• Climatic data • Soil hydraulic properties • Depth to water table • In situ pressure head or water contents	Homogeneous soil profiles above a shallow water table; moist soils	Low to moderate	Moderate to high	Uncertainty is due to climatic data and hydraulic properties. Extensive computational effort for deep water tables, dry heterogeneous soil.
Groundwater models	• Aquifer geometry • Transmissivity, storage • Aquifer boundary conditions • Initial head field	Applicable to any conditions and any scale	Moderate	Moderate to high	Cost can be considerable if data are not compiled. Requires thorough calibration.

[1] Relative accuracy is a qualitative comparison of the accuracy of the method when applied under optimal site conditions and, like the relative cost tabulation, is based on the authors' professional judgment and experience.

[2] Relative cost is based on the authors' past experience and the following criteria: low-cost: less than \$10,000; moderate cost: \$10,000 to \$50,000; high cost: greater than \$50,000.

recharge estimates based on precipitation and evapotranspiration can vary by a factor of three or more. Where calculation of the infiltration component of a water balance relies on the measurement of unsaturated hydraulic conductivity, errors of a factor of ten or more can be expected. The water balance method is most applicable to humid or temperate regions and on a small scale, as for a point release, a land plot, or a small watershed, where input parameters that rely on climatic data have low variability.

Stream Gauging

The estimation technique that uses streamflow data is widely applied in humid environments with stream-connected, shallow aquifers. The method relies on the basic concepts of a basin-wide water balance and recognizes that the portion of precipitation that infiltrates past the root zone and recharges the aquifer is then available as baseflow to streams. This method is not appropriate for arid or semiarid environments where ephemeral or losing streams dominate, but is practical for application elsewhere. Some investigations (e.g., the Tennessee River Basin [Hoos, 1990]) have applied a combined hydrologic balance and streamflow study wherein hydrographs of stream discharge were used to solve the water balance equation under the assumption that the aquifer was in steady state so that discharge approximately equaled recharge. Where the steady-state assumption is correct, this approach avoids the need to directly measure climatic parameters or soil hydraulic properties and thus offers a simple, accurate alternative to the conventional water balance.

Darcy Flux and Plane of Zero Flux Methods

Other physical methods commonly applied to the estimation of recharge, such as the Darcy flux and the plane of zero flux methods, also have limitations. The Darcy flux method relies on estimates of unsaturated hydraulic conductivity, which can vary by up to six orders of magnitude over the seasonal range of moisture contents typically found in near-surface soils. It is difficult and time consuming to determine hydraulic conductivity as a function of moisture content and pressure head, and the difficulty and uncertainty increase with soil dryness. Measurements of unsaturated hydraulic conductivity, and hence estimates of recharge based on that parameter, are only considered accurate within an order of magnitude at best. Other significant sources of uncertainty include hysteresis of the unsaturated hydraulic conductivity-pressure head relationship, spatial variability of conductivity and moisture content, and the occurrence of preferential flow.

The work of Stephens and Knowlton (1986) at an arid site in New Mexico demonstrates the problems with the Darcy flux method by showing variations of more than two orders of magnitude in recharge. In addition, this study found that the annual recharge rate varied by more than a factor of five depending on how the mean unsaturated hydraulic conductivity was calculated. When sufficient number of measurements can be collected over a range of site moisture conditions (including measurements during different seasons), this method is considered more accurate than water balance techniques. However, errors of at least an order of magnitude or more should be expected.

The plane of zero flux (or zero flux plane) method, which relies on distinguishing the evapotranspiration and percolation components of total soil-water flux across a zero flux plane, offers an alternative to the Darcy flux method. The zero flux plane can be accurately identified with inexpensive tensiometers, and water content changes below the zero flux plane, which constitute recharge, are also easily obtainable with neutron logging. The overall accuracy of recharge estimates computed with this method depends on the frequency of water content measurements and the spatial variability of soil properties.

A number of case studies in the United Kingdom have demonstrated that the zero flux plane technique is a robust approach with a precision of about 20 mm/yr. (Gardner *et al.*, 1991). Dreiss and Anderson (1985) combined the zero flux plane and water balance methods using weekly measurements at a land-plot scale to obtain recharge estimates accurate within 15 percent or less. The main problem with the approach is that it fails during extended periods of rainfall when soils are close to saturation and the zero flux plane is not detectable. In such cases, Darcy flux or soil-water balance methods, although accompanied by greater sources of uncertainty and error, will prove more useful.

Soil-Water and Groundwater Models

A number of large-scale studies of diffuse natural recharge have relied on soil-water models or groundwater models. Modeling tools are an extremely valuable means of evaluating recharge (as well as other flow characteristics) and can be applied at sites under any environmental or climatic conditions or on any scale. It is important to keep in mind that errors in the model input parameters, which are frequently significant, will be accentuated in the back-calibrated recharge estimate. For example, most soil-water models rely directly or indirectly on estimates of hydraulic conductivity and actual evapotranspiration, which as noted above can introduce errors of an order of magnitude or more.

Lysimetry

Whereas lysimetry provides the only means of directly measuring recharge, the accuracy of the technique depends on the validity of two important underlying assumptions: (1) that the soil materials and conditions inside the lysimeter are representative of *in situ* conditions outside the instrument, and (2) that all moisture entering and stored in the lysimeter is accurately measured and accounted for. Because of the nature of lysimeter construction, the first assumption is never completely valid, and great care must be taken to ensure the validity of the second assumption.

Relative to other techniques, lysimeter measurements require complex field instrumentation and long-term monitoring and maintenance at high cost for both materials and labor. In humid environments, seasonal monitoring may be adequate, but years of data are required in arid climates in order to achieve an accurate estimate of average recharge. Additionally, the method provides very localized estimates of recharge and cannot account for spatial variability. Because of the practical time and cost limitations, the technique is not commonly used, but when carefully applied, lysimetry provides estimates of recharge with a precision as low as 1 mm/yr. (Gee and Hillel, 1988).

Other Physical Techniques

Other techniques relying on basin outflow analyses, soil temperature gradients, or water-level fluctuations provide regionally averaged estimates of diffuse recharge. The basin outflow method has the potential to work well if the aquifer boundaries and recharge areas are well defined and transmissivity estimates are accurate. Under these conditions, this method has been shown to produce recharge estimates comparable to those obtained by soil-water balance and groundwater modeling methods. The Bredehoeft and Papadopulos (1965) type-curve method is the only technique relying on soil temperature gradient data that has been shown to provide reasonable quantitative estimates of a recharge flux. Where the method has been applied to deep aquifers with an upward temperature gradient, results have also been shown to compare well with recharge estimates obtained from basin water balance calculations. However, as mentioned above, these methods provide a regionally averaged recharge estimate, and the comparative agreement with results from water balance studies reflects a relatively low degree of accuracy compared with other site-specific techniques. The estimation method relying on *water-level fluctuations* can also provide reasonable, regionally averaged recharge values, and when combined with other site-specific point measurements, as in the work by Sophocleous and Perry (1985), water-level fluctuations provide a clear, quantitative depiction of the recharge process.

Chemical Methods

Chemical tracer techniques in the vadose zone are most appropriately applied in arid or semiarid environments and can provide quantitative, reproducible, and consistent estimates of recharge (e.g., Phillips *et al.*, 1988; Scanlon, 1992). The use of environmental tracers avoids many of the limitations and uncertainties inherent in measuring physical parameters. Tracers allow direct measurement of water and solute displacement in both the vadose zone and groundwater. Of the variety of chemical methods available for estimating recharge, the most widely applied techniques in the vadose zone are chloride mass balance and the bomb tracers, chlorine-36 and tritium.

The conceptual model for movement of environmental tracers through the vadose zone assumes piston displacement (thereby ignoring preferential flow and dispersion). Furthermore, the application of chloride mass balance assumes a constant input of meteoric chloride over time. Chloride concentration in precipitation is the parameter with the greatest uncertainty in calculating a recharge flux to groundwater when using the chloride mass balance technique. In assessing the uncertainty in recharge estimates to the High Plains aquifer of Texas and New Mexico, Wood and Sanford (1995) found that for every 0.1-mg/L difference in weighted average chloride in precipitation, the calculated recharge flux was changed by about 1.8 mm/yr.

Aquifer Hydrochemistry

Hydrochemistry studies of aquifers, which rely on the same chemical signatures as those used in vadose zone and other techniques, can also provide valuable information concerning groundwater ages, sources of recharge, and aquifer flow characteristics (e.g., Panno *et al.*, 1994). However, these types of studies often tend to provide more semiquantitative results and, if intended to provide quantitative recharge estimates, require careful application with a thorough understanding of the aquifer flow system. Although it has not received wide application due to its relatively recent development, the technique involving co-measurement of tritium and helium-3 is probably the most precise and definitive chemical method available for determining groundwater ages and, hence, recharge (e.g., Ekwurzel *et al.*, 1994; Solomon *et al.*, 1993).

Section 5

Considerations and Recommendations for Selecting Recharge Estimation Techniques

Section 5
CONSIDERATIONS AND RECOMMENDATIONS FOR SELECTING RECHARGE ESTIMATION TECHNIQUES

In the previous sections, we have reviewed some of the concepts relevant to groundwater recharge, identified methods to quantify it, and summarized some of the applications of recharge analyses throughout the United States. This section highlights things to consider when attempting to estimate site-specific recharge rates and provides some basic recommendations for both humid and arid climates.

CONSIDERATIONS

When assessing the risk associated with contaminant migration in the vadose zone, one should seek methods to assess the recharge that is likely to occur in the future. The ideal approach would include methods that integrate over the complete range of recharge events that have occurred in the past, as these are considered a good indicator of mean recharge in the future. The usual alternative is to establish field instrumentation at the site of interest and apply methods that rely on the spatially distributed data or time series that are generated; however, time and budget constraints usually do not permit such extensive data collection efforts.

There are no universally acceptable methods to compute diffuse recharge that can be applied to all sites. The method selected will depend on the site geology, soil characteristics, depth to the water table, vegetation, and climatic conditions, along with other factors such as time constraints, available budget, and the importance of recharge to the success of the particular project. The comparison of methods presented in Table 4-1 should provide a guide to the appropriate selection of an estimation method based on optimal site characteristics, cost, and accuracy.

Regardless of the method used, however, it is important to recognize the *uncertainty in the estimate.*

Attempt to understand method uncertainty. The current state of the art is that, except for lysimeter methods, all recharge methods are indirect techniques. Therefore, one must recognize the uncertainty in the data required to compute recharge and the degree to which this uncertainty translates into uncertainty in the calculated recharge.

There can be many sources of uncertainty, including uncertainty in the conceptual model and inherent assumptions that form the basis for each of the recharge methods and uncertainty in the parameters for each method. Because the sources of uncertainty in each method differ, the results of the various methods are expected to differ as well.

At present, there is no comprehensive analysis of uncertainty associated with each of the recharge methods as they are applied to dry and humid climates. A site-specific uncertainty analysis is recommended, when practical, to facilitate selecting the appropriate recharge technique. Such analysis should quantify the precision and accuracy that may be expected for each method applied under specific site physical conditions. One approach to evaluate uncertainty is to apply different techniques, including physical and chemical methods, to quantify recharge and to then evaluate the range of values obtained by the different methods. Another approach is to estimate the uncertainty associated with each parameter used in the recharge calculation. Section 3 of this report provides both general and specific information regarding parameter uncertainty where such information is available. However, this information should be used only as a decision-making guide and not as a substitute for a site- and method-specific uncertainty analysis.

When time and budget are limited, one may find the local and regional hydrogeological publications useful for estimating recharge. Excellent resource materials are available at district offices of the U.S. Geological Survey and state water or geological survey offices. From these sources one can compile recharge estimates from previous investigations and evaluate their relevance to a particular location. One should pay special attention to similarity of precipitation ratios, soil type, and topographic features, as well as to vegetation type and density. Where a range of possible appropriate recharge estimates are available in the literature, use the maximum and minimum of those recharge values in the risk assessment calculations.

Where site-specific measurements are required but time and resources are still limited, one may consider an approach based on a one-time data collection. The Darcy flux analysis based on laboratory tests of core samples or field tests of hydraulic properties deep in the vadose zone is a reasonable approach.

Where a unit gradient can be assumed (i.e., after prolonged periods of redistribution or below the depth where evapotranspiration is significant), the Darcy flux analysis requires an estimate of unsaturated hydraulic conductivity at the *in situ* moisture content. Moisture content can be measured economically in the field by neutron logging, and unsaturated hydraulic conductivity can be calculated using the Brooks-Corey (1964) or van Genuchten (1978a) relationships (Section 3) from field or core measurements of the moisture retention characteristic curves. Again, order of magnitude uncertainty in the result may occur even under optimum (i.e., humid climate) conditions.

Where adequate resources are available, it intuitively makes good sense to use both the chemical and physical methods for a specific site. Finally, if time and resources are plentiful, lysimetry can be employed for direct measurement of recharge.

RECOMMENDATIONS

Based on the discussion presented in this report, we offer some recommendations for the most effective and appropriate methods for humid and arid climates. These recommendations provide a general overview of the approaches that should be considered for each setting. However, the specific physical, chemical, and climatic conditions at each site and the data needs and objectives of each program must also be taken into consideration when designing a site- and program-specific recharge study. Accordingly, these recommendations may serve only as a guide to the available methods in order to support the selection of the most suitable approach for a given site and program.

Estimation Methods for Humid Climates

In humid climates where the water table is relatively shallow and much of the groundwater recharge occurs by percolation of precipitation, the mean recharge may best be derived from data collected in the upper portions of the aquifer. Abundant recharge in relatively humid climates often produces significant water-table fluctuations that facilitate application of a variety of physical methods to quantify recharge using well hydrograph data. Chemical tracer techniques may offer the best approach to obtain a time- and space-integrated estimate of recharge. However, these techniques have not been widely used on site-specific risk assessments. For areas where recharge is rapid and the groundwater age is less than about 50 years, the tritium/helium-3 method may be most appropriate.

Estimation Methods for Arid Climates

In areas of very low precipitation and scant, infrequent recharge, the long-term mean recharge is best obtained from chemical information in the vadose zone. This conclusion is especially true for deep water table conditions. The chloride mass balance or chlorine-36 methods applied to soil samples are potentially useful tracers of soil-water movement that potentially becomes recharge. Where wells are available for collecting groundwater samples and where recharge takes place over thousands of years, recharge analysis by carbon-14 and chlorine-36 methods are good choices. At some sites, such as near ephemeral streams and arroyos, rare recharge events produce significant water-table responses that can also be applied to recharge analyses.

Section 6

Future Research

Section 6
FUTURE RESEARCH

In spite of considerable research over the past 50 years into developing methods to quantify recharge, a great deal of work remains to be completed regarding the application of recharge data to contaminant transport problems. For these problems, the contaminant transport rate is strongly dependent upon the recharge rate, that is, the specific discharge below the root zone. In many soils, deep soil-water movement is enhanced where precipitation at the surface infiltrates rapidly through the zone of evapotranspiration via macropores such as root holes, rodent and insect burrows, and fractures. In cases where surface water ponds above a vertically continuous macropore network, rapid contaminant migration would occur compared to the migration rate within the adjacent porous media. Research is needed to develop methods to determine whether macropore transport will occur at a specific site and, if it does, over what portions of the site matrix and macropore flow will the transport occur.

Macropores contribute to one of the sources of spatial variability in recharge; however, there are others. For example, unstable flow, which may occur in some layered or hydrophobic soils, leads to irregular wetting fronts and preferential recharge through only a portion of a site. Research is needed to develop methods for evaluating site conditions conducive to producing unstable flow. Additionally, where preferential flow occurs, tools are needed to quantify its significance as it relates to contaminant migration.

A better understanding of uncertainty in the estimates of recharge between different methods and in various hydrogeologic and climatic settings is also needed. As mentioned in Section 5, no comprehensive analysis of uncertainty associated with each of the recharge methods exists as they are applied to dry and humid climates. *Additional field* tests of the low cost, simpler estimation techniques compared against more rigorous measurement systems, under a variety of conditions, would allow quantification of the uncertainty in recharge estimates likely used in RBCA and other site-specific modeling efforts.

A final recommendation pertains to evaluating the existing database compiled in the Appendix A of this report and making the information more usable. The existing data should be statistically analyzed to identify a correlation between precipitation and recharge for various physiographic provinces and climatic regimes. Factors such as evapotranspiration, soil texture, water-table depth, and topography, for example, could be incorporated into the recharge assessment through step-wise, multiple regression methods. There may be different relationships that apply to different areas of the country, in which case the results of the statistical correlation could be presented graphically on a set of maps showing areas where recharge studies have been conducted, the results of those studies, and the regressed relationships. Such a set of maps would likely prove extremely useful to those interested in inexpensive, approximate estimates of recharge at a given location and setting.

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Glossary

GLOSSARY

Advection: fluid migration induced by hydraulic pressure gradients.

Air-entry value: the value for matric potential at which water starts to drain from, and air begins to enter into, the larger pores of a porous medium.

Anion exclusion: electrostatic repulsion of negatively charged molecules by negatively charged particles within a fluid-solid system.

Anistropy: a characteristic of the transmissive properties of geologic media in which the value of the property depends upon the direction that the property is measured.

Available water: the difference between water content at field capacity and permanent wilting.

Bulk density: mass of dry soil per unit volume of bulk soil; dry bulk density.

Capillary pressure: the difference in non-wetting and wetting fluid potentials; identical to matric potential if the air, the non-wetting fluid, is at atmospheric pressure.

Capillary fringe: that part of the vadose zone immediately above a water table where the media is satiated but the water is under tension.

Contact angle: the angle created by the interface between a solid surface and fluid phases in contact with that surface.

Deep percolation: infiltration below the root zone depth, usually in the context of irrigation.

Diffuse recharge: recharge from the vadose zone which originates as infiltration of rain and snow melt over large portions of a watershed.

Diffusion: a transport process in which chemicals migrate in fluid due to concentration gradients.

Dispersivity: a characteristic of the geologic medium describing tortuosity and heterogeneity which affect mechanical mixing of chemicals during advection.

Distribution coefficient: the partitioning coefficient for a chemical, usually between water and the solid phases.

Elevation head: gravitation (elevation) potential expressed in units of potential energy per unit weight of fluid.

Evapotranspiration: the process of water discharge from the vadose zone by direct evaporation of soil water and uptake by plant roots.

Field capacity: water content of a field soil two to three days following a thorough irrigation; water content at -0.1 or -0.33 bars soil-water potential.

Field-saturated hydraulic conductivity: the term used to describe the hydraulic conductivity of soils that are "field saturated" wherein entrapped air exists within a portion of the pores.

Gaining stream: a stream or reach of stream that receives water from the zone of saturation; its channel lies below the water table.

Gravimetric water content: mass of water per unit mass of dry soil.

Gravitational potential: potential energy of soil-water due to its position above a reference datum; identical to elevation potential.

Henry's Law constant: the partitioning coefficient for a chemical between the gas and liquid phases.

Hydraulic conductivity: volumetric rate of fluid flow per unit cross-sectional area under a unit hydraulic gradient; ability of media to conduct a particular fluid.

Hydraulic gradient: fluid driving force per unit weight of fluid.

Hydrodynamic dispersion: a transport process causing chemical mixing in the geologic media attributable to the net effects of molecular diffusion and mechanical mixing in the advected liquid.

Hysteresis: a phenomenon describing a relationship between parameters, usually pertaining to water transmission characteristic or storage parameters, in which the parameter values depend on nature of the process.

Imbibition: the process of wetting a geologic median with water.

Infiltration: The entry into the soil of water made available at the ground surface and the associated downward flow within the unsaturated zone.

Infiltration rate: the rate of water movement into the soil per unit area due to gravity, capillary and pressure forces, usually associated with vertical flux density across the soil surface.

Interfacial tension: represents the potential energy associated with the surface area separating two immiscible fluids that is attributed to differences in cohesive forces between molecules of the respective fluids.

Local recharge: recharge from the vadose zone which originates from concentrated surface runoff in channels or seepage from impoundments.

Losing stream: a stream or reach of stream that contributes water to the zone of saturation; its channel lies above the water table.

Macropore flow: fluid flow within macropores such as root and worm holes, animal burrows, fractures, and large interconnected pores.

Matric potential: the component of soil-water potential relative to a reference state due to capillary and adsorptive forces which hold water in porous or fractured media.

Mean pore water velocity: volumetric flow rate of groundwater per unit area of connected pore space. The mean pore water velocity varies from specific discharge by incorporating effective porosity or effective water content to specify the cross-sectional area of pore space across which groundwater flow actually occurs.

Non-wetting fluid: in a mixture of phase-separated fluids, the non-wetting fluid is the fluid which does not preferentially wet soil particles.

Osmotic potential: the potential energy relative to a reference state attributed to chemical concentrations in the soil water.

Partitioning: a transport process in which chemicals migrate between solid, liquid, and gas phases.

Perched aquifer: an aquifer within the vadose zone created by a relatively low permeable perching layer.

Permanent wilting: water content of soil at which plants become so dry that the plant cannot survive even if the soil is rewetted; water content at -15 bars soil-water potential.

Permeability: an intrinsic property of the porous or fractured media describing the fluid transmissive character.

Phreatic zone: regional zone of saturation underlying the vadose zone.

Phreatophytes: water-loving plants that live with their roots below the water table and extract their moisture requirements directly from the saturated zone.

Playa: a dry or shallow, ephemeral lake in the lowest part of an undrained arid basin.

Porosity: volume of voids per unit bulk volume of soil.

Pressure head: soil-water potential expressed in units of potential energy per unit weight of fluid.

Pressure potential: the potential energy relative to a reference state due to air pressure or hydrostatic pressure.

Recharge: the entry of water into an aquifer across the water table surface, expressed as a rate with units of velocity.

Redistribution: the simultaneous movement of soil water upward due to evapotranspiration and downward due to infiltration.

Relative permeability: ratio of permeability at field water content to the permeability at saturation.

Relative humidity: ratio of vapor pressure at ambient conditions to saturated vapor pressure under the same conditions.

Relative hydraulic conductivity: ratio of hydraulic conductivity at field water content to the saturated hydraulic conductivity.

Retardation factor: velocity of a chemical relative to the mean pore water velocity.

Satiated: maximum saturation achievable under prevailing field conditions.

Saturation percentage: volume of fluid per unit volume of void space.

Soil-water characteristic curve: the relationship between pressure head or soil-water potential and water content of a porous or fractured media; the soil-water retention curve.

Soil-water diffusion coefficient: a mass transfer property of the medium describing the chemical mass flux due to the concentration gradient; constant of proportionality in Fick's law of diffusion for soil.

Soil-water diffusivity: an unsaturated soil property which is the ratio of hydraulic conductivity and specific water capacity; a constant of proportionality relating soil-water flux and water content gradient.

Soil-water flux: volumetric rate of fluid flow per unit cross-sectional area perpendicular to flow.

Soil-water potential: the potential energy of water in the vadose zone relative to a reference state due to the sum of matric, pressure, and osmotic potential.

Soil-water retention curve: soil-water characteristic curve.

Specific discharge: volumetric flow rate of groundwater per unit surface area of aquifer material or of porous media transmitting water. Specific discharge ("q") has units of velocity and is proportional to hydraulic conductivity (see mean pore water velocity).

Specific yield: volume of water released from or taken into storage per unit horizontal area of an unconfirmed porous or fractured media per unit change in water table elevation; storage coefficient of media under unconfined conditions.

Specific retention: water content at which the water phase becomes virtually discontinuous.

Specific moisture (water) capacity: the volume of water released from or taken into storage per unit horizontal area of geologic medium per unit change in pressure head; slope of the soil-water characteristic (water content versus pressure head) curve.

Surface Run-On: surface-water flow onto a given area.

Stochastic: of, or having to do with, processes that are controlled by random mechanisms or events.

Thermal diffusion coefficient: constant of proportionality relating soil-water plus vapor flux and temperature gradients.

Thermal diffusivity: the ratio of thermal conductivity to heat capacity.

Total hydraulic head: total soil-water potential expressed in units of potential energy per unit weight of fluid.

Total soil-water potential: the sum of soil-water potential and gravitational potential.

Unsaturated media: porous or fractured media where the voids are occupied by both water and air phases.

Vadose zone: geologic media between the land surface and the regional water table.

Void ratio: volume of voids per unit volume of solids.

Volumetric water content: volume of water per unit volume of bulk soil; also, water content.

Water-entry value: the value for matric potential at which water starts to enter the finer pores of a porous medium.

Water-retention curve: soil-water retention curve; soil-water characteristic curve.

Water content: volumetric water content.

Water table: surface in a geologic medium where water pressure equals atmospheric pressure.

Wetting fluid: in a mixture of phase-separated fluids, the wetting fluid is the fluid which preferentially wets soil particles.

Appendix A

Estimates of Diffuse Annual Recharge

Table A-1. Estimates of diffuse annual recharge by geographic region
Page 1 of 14

Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Arkansas, White and Red River Basins	High Plains of Colorado	300-350	No report	Sparse	Plains	Sand and sandstone	<25 [<~7.7]	No report	Bedinger and Sniegocki, 1976
Arkansas, White and Red River Basins	High Plains of Colorado	360-760	1015-1650 [a]	Sparse	Plains	Ogallala sandstone	6.4-152 [1-27]	Soil water balance	Dugan et al., 1994
Arkansas, White and Red River Basins	Great Plains of New Mexico, Colorado, Texas, Kansas, and Oklahoma	400-600	No report	Sparse	Plains	Terrace alluvium (sand and gravel)	1.3-13 [~0.26-2.6]	No report	Bedinger and Sniegocki, 1976
Arkansas, White and Red River Basins	Central Midwest Regional Aquifer	305-1270	810-1370 [a]	Sparse	Plains	Ogallala sandstone	2.54-381 [0.4-48]	Soil water balance	Dugan and Pekenpaugh, 1985
Arkansas, White and Red River Basins	Ozark Plateau, Ouachita Highlands and Coastal Plain	910-1420	No report	Abundant	River valley	Alluvium	6-13 [~0.5-1.1]	No report	Bedinger and Sniegocki, 1976
Arkansas, White and Red River Basins	Great Bend Prairie, Arkansas Basin, central Kansas	585	47-60 [b]	Sparse	Low relief sand-dune prairie	Alluvial	56 [10]	Hybrid water-level fluctuation	Sophocleous, 1992
Arkansas, White and Red River Basins	Southern Kansas	400-1015	1170-1625 [a]	Sparse to moderate	Rolling plain	Alluvial	83 [~12]	Soil-water model	Hansen, 1991
Arkansas, White and Red River Basins	Coastal Plain of Arkansas, Texas, and Louisiana	1000-1270	No report	Abundant	Low plain	Unconsolidated sand	100-150 [~8.8-13.0]	No report	Bedinger and Sniegocki, 1976
Arkansas, White and Red River Basins	Ozark Plateau of Missouri and Arkansas	1120-1220	No report	Abundant	Plateau and river valley	Carbonate and gypsum	~ 500 [~43]	No report	Bedinger and Sniegocki, 1976

Table A-1. Estimates of diffuse annual recharge by geographic region

Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Arkansas, White and Red River Basins	Southern Rocky Mountains, Great Plains, and Central Lowland	305-910	No report	Variable croplands	River valley	Alluvium	6-13 [~1-2.1]	No report	Bedinger and Sniegocki, 1976
California Region	West San Francisco, California	503	890 [a]	None	Hilly with impervious surface	Unconsolidated sands and silty sands	46 [9.1]	Soil-water balance using hydrologic routing model	Phillips et al., 1993
California Region	West San Francisco, California	503	890 [a]	Abundant	Hilly with irrigated surface	Unconsolidated sands and silty sands	320 [64]	Soil-water balance using hydrologic routing model	Phillips et al., 1993
California Region	West San Francisco, California	503	890 [a]	Sparse	Hilly with nonirrigated surface	Unconsolidated sands and silty sands	27 [5.4]	Soil-water balance using hydrologic routing model	Phillips et al., 1993
California Region	East San Francisco, California	503	890 [a]	None	Hilly with impervious surface	Unconsolidated sands and silty sands	33 [6.6]	Soil-water balance using hydrologic routing model	Phillips et al., 1993
California Region	East San Francisco, California	503	890 [a]	Abundant	Hilly with irrigated surface	Unconsolidated sands and silty sands	310 [62]	Soil-water balance using hydrologic routing model	Phillips et al., 1993
California Region	East San Francisco, California	503	890 [a]	Sparse	Hilly with nonirrigated surface	Unconsolidated sands and silty sands	23 [4.6]	Soil-water balance using hydrologic routing model	Phillips et al., 1993
California Region	Golden Gate Park, San Francisco, California	503	890 [a]	Abundant/ varied	Hilly	Unconsolidated sands and silty sands	127 [25]	Tritium	Phillips et al., 1993
California	Pacific Coast at Santa Cruz	~600	No report	None	Gently sloped	Clays, clayey sands, mudstone	90-640 [15-53]	Plane of zero flux	Dreiss and Anderson, 1985

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Great Basin	Beatty, Nye County, Nevada	74	1900	Sparse to none	Basin floor	Unconsolidated sand and gravel	0.04 [0.054]	Soil-water balance	Nichols, 1987
Great Basin	NA	280	No report	Variable/ sparse	Mountains, alluvial valleys, and closed basins	Variable	14 [5]	No report	Eakin et al., 1976
Great Basin	Amargosa River Basin	100-200 (avg. 180)	No report	Sparse	Basin and mountain front	Alluvial basin	0.3 [0.16]	Field scale water balance model	Osterkamp et al., 1994
Great Lakes Region	Hiram, Ohio	1,043	610	Abundant	Flat-lying river valley	Sand and gravel	433 [42]	Soil-water balance	Lyford and Cohen, 1988
Great Lakes Region	Penn Yan, New York	792	508	Abundant	Low plain	Sand and gravel	284 [36]	Soil-water balance	Lyford and Cohen, 1988
Lower Colorado River Basin	NA	150-1020	No report	Sparse	Varied, mountainous	Alluvial basin sediments	1.5-10 [1]	Soil-water balance	Davidson, 1979
Lower Mississippi River Basin	Northern section	1170-1320	840-990 [a]	Abundant	Flat-lying river valley	Alluvial	~ 25 [~2]	Soil-water balance [3]	Terry et al., 1979
Lower Mississippi River Basin	Southern section	1320-1625	990-1120	Abundant	Flat-lying river valley	Alluvial	~ 50 [~3.4]	Soil-water balance [3]	Terry et al., 1979
Mid-Atlantic Region	Copperstown, New York	998	508	Abundant	Flat-lying river plain?	Sand and gravel	491 [49]	Soil-water balance	Lyford and Cohen, 1988
Mid-Atlantic Region	Woodstock, Vermont	1,002	483	Abundant	Mountains?	Sand and gravel	519 [52]	Soil-water balance	Lyford and Cohen, 1988
Mid-Atlantic Region	Stroudsberg, Pennsylvania	1,219	559	Abundant	Flat-lying river valley?	Sand and gravel	661 [54]	Soil-water balance	Lyford and Cohen, 1988

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Missouri River Basin	Ogallala aquifer, Colorado, Kansas, and Nebraska	400-500	1170-1520 [c]	Sparse	High plain	Semiconsolidated gravel, sand, silt and clay	12-75 [~2.7-17]	No report	Taylor, 1978
Missouri River Basin	Akron, Colorado	No report	No report	None	High plains	Fallow soil	61	Soil-water model based on Richard's Equation	Krishnamurthi et al., 1977; Longenbaugh, 1975
Missouri River Basin	High Plains	356-762	1016-1651 [a]	Sparse	High plains	Ogallala sandstone	6.4-152 [~1-27]	Soil-water balance	Dugan et al., 1994
Missouri River Basin	Central midwest regional aquifer	305-1270	813-1372 [a]	Sparse	High plains	Ogallala sandstone	2.54-381 [~0.4-48]	Soil-water balance	Dugan and Pekenpaugh, 1985
Missouri River Basin	Northeast Kansas	760-1015	1170-1220 [a]	Sparse to moderate	Rolling plains	Glacial drift	120 [~14]	Soil-water model	Hansen, 1991
Missouri River Basin	North central Kansas	630-1015	1170-1370 [a]	Sparse to moderate	Rolling plains	No report	60 [~9]	Soil-water model	Hansen, 1991
Missouri River Basin	Cheyenne, Wyoming	381	No report	Grasslands	Rolling plains	Sand and gravel	22 [5.8]	Streamflow measurements and soil-water balance	Morgan, 1946
New England	NA	1000-1140	No report	Abundant	Low mountains and coastal	Glacial drift and alluvium	530 [49]	No report	Sinnott, 1982
New England	NA	1000-1140	No report	Abundant	Low mountains and coastal	Glacial till and bedrock	107 [10]	No report	Sinnott, 1982
New England	Milford-Souhegan	1118	No report	Abundant	Low mountain	Glacial drift	333 [30]	Basin outflow	Harte and Mack, 1992

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type			
New England	Lower Merrimale and Coastal River Basins	No report	No report	Abundant	Low mountain and coastal	Stratified drift	483	Streamflow measurements	Stekl and Flanagan, 1992
New England	Bellamy, Cochero, and Salmon Falls River Basins	1092	457-610	Abundant	Low mountain and coastal	Stratified drift	508 [46]	Streamflow measurements	Mack and Lawlor, 1992
New England	Nashua, New Hampshire	1099	508	Abundant	River valley?	Sand and gravel	591 [54]	Soil-water balance	Lyford and Cohen, 1988
New England	Middlestown, Connecticut	1273	610	Abundant	River valley?	Sand and gravel	663 [52]	Soil-water balance	Lyford and Cohen, 1988
Ohio River Basin	Lower Ohio River Subbasin	1220	914[c]	Abundant	River valley	Alluvium, glacial outwash	110 [9]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	Muskingum Subbasin	1003	826[c]	Abundant	River valley	Alluvium, glacial outwash	90 [9]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	White Subbasin	1067	838[c]	Abundant	River valley	Alluvium, glacial outwash	107 [10]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	Wabash Subbasin	1016	864[c]	Abundant	River valley	Alluvium, glacial outwash	111 [11]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	Allegheny Subbasin	1067	686[c]	Abundant	River valley	Alluvium, glacial outwash	190 [18]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	Sandy-Guyandotte Subbasins	1118	838[c]	Abundant	River valley	Alluvium, glacial outwash	90 [8]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	Upper Ohio River Subbasin	1016	788[c]	Abundant	River valley	Alluvium, glacial outwash	100 [10]	Streamflow measurements[4]	Bloyd, 1974
Ohio River Basin	Cumberland Subbasin	1245	914[c]	Abundant	River valley	Alluvium, glacial outwash	125 [10]	Streamflow measurements[4]	Bloyd, 1974

Table A-1. Estimates of diffuse annual recharge by geographic region

Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Ohio River Basin	Miami Subbasins	978	838 [c]	Abundant	River valley	Alluvium, glacial outwash	98 [10]	Streamflow measurements [4]	Bloyd, 1974
Ohio River Basin	Scioto Subbasin	1003	864 [c]	Abundant	River valley	Alluvium, glacial outwash	30 [3]	Streamflow measurements [4]	Bloyd, 1974
Ohio River Basin	Kanawha Subbasins	1130	813 [c]	Abundant	River valley	Alluvium, glacial outwash	180 [16]	Streamflow measurements [4]	Bloyd, 1974
Ohio River Basin	Licking-Kentucky Subbasins	1143	889 [c]	Abundant	River valley	Alluvium, glacial outwash	60 [5]	Streamflow measurements [4]	Bloyd, 1974
Ohio River Basin	Monongahela Subbasin	1194	737 [c]	Abundant	River valley	Alluvium, glacial outwash	215 [18]	Streamflow measurements [4]	Bloyd, 1974
Ohio River Basin	Wright-Patterson AFB	965	No report	Abundant	Central lowlands	Glacial sand and gravels	315-401 [33-42]	Water level fluctuation	Dumouchelle et al., 1993
Ohio River Basin	Geauga County	991-1092	No report	Abundant	River valley	Sandstone/shale uplands, glacial deposits	51-203 [4.9-19]	Basin outflow (stream hydrograph)	WRIR 90-4026
Ohio River Basin	Williams County	864	No report	Abundant	Flat to rolling	Morraine and lake deposits	51-203 [5.9-23]	Basin outflow (stream hydrograph)	Coen, 1989
Ohio River Basin	Killbuckle	No report	No report	Abundant	River valley	Glacial till and valley train outwash	72-236	Groundwater model	Breen et al., 1994
Pacific Northwest	Hanford, Washington	162	1600 [c]	Varied thistle and cheatgrass	Flat	Loamy sand	5 [3]	Lysimeter/Darcy flux with unit gradient	Gee et al., 1994
Pacific Northwest	Columbia Plateau, Snake River Plain	<300	No report	Sparse	Plains and low mountains	Basalt and sedimentary	<25 [<8.3]	No report	Foxworthy, 1979

Table A-1. Estimates of diffuse annual recharge by geographic region

Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Pacific Northwest	Columbia Plateau of Washington, Oregon, and Idaho	380-620	710-1140 [a]	Grasslands and forest	Rolling hills	Basalt with overlying unconsolidated sediments	Up to 190 [~38]	Soil-water model	Bauer and Vaccaro, 1990
Pacific Northwest	Columbia Plateau of Washington, Oregon, and Idaho	165-380	710-1140 [a]	Sage and grasslands	Low lying basins and rolling hills	Basalt with overlying unconsolidated sediments	25-30 [7.5-17]	Soil-water model	Bauer and Vaccaro, 1990
Rio Grande Basin	NA	200-760 Avg = 305	290 [a]	Variable and sparse	Mountains and alluvial valleys	Variable	~ 14 [~4.6]	Soil-water balance	West and Broadhurst, 1975
Rio Grande Basin	Socorro, New Mexico	179	1780 [c]	Sparse		Sand	7-37 [20]	Darcy flux	Stephens & Knowlton, 1986
Rio Grande Basin	Socorro, New Mexico	200	1780 [c]	Sparse (four-winged saltbrush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	8.4 [4.2]	Tritium peak	Phillips et al., 1988
Rio Grande Basin	Socorro, New Mexico	200	1780 [c]	Sparse (four-winged saltbush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	2-2.5 [1-1.3]	Chloride mass balance	Phillips et al., 1988
Rio Grande Basin	Socorro, New Mexico	200	1780 [c]	Sparse (four-winged saltbush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	7.0 [3.5]	Darcy flux (harmonic mean K)	Phillips et al., 1988

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Rio Grande Basin	Socorro, New Mexico	200	1780 [c]	Sparse (four-winged saltbush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	2.6-3.0 [1.3-1.5]	Chlorine-36 peak	Phillips et al., 1988
Rio Grande Basin	Socorro, New Mexico	233	1780 [a]	Sparse	No report	No report	2.8 [1.2]	Chlorine-36	Trotman, 1983
Rio Grande Basin	El Paso, Hudspeth County, Texas	280	~1960 [c]	Sparse	Flat alluvial plain	Clay, silt, sand	1.4 [0.5]	Chlorine-36 peak	Scanlon, 1992
Rio Grande Basin	El Paso, Hudspeth County, Texas	280	~1960 [c]	Sparse	Flat alluvial plain	Clay, silt, sand	0.9 [0.3]	Chloride mass balance	Scanlon, 1992
Rio Grande Basin	El Paso, Hudspeth County, Texas	280	~1960 [c]	Sparse	Flat alluvial plain	Clay, silt, sand	7 [2.5]	Tritium peak	Scanlon, 1992
Rio Grande Basin	Las Cruces, New Mexico	230	2390 [c]	None	Flat	Loamy fine sand	87 [25]	Lysimeter	Gee et al., 1994
Rio Grande Basin	Las Cruces, New Mexico	230	2390 [c]	Sparse (grass and shrubs)	Terrace/ pediment	Sandy loam	1.5 [0.65]	Chloride mass balance	Phillips et al., 1988
Rio Grande Basin	Las Cruces, New Mexico	230	2390 [c]	Sparse (grass and shrubs)	Terrace/ pediment	Sandy loam	9.5 [4]	Tritium peak	Phillips et al., 1988
Rio Grande Basin	Las Cruces, New Mexico	230	2390 [c]	Sparse (grass and shrubs)	Terrace/ pediment	Sandy loam	2.5 [1.1]	Chlorine-36 peak	Phillips et al., 1988
Rio Grande Basin	Las Cruces, New Mexico	228	2464 [a]	Sparse to none	Playa	Silts, sands, clays	5.59 [2.44]	Chloride mass balance	Stone, 1986
Rio Grande Basin	Roswell Basin	305-373	No report	Sparse	Mountain slope to river	Limestone	[7.3]	Basin outflow	Fleiller & Nye, 1933
Rio Grande Basin	Roswell Basin	305-373	No report	Sparse	Mountain slope to river	Limestone	[10]	Basin outflow	Hantush, 1957

Table A-1. Estimates of diffuse annual recharge by geographic region

Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type			
Rio Grande Basin	Roswell Basin	305-373	No report	Sparse	Mountain slope to river	Limestone	[3.3-7.8]	Tritium in aquifer	Gross et al., 1979
Rio Grande Basin	Curry County, east-central New Mexico	385	>1000[a]	Sparse	High plains	Sand	0.75 [0.19]	Chloride mass balance	Stone, 1986
Rio Grande Basin	San Juan Basin, northwest New Mexico	145	1420[a]	Sparse to none	Alluvial terrace	Alluvium	2.29 [1.58]	Chloride mass balance	Stone, 1986
Rio Grande Basin	San Juan Basin, northwest New Mexico	145	1420[a]	Sparse to none	Badlands	Eolian sand, unconsolidated sediments	0.25-0.89 [0.17-0.61]	Chloride mass balance	Stone, 1986
Rio Grande Basin	Sunland Park, New Mexico	200	1270[a]	Sparse (creosote bush)	Gently sloping escarpment	Sandy alluvium	0.19 [0.095]	Chloride mass balance	Stephens & Coons, 1994
Rio Grande Basin	Sunland Park, New Mexico	200	1270[a]	Sparse (creosote bush)	Gently sloping escarpment	Sandy alluvium	0.066 [0.033]	Soil-water model (HELP)	Stephens & Coons, 1994
Rio Grande Basin	Sunland Park, New Mexico	200	1270[a]	Sparse (creosote bush)	Gently sloping escarpment	Sandy alluvium	0.16 [0.08]	Darcy flux	Stephens & Coons, 1994
Rio Grande Basin	Quemado, west-central New Mexico	251	788[a]	Sparse	Mesa, valley floor	Alluvium and sedimentary bedrock	1.27-2.03 [0.51-0.81]	Chloride mass balance	Stone, 1986
South Atlantic Gulf Region	NA	1120-1630	No report	Abundant	Coastal plain and rolling highlands	Unconsolidated and consolidated sediments	150 [10.9]	Streamflow measurements[5]	Cederstrom et al., 1979
South Atlantic Gulf Region	South Carolina	No report	No report	Abundant	Steep slopes	Clayey	<25	No report	Patterson, 1995

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type			
South Atlantic Gulf Region	South Carolina	1450	1015[a]	Sparse	Sand hills of coastal plain	Sand	350-380 [~25]	Soil-water balance	Patterson, 1995
South Atlantic Gulf Region	Florida	No report	No report	Abundant	Flat coastal plain	Unconsolidated sediments	0-110	Back calibration of groundwater model	Graham and Neff, 1994
Tennessee River Basin	Cumberland Plateau	1270-1520	760-910[a]	Abundant	Level highland	Fractured noncarbonate	230-355 [10-31]	Streamflow measurements[6]	Zurawski, 1978
Tennessee River Basin	Cumberland Plateau, Tennessee	No report	No report	Abundant	Gently rolling to hilly	Flat-lying sandstone, shale and conglomerate	165 [~12][d]	Streamflow measurements and soil-water balance	Hoos, 1990
Tennessee River Basin	Highland Rim	1270	760[a]	Abundant	Rolling plateau	Carbonate with alluvium	290-405 [23-33]	Streamflow measurements[6]	Zurawski, 1978
Tennessee River Basin	Highland Rim	No report	No report	Abundant	Undulating to hilly and steep	Flat-lying carbonate	190 [~15][d]	Streamflow measurements and soil-water balance	Hoos, 1990
Tennessee River Basin	Coastal Plain/Big Sandy River	1270	760[a]	Abundant	River valley and rolling plain	Unconsolidated sand	215-240 [17-20]	Streamflow measurements[6]	Zurawski, 1978
Tennessee River Basin	Valley and Ridge	1020-1520	610-910[a]	Abundant	Parallel valleys and ridges	Carbonate with alluvium	140-240 [13-23]	Streamflow measurements[6]	Zurawski, 1978
Tennessee River Basin	Valley and Ridge, Tennessee	No report	No report	Abundant	Parallel valleys and ridges	Folded/faulted sandstone, carbonate and shale	170 [~12][d]	Streamflow measurements and soil-water balance	Hoos, 1990

Table A-1. Estimates of diffuse annual recharge by geographic region

Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type			
Tennessee River Basin	Blue Ridge	1020-1780	610-1070[a]	Abundant	Mountainous	Fractured noncarbonate with alluvium	215-240 [17-20]	Streamflow measurements[6]	Zurawski, 1978
Tennessee River Basin	Blue Ridge	No report	No report	Abundant	Mountainous	Fractured crystalline bedrock	300 [~21][d]	Streamflow measurements and soil-water balance	Hoos, 1990
Tennessee River Basin	Central Basin	1270	760[a]	Abundant	Interior low plateau	Carbonate	255-380 [20-31]	Streamflow measurements[6]	Zurawski, 1978
Tennessee River Basin	Central Basin, Tennessee	No report	No report	Abundant	Gently rolling to hilly and steep	Flat-lying carbonate	140 [~11][d]	Streamflow measurements and soil-water balance	Hoos, 1990
Texas-Gulf Region	Regional alluvial aquifers	510-1120	510-1520[c]	Moderate	Floodplain	Sand, gravel, and clay	34 [4.2]	Soil-water balance[7]	Baker & Wall, 1976
Texas-Gulf Region	Ogallala Aquifer of Texas High Plains	410-510	1520[c]	Sparse	High plain	Sand and gravel	2.3 [0.5]	Soil-water balance[7]	Baker & Wall, 1976
Texas-Gulf Region	Ogallala Aquifer at Portales, New Mexico	~355	No report	Sparse	Depression in high plains with sand dunes	Sand and gravel	12.7 [3.6]	Basin outflow	Theis, 1937
Texas-Gulf Region	Ogallala Aquifer in Lea County, New Mexico	~355	No report	Sparse	High plains	Sand and gravel	3.2-16.9 [0.9-4.8]	Basin outflow	Theis, 1937
Texas-Gulf Region	Ogallala Aquifer in Lea County, New Mexico	~355	No report	Sparse	High plains	Sand and gravel	6.4-10.7 [1.8-3.0]	Water level fluctuations	Theis, 1937
Texas-Gulf Region	Ogallala Aquifer in Lea County, New Mexico	~355	460[a]	Sparse	High plains	Sand and gravel	9.6	Back calibration of groundwater model	McAda, 1984

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E^2	Vegetation	Topography	Soil/Aquifer Type			
Texas-Gulf Region	Carrizo-Wilcox, Queen City and Sparta Aquifers, Texas Coastal Plain	610-1320	0-1270	Moderate	Low plain	Interbedded sand and clay	~ 7 [~0.7]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Woodbine Aquifer of Texas Coastal Plain	860-1070	760-1020 [c]	Moderate	Low plain	Sand, sandstone, shale	<1 [<~0.1]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Edwards-Trinity Plateau	400-710	1270-1520 [c]	Sparse	Rolling plain	Limestone, sand, sandstone	17 [3.1]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Edwards-Balcones fault zone	610-810	900-1400 [c]	Moderate	Fractured escarpment	Fractured limestone and dolomite	78 [11]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Santa Rosa aquifer, west Texas	510	1520 [c]	Variable and sparse	Low plain	Sand/gravel	14 [2.7]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Trinity Aquifer, central Texas	710-910	760-1020 [c]	Variable and sparse	Low plain	Sand, shale, limestone	1.7 [0.2]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Ellenburger-San Saba Aquifer of Llano Uplift, Texas	610-810	1020-1270 [c]	Variable and sparse	Rolling plain	Limestone and dolomite	2.4 [0.34]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Gulf Coast aquifer of Texas Coastal Plain	610-1420	250-1270 [c]	Moderate	Coastal plain	Interbedded sand, clay, and gravel	34 [3.3]	Soil-water balance [7]	Baker & Wall, 1976
Texas-Gulf Region	Hickory Aquifer of Edwards Plateau, Texas	560-710	1020-1520 [c]	Variable and sparse	Rolling plain	Sand and sandstone	4 [0.63]	Soil-water balance [7]	Baker & Wall, 1976
Upper Colorado River Basin	NA	400	No report	Variable	Mountains and alluvial valleys	Variable	16 [4]	Soil-water balance	Price and Arnow, 1974
Upper Mississippi River Basin	Wisconsin Subbasin	813	610 [a]	Abundant	River valley	Alluvium, glacial outwash	120 [15]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Kaskaskia Subbasin	965	775 [a]	Abundant	River valley	Alluvium, glacial outwash	60 [6]	Streamflow measurements [4]	Bloyd, 1975

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Upper Mississippi River Basin	Rock Subbasin	838	673 [a]	Abundant	River valley	Alluvium, glacial outwash	110 [13]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Chippewa Black Subbasin	762	610 [a]	Abundant	River valley	Alluvium, glacial outwash	90 [12]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Big Muddy Subbasin	1067	787 [a]	Abundant	River valley	Alluvium, glacial outwash	32 [3]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Mississippi Headwaters	660	610 [a]	Abundant	River valley	Alluvium, glacial outwash	90 [14]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Merames Subbasin	1016	787 [a]	Abundant	River valley	Alluvium, glacial outwash	90 [9]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Minnesota Subbasin	610	635 [a]	Abundant	River valley	Alluvium, glacial outwash	18 [3]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Illinois Subbasin	864	724 [a]	Abundant	River valley	Alluvium, glacial outwash	50 [6]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Mahomet Valley, east-central Illinois	No report	No report	Abundant	River valley	Glacial outwash, sand and gravel	90	Soil-water balance	Panno et al., 1994 from Visocky and Schicht (1969)
Upper Mississippi River Basin	Lake County, Illinois	No report	No report	Moderate	River valley	Glacial drift (clay, sand and gravel); dolomite bedrock	78-310	Soil temperature	Boyle and Saleem, 1979
Upper Mississippi River Basin	Lake County, Illinois	No report	No report	Moderate	River valley	Glacial drift (clay, sand and gravel); dolomite bedrock	28	Basin outflow [8]	Prickett et al., 1964; Walton, 1965

Table A-1. Estimates of diffuse annual recharge by geographic region
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Geographic Region	Site	Site Hydrologic Conditions*					Estimated Annual Recharge (mm/yr [%P])	Estimation Technique	Reference
		MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type			
Upper Mississippi River Basin	Lake County, Illinois	No report	No report	Moderate	River valley	Glacial drift (clay, sand and gravel); dolomite bedrock	80-107	Basin outflow [8]	Schicht et al., 1976
Upper Mississippi River Basin	Des Moines Subbasin	787	686 [a]	Abundant	River valley	Alluvium, glacial outwash	24 [3]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Skunk Subbasin	813	686 [a]	Abundant	River valley	Alluvium, glacial outwash	50 [6]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Cannon-Zumbro-Root Subbasins	737	635 [a]	Abundant	River valley	Alluvium, glacial outwash	74 [10]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Iowa-Cedar Subbasins	813	686 [a]	Abundant	River valley	Alluvium, glacial outwash	50 [6]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Salt Subbasin	940	762 [a]	Abundant	River valley	Alluvium, glacial outwash	19 [2]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Wapsipinicon Subbasin	813	660 [a]	Abundant	River valley	Alluvium, glacial outwash	80 [10]	Streamflow measurements [4]	Bloyd, 1975
Upper Mississippi River Basin	Fox-Wyaconda-Fabius Subbasins	889	737 [a]	Abundant	River valley	Alluvium, glacial outwash	18 [2]	Streamflow measurements [4]	Bloyd, 1975

* Climate classifications according to Thornthwaite (1948):
Arid: MAP/E <0.5
Semiarid: 0.5 < MAP/E < 1.0
Humid: MAP/E >1.0

1 Mean annual precipitation (mm/yr)
2 Evaporation (mm/yr)
3 Soil-water balance computations based on data provided for mean annual precipitation, evapotranspiration and runoff
4 Assumes a 60-percent flow parameter equals baseflow
5 Average recharge estimated from average baseflow
6 Range of recharge estimated from average baseflow
7 Recharge, approximated by steady-state yield, may be overestimated
8 Where basin outflow is limited to known pumping discharge

a Potential evapotranspiration
b Complimentary relationship areal evapotranspiration (CRAE) (Morton, 1983)
c Lake evaporation
d Report of annual recharge estimate as %P based on MAP reported by Zurawski (1978)

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Basin outflow [3]	Upper Mississippi River Basin	Lake County, Illinois	No report	No report	Moderate	River valley	Glacial drift (clay, sand and gravel); dolomite bedrock	28	Prickett et al., 1964; Walton, 1965
Basin outflow [3]	Upper Mississippi River Basin	Lake County, Illinois	No report	No report	Moderate	River valley	Glacial drift (clay, sand and gravel); dolomite bedrock	80-107	Schicht et al., 1976
Basin outflow	New England	Milford-Souhegan	1118	No report	Abundant	Low mountain	Glacial drift	333 [30]	Harte and Mack, 1992
Basin outflow	Texas-Gulf Region	Ogallala Aquifer at Portales, New Mexico	~355	No report	Sparse	Depression in high plains with sand dunes	Sand and gravel	12.7 [3.6]	Theis, 1937
Basin outflow	Texas-Gulf Region	Ogallala Aquifer in Lea County, New Mexico	~355	No report	Sparse	High plains	Sand and gravel	3.2-16.9 [0.9-4.8]	Theis, 1937
Basin outflow	Rio Grande Basin	Roswell Basin	305-373	No report	Sparse	Mountain slope to river	Limestone	[7.3]	Fleiller & Nye, 1933
Basin outflow	Rio Grande Basin	Roswell Basin	305-373	No report	Sparse	Mountain slope to river	Limestone	[10]	Hantush, 1957
Basin outflow (stream hydrograph)	Ohio River Basin	Geauga County	991-1092	No report	Abundant	River valley	Sandstone/shale uplands, glacial deposits	51-203 [4.9-19]	USGS, 1990 WRIR 90-4026
Basin outflow (stream hydrograph)	Ohio River Basin	Williams County	864	No report	Abundant	Flat to rolling	Morraine and lake deposits	51-203 [5.9-23]	Coen, 1989
Chlorine-36	Rio Grande Basin	Socorro, New Mexico	233	1780 [a]	Sparse	No report	No report	2.8 [1.2]	Trotman, 1983

Table A-2. Estimates of diffuse annual recharge by estimation technique

Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Chlorine-36 peak	Rio Grande Basin	El Paso, Hudspeth County, Texas	280	~1960 [b]	Sparse	Flat alluvial plain	Clay, silt, sand	1.4 [0.5]	Scanlon, 1992
Chlorine-36 peak	Rio Grande Basin	Las Cruces, New Mexico	230	2390 [b]	Sparse (grass and shrubs)	Terrace/ pediment	Sandy loam	2.5 [1.1]	Phillips et al., 1988
Chlorine-36 peak	Rio Grande Basin	Socorro, New Mexico	200	1780 [b]	Sparse (four-winged saltbush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	2.6-3.0 [1.3-1.5]	Phillips et al., 1988
Chloride mass balance	Rio Grande Basin	Curry County, east-central New Mexico	385	>1000 [a]	Sparse	High plains	Sand	0.75 [0.19]	Stone, 1986
Chloride mass balance	Rio Grande Basin	El Paso, Hudspeth County, Texas	280	~1960 [b]	Sparse	Flat alluvial plain	Clay, silt, sand	0.9 [0.3]	Scanlon, 1992
Chloride mass balance	Rio Grande Basin	Las Cruces, New Mexico	228	2464 [a]	Sparse to none	Playa	Silts, sands, clays	5.59 [2.44]	Stone, 1986
Chloride mass balance	Rio Grande Basin	Las Cruces, New Mexico	230	2390 [b]	Sparse (grass and shrubs)	Terrace/ pediment	Sandy loam	1.5 [0.65]	Phillips et al., 1988
Chloride mass balance	Rio Grande Basin	Quemado, west-central New Mexico	251	788 [a]	Sparse	Mesa, valley floor	Alluvium and sedimentary bedrock	1.27-2.03 [0.51-0.81]	Stone, 1986
Chloride mass balance	Rio Grande Basin	San Juan Basin, northwest New Mexico	145	1420 [a]	Sparse to none	Alluvial terrace	Alluvium	2.29 [1.58]	Stone, 1986
Chloride mass balance	Rio Grande Basin	San Juan Basin, northwest New Mexico	145	1420 [a]	Sparse to none	Badlands	Eolian sand, unconsolidated sediments	0.25-0.89 [0.17-0.61]	Stone, 1986

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Chloride mass balance	Rio Grande Basin	Socorro, New Mexico	200	1780 [b]	Sparse (four-winged saltbush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	2-2.5 [1-1.3]	Phillips et al., 1988
Chloride mass balance	Rio Grande Basin	Sunland Park, New Mexico	200	1270 [a]	Sparse (creosote bush)	Gently sloping escarpment	Sandy alluvium	0.19 [0.095]	Stephens & Coons, 1994
Darcy flux	Rio Grande Basin	Socorro, New Mexico	179	1780 [b]	Sparse		Sand	7-37 [20]	Stephens & Knowlton, 1986
Darcy flux	Rio Grande Basin	Sunland Park, New Mexico	200	1270 [a]	Sparse (creosote bush)	Gently sloping escarpment	Sandy alluvium	0.16 [0.08]	Stephens & Coons, 1994
Darcy flux (harmonic mean K)	Rio Grande Basin	Socorro, New Mexico	200	1780 [b]	Sparse (four-winged saltbush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	7.0 [3.5]	Phillips et al., 1988
Field scale water balance model	Great Basin	Amargosa River Basin	100-200 (avg. 180)	No report	Sparse	Basin and mountain front	Alluvial basin	0.3 [0.16]	Osterkamp et al., 1994
Groundwater model	Ohio River Basin	Killbuckle	No report	No report	Abundant	River valley	Glacial till and valley train outwash	72-236	Breen et al., 1994
Groundwater model back calibration	South Atlantic Gulf Region	Florida	No report	No report	Abundant	Flat coastal plain	Unconsolidated sediments	0-110	Graham and Neff, 1994
Groundwater model back calibration	Texas-Gulf Region	Ogallala Aquifer in Lea County, New Mexico	~355	460 [a]	Sparse	High plains	Sand and gravel	9.6	McAda, 1984

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Plane of zero flux	California	Pacific coast at Santa Cruz	~600	No report	None	Gently sloped	Clays, clayey sands, mudstone	90-640 [15-53]	Dreiss and Anderson, 1985
Lysimeter	Rio Grande Basin	Las Cruces, New Mexico	230	2390 [b]	None	Flat	Loamy fine sand	87 [25]	Gee et al., 1994
Lysimeter/Darcy flux with unit gradient	Pacific Northwest	Hanford, Washington	162	1600 [b]	Varied thistle and cheatgrass	Flat	Loamy sand	5 [3]	Gee et al., 1994
Soil temperature	Upper Mississippi River Basin	Lake County, Illinois	No report	No report	Moderate	River valley	Glacial drift (clay, sand and gravel); dolomite bedrock	78-310	Boyle and Saleem, 1979
Soil-water balance	Great Basin	Beatty, Nye County, Nevada	74	1900	Sparse to none	Basin floor	Unconsolidated sand and gravel	0.04 [0.054]	Nichols, 1987
Soil-water balance [4]	Texas-Gulf Region	Carrizo-Wilcox, Queen City and Sparta Aquifers, Texas Coastal Plain	610-1320	0-1270	Moderate	Low plain	Interbedded sand and clay	~ 7 [~0.7]	Baker & Wall, 1976
Soil-water balance	Arkansas, White and Red River Basins	Central Midwest Regional Aquifer	305-1270	810-1370 [a]	Sparse	Plains	Ogallala sandstone	2.54-381 [0.4-48]	Dugan and Pekenpaugh, 1985
Soil-water balance	Missouri River Basin	Central midwest regional aquifer	305-1270	813-1372 [a]	Sparse	High plains	Ogallala sandstone	2.54-381 [~0.4-48]	Dugan and Pekenpaugh, 1985
Soil-water balance	Mid-Atlantic Region	Copperstown, New York	998	508	Abundant	Flat-lying river plain?	Sand and gravel	491 [49]	Lyford and Cohen, 1988
Soil-water balance [4]	Texas-Gulf Region	Edwards-Balcones fault zone	610-810	900-1400 [b]	Moderate	Fractured escarpment	Fractured limestone and dolomite	78 [11]	Baker & Wall, 1976

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Soil-water balance [4]	Texas-Gulf Region	Edwards-Trinity Plateau	400-710	1270-1520 [b]	Sparse	Rolling plain	Limestone, sand, sandstone	17 [3.1]	Baker & Wall, 1976
Soil-water balance [4]	Texas-Gulf Region	Ellenburger-San Saba Aquifer of Llano Uplift, Texas	610-810	1020-1270 [b]	Variable and sparse	Rolling plain	Limestone and dolomite	2.4 [0.34]	Baker & Wall, 1976
Soil-water balance [4]	Texas-Gulf Region	Gulf Coast aquifer of Texas Coastal Plain	610-1420	250-1270 [b]	Moderate	Coastal plain	Interbedded sand, clay, and gravel	34 [3.3]	Baker & Wall, 1976
Soil-water balance [4]	Texas-Gulf Region	Hickory Aquifer of Edwards Plateau, Texas	560-710	1020-1520 [b]	Variable and sparse	Rolling plain	Sand and sandstone	4 [0.63]	Baker & Wall, 1976
Soil-water balance	Missouri River Basin	High Plains	356-762	1016-1651 [a]	Sparse	High plains	Ogallala sandstone	6.4-152 [~1-27]	Dugan et al., 1994
Soil water balance	Arkansas, White and Red River Basins	High Plains of Colorado	360-760	1015-1650 [a]	Sparse	Plains	Ogallala sandstone	6.4-152 [1-27]	Dugan et al., 1994
Soil-water balance	Great Lakes Region	Hiram, Ohio	1,043	610	Abundant	Flat-lying river valley	Sand and gravel	433 [42]	Lyford and Cohen, 1988
Soil-water balance	Upper Mississippi River Basin	Mahomet Valley, east-central Illinois	No report	No report	Abundant	River valley	Glacial outwash, sand and gravel	90	Panno et al., 1994 from Visocky and Schicht (1969)
Soil-water balance	New England	Middlestown, Connecticut	1273	610	Abundant	River valley?	Sand and gravel	663 [52]	Lyford and Cohen, 1988
Soil-water balance	Lower Colorado River Basin	NA	150-1020		Sparse	Varied, mountainous	Alluvial basin sediments	1.5-10 [1]	Davidson, 1979
Soil-water balance	Rio Grande Basin	NA	200-760 Avg = 305	290 [a]	Variable and sparse	Mountains and alluvial valleys	Variable	~ 14 [~4.6]	West and Broadhurst, 1975

Table A-2. Estimates of diffuse annual recharge by estimation technique

Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions: MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type	Estimated Annual Recharge (mm/yr [%P])	Reference
Soil-water balance	Upper Colorado River Basin	NA	400	No report	Variable	Mountains and alluvial valleys	Variable	16 [4]	Price and Arnow, 1974
Soil-water balance	New England	Nashua, New Hampshire	1099	508	Abundant	River valley?	Sand and gravel	591 [54]	Lyford and Cohen, 1988
Soil-water balance[5]	Lower Mississippi River Basin	Northern section	1170-1320	840-990[a]	Abundant	Flat-lying river valley	Alluvial	~ 25 [~2]	Terry et al., 1979
Soil-water balance[4]	Texas-Gulf Region	Ogallala Aquifer of Texas High Plains	410-510	1520[b]	Sparse	High plain	Sand and gravel	2.3 [0.5]	Baker & Wall, 1976
Soil-water balance	Great Lakes Region	Penn Yan, New York	792	508	Abundant	Low plain	Sand and gravel	284 [36]	Lyford and Cohen, 1988
Soil-water balance[4]	Texas-Gulf Region	Regional alluvial aquifers	510-1120	510-1520[b]	Moderate	Floodplain	Sand, gravel, and clay	34 [4.2]	Baker & Wall, 1976
Soil-water balance[4]	Texas-Gulf Region	Santa Rosa aquifer, west Texas	510	1520[b]	Variable and sparse	Low plain	Sand/gravel	14 [2.7]	Baker & Wall, 1976
Soil-water balance	South Atlantic Gulf Region	South Carolina	1450	1015[a]	Sparse	Sand hills of coastal plain	Sand	350-380 [~25]	Patterson, 1995
Soil-water balance[5]	Lower Mississippi River Basin	Southern section	1320-1625	990-1120	Abundant	Flat-lying river valley	Alluvial	~ 50 [~3.4]	Terry et al., 1979
Soil-water balance	Mid-Atlantic Region	Stroudsberg, Pennsylvania	1,219	559	Abundant	Flat-lying river valley?	Sand and gravel	661 [54]	Lyford and Cohen, 1988
Soil-water balance[4]	Texas-Gulf Region	Trinity Aquifer, central Texas	710-910	760-1020[b]	Variable and sparse	Low plain	Sand, shale, limestone	1.7 [0.2]	Baker & Wall, 1976

Table A-2. Estimates of diffuse annual recharge by estimation technique

Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type		
Soil-water balance[4]	Texas-Gulf Region	Woodbine Aquifer of Texas Coastal Plain	860-1070	760-1020[b]	Moderate	Low plain	Sand, sandstone, shale	<1 [<~0.1]	Baker & Wall, 1976
Soil-water balance	Mid-Atlantic Region	Woodstock, Vermont	1,002	483	Abundant	Mountains?	Sand and gravel	519 [52]	Lyford and Cohen, 1988
Soil-water balance using hydrologic routing model	California Region	East San Francisco, California	503	890[a]	Abundant	Hilly with irrigated surface	Unconsolidated sands and silty sands	310 [62]	Phillips et al., 1993
Soil-water balance using hydrologic routing model	California Region	East San Francisco, California	503	890[a]	None	Hilly with impervious surface	Unconsolidated sands and silty sands	33 [6.6]	Phillips et al., 1993
Soil-water balance using hydrologic routing model	California Region	East San Francisco, California	503	890[a]	Sparse	Hilly with nonirrigated surface	Unconsolidated sands and silty sands	23 [4.6]	Phillips et al., 1993
Soil-water balance using hydrologic routing model	California Region	West San Francisco, California	503	890[a]	Abundant	Hilly with irrigated surface	Unconsolidated sands and silty sands	320 [64]	Phillips et al., 1993
Soil-water balance using hydrologic routing model	California Region	West San Francisco, California	503	890[a]	None	Hilly with impervious surface	Unconsolidated sands and silty sands	46 [9.1]	Phillips et al., 1993
Soil-water balance using hydrologic routing model	California Region	West San Francisco, California	503	890[a]	Sparse	Hilly with nonirrigated surface	Unconsolidated sands and silty sands	27 [5.4]	Phillips et al., 1993

Table A-2. Estimates of diffuse annual recharge by estimation technique

Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Soil-water model	Pacific Northwest	Columbia Plateau of Washington, Oregon, and Idaho	165-380	710-1140 [a]	Sage and grasslands	Low lying basins and rolling hills	Basalt with overlying unconsolidated sediments	25-30 [7.5-17]	Bauer and Vaccaro, 1990
Soil-water model	Pacific Northwest	Columbia Plateau of Washington, Oregon, and Idaho	380-620	710-1140 [a]	Grasslands and forest	Rolling hills	Basalt with overlying unconsolidated sediments	Up to 190 [~38]	Bauer and Vaccaro, 1990
Soil-water model	Missouri River Basin	North central Kansas	630-1015	1170-1370 [a]	Sparse to moderate	Rolling plains	No report	60 [~9]	Hansen, 1991
Soil-water model	Missouri River Basin	Northeast Kansas	760-1015	1170-1220 [a]	Sparse to moderate	Rolling plains	Glacial drift	120 [~14]	Hansen, 1991
Soil-water model	Arkansas, White and Red River Basins	Southern Kansas	400-1015	1170-1625 [a]	Sparse to moderate	Rolling plain	Alluvial	83 [~12]	Hansen, 1991
Soil-water model based on Richard's Equation	Missouri River Basin	Akron, Colorado	No report	No report	None	High plains	Fallow soil	61	Krishnamurthi et al., 1977; Longenbaugh, 1975
Soil-water model (HELP)	Rio Grande Basin	Sunland Park, New Mexico	200	1270 [a]	Sparse (creosote bush)	Gently sloping escarpment	Sandy alluvium	0.066 [0.033]	Stephens & Coons, 1994
Streamflow measurements [6]	Ohio River Basin	Allegheny Subbasin	1067	686 [b]	Abundant	River valley	Alluvium, glacial outwash	190 [18]	Bloyd, 1974
Streamflow measurements	New England	Bellamy, Cochero, and Salmon Falls River Basins	1092	457-610	Abundant	Low mountain and coastal	Stratified drift	508 [46]	Mack and Lawlor, 1992

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions: MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type	Estimated Annual Recharge (mm/yr [%P])	Reference
Streamflow measurements [6]	Upper Mississippi River Basin	Big Muddy Subbasin	1067	787 [a]	Abundant	River valley	Alluvium, glacial outwash	32 [3]	Bloyd, 1975
Streamflow measurements [7]	Tennessee River Basin	Blue Ridge	1020-1780	610-1070 [a]	Abundant	Mountainous	Fractured noncarbonate with alluvium	215-240 [17-20]	Zurawski, 1978
Streamflow measurements [6]	Upper Mississippi River Basin	Cannon-Zumbro-Root Subbasins	737	635 [a]	Abundant	River valley	Alluvium, glacial outwash	74 [10]	Bloyd, 1975
Streamflow measurements [7]	Tennessee River Basin	Central Basin	1270	760 [a]	Abundant	Interior low plateau	Carbonate	255-380 [20-31]	Zurawski, 1978
Streamflow measurements [6]	Upper Mississippi River Basin	Chippewa Black Subbasin	762	610 [a]	Abundant	River valley	Alluvium, glacial outwash	90 [12]	Bloyd, 1975
Streamflow measurements [7]	Tennessee River Basin	Coastal Plain/Big Sandy River	1270	760 [a]	Abundant	River valley and rolling plain	Unconsolidated sand	215-240 [17-20]	Zurawski, 1978
Streamflow measurements [6]	Ohio River Basin	Cumberland Subbasin	1245	914 [b]	Abundant	River valley	Alluvium, glacial outwash	125 [10]	Bloyd, 1974
Streamflow measurements [7]	Tennessee River Basin	Cumberland Plateau	1270-1520	760-910 [a]	Abundant	Level highland	Fractured noncarbonate	230-355 [10-31]	Zurawski, 1978
Streamflow measurements [6]	Upper Mississippi River Basin	Des Moines Subbasin	787	686 [a]	Abundant	River valley	Alluvium, glacial outwash	24 [3]	Bloyd, 1975
Streamflow measurements [6]	Upper Mississippi River Basin	Fox-Wyaconda-Fabius Subbasins	889	737 [a]	Abundant	River valley	Alluvium, glacial outwash	18 [2]	Bloyd, 1975

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP[1]	E[2]	Vegetation	Topography	Soil/Aquifer Type		
Streamflow measurements[7]	Tennessee River Basin	Highland Rim	1270	760[a]	Abundant	Rolling plateau	Carbonate with alluvium	290-405 [23-33]	Zurawski, 1978
Streamflow measurements[6]	Upper Mississippi River Basin	Illinois Subbasin	864	724[a]	Abundant	River valley	Alluvium, glacial outwash	50 [6]	Bloyd, 1975
Streamflow measurements[6]	Upper Mississippi River Basin	Iowa-Cedar Subbasins	813	686[a]	Abundant	River valley	Alluvium, glacial outwash	50 [6]	Bloyd, 1975
Streamflow measurements[6]	Ohio River Basin	Kanawha Subbasins	1130	813[b]	Abundant	River valley	Alluvium, glacial outwash	180 [16]	Bloyd, 1974
Streamflow measurements[6]	Upper Mississippi River Basin	Kaskaskia Subbasin	965	775[a]	Abundant	River valley	Alluvium, glacial outwash	60 [6]	Bloyd, 1975
Streamflow measurements[6]	Ohio River Basin	Licking-Kentucky Subbasins	1143	889[b]	Abundant	River valley	Alluvium, glacial outwash	60 [5]	Bloyd, 1974
Streamflow measurements	New England	Lower Merrimale and Coastal River Basins	No report	No report	Abundant	Low mountain and coastal	Stratified drift	483	Stekl and Flanagan, 1992
Streamflow measurements[6]	Ohio River Basin	Lower Ohio River Subbasin	1220	914[b]	Abundant	River valley	Alluvium, glacial outwash	110 [9]	Bloyd, 1974
Streamflow measurements[6]	Upper Mississippi River Basin	Merames Subbasin	1016	787[a]	Abundant	River valley	Alluvium, glacial outwash	90 [9]	Bloyd, 1975
Streamflow measurements[6]	Ohio River Basin	Miami Subbasins	978	838[b]	Abundant	River valley	Alluvium, glacial outwash	98 [10]	Bloyd, 1974

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Streamflow measurements [6]	Upper Mississippi River Basin	Minnesota Subbasin	610	635 [a]	Abundant	River valley	Alluvium, glacial outwash	18 [3]	Bloyd, 1975
Streamflow measurements [6]	Upper Mississippi River Basin	Mississippi Headwaters	660	610 [a]	Abundant	River valley	Alluvium, glacial outwash	90 [14]	Bloyd, 1975
Streamflow measurements [6]	Ohio River Basin	Monongahela Subbasin	1194	737 [b]	Abundant	River valley	Alluvium, glacial outwash	215 [18]	Bloyd, 1974
Streamflow measurements [6]	Ohio River Basin	Muskingum Subbasin	1003	826 [b]	Abundant	River valley	Alluvium, glacial outwash	90 [9]	Bloyd, 1974
Streamflow measurements [8]	South Atlantic Gulf Region	NA	1120-1630	No report	Abundant	Coastal plain and rolling highlands	Unconsolidated and consolidated sediments	150 [10.9]	Cederstrom et al., 1979
Streamflow measurements [6]	Upper Mississippi River Basin	Rock Subbasin	838	673 [a]	Abundant	River valley	Alluvium, glacial outwash	110 [13]	Bloyd, 1975
Streamflow measurements [6]	Upper Mississippi River Basin	Salt Subbasin	940	762 [a]	Abundant	River valley	Alluvium, glacial outwash	19 [2]	Bloyd, 1975
Streamflow measurements [6]	Ohio River Basin	Sandy-Guyandotte Subbasins	1118	838 [b]	Abundant	River valley	Alluvium, glacial outwash	90 [8]	Bloyd, 1974
Streamflow measurements [6]	Ohio River Basin	Scioto Subbasin	1003	864 [b]	Abundant	River valley	Alluvium, glacial outwash	30 [3]	Bloyd, 1974
Streamflow measurements [6]	Upper Mississippi River Basin	Skunk Subbasin	813	686 [a]	Abundant	River valley	Alluvium, glacial outwash	50 [6]	Bloyd, 1975

Table A-2. Estimates of diffuse annual recharge by estimation technique

Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Streamflow measurements [6]	Ohio River Basin	Upper Ohio River Subbasin	1016	788 [b]	Abundant	River valley	Alluvium, glacial outwash	100 [10]	Bloyd, 1974
Streamflow measurements [7]	Tennessee River Basin	Valley and Ridge	1020-1520	610-910 [a]	Abundant	Parallel valleys and ridges	Carbonate with alluvium	140-240 [13-23]	Zurawski, 1978
Streamflow measurements [6]	Ohio River Basin	Wabash Subbasin	1016	864 [b]	Abundant	River valley	Alluvium, glacial outwash	111 [11]	Bloyd, 1974
Streamflow measurements [6]	Upper Mississippi River Basin	Wapsipinicon Subbasin	813	660 [a]	Abundant	River valley	Alluvium, glacial outwash	80 [10]	Bloyd, 1975
Streamflow measurements [6]	Ohio River Basin	White Subbasin	1067	838 [b]	Abundant	River valley	Alluvium, glacial outwash	107 [10]	Bloyd, 1974
Streamflow measurements [6]	Upper Mississippi River Basin	Wisconsin Subbasin	813	610 [a]	Abundant	River valley	Alluvium, glacial outwash	120 [15]	Bloyd, 1975
Streamflow measurements and soil-water balance	Tennessee River Basin	Blue Ridge	No report	No report	Abundant	Mountainous	Fractured crystalline bedrock	300 [~21] [c]	Hoos, 1990
Streamflow measurements and soil-water balance	Tennessee River Basin	Central Basin, Tennessee	No report	No report	Abundant	Gently rolling to hilly and steep	Flat-lying carbonate	140 [~11] [c]	Hoos, 1990
Streamflow measurements and soil-water balance	Missouri River Basin	Cheyenne, Wyoming	381	No report	Grasslands	Rolling plains	Sand and gravel	22 [5.8]	Morgan, 1946

Table A-2. Estimates of diffuse annual recharge by estimation technique
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Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E [2]	Vegetation	Topography	Soil/Aquifer Type		
Streamflow measurements and soil-water balance	Tennessee River Basin	Cumberland Plateau, Tennessee	No report	No report	Abundant	Gently rolling to hilly	Flat-lying sandstone, shale and conglomerate	165 [~12] [c]	Hoos, 1990
Streamflow measurements and soil-water balance	Tennessee River Basin	Highland Rim	No report	No report	Abundant	Undulating to hilly and steep	Flat-lying carbonate	190 [~15] [c]	Hoos, 1990
Streamflow measurements and soil-water balance	Tennessee River Basin	Valley and Ridge, Tennessee	No report	No report	Abundant	Parallel valleys and ridges	Folded/faulted sandstone, carbonate and shale	170 [~12] [c]	Hoos, 1990
Tritium	California Region	Golden Gate Park, San Francisco, California	503	890 [a]	Abundant/ varied	Hilly	Unconsolidated sands and silty sands	127 [25]	Phillips et al., 1993
Tritium in aquifer	Rio Grande Basin	Roswell Basin	305-373	No report	Sparse	Mountain slope to river	Limestone	[3.3-7.8]	Gross et al., 1979
Tritium peak	Rio Grande Basin	El Paso, Hudspeth County, Texas	280	~1960 [b]	Sparse	Flat alluvial plain	Clay, silt, sand	7 [2.5]	Scanlon, 1992
Tritium peak	Rio Grande Basin	Las Cruces, New Mexico	230	2390 [b]	Sparse (grass and shrubs)	Terrace/ pediment	Sandy loam	9.5 [4]	Phillips et al., 1988
Tritium peak	Rio Grande Basin	Socorro, New Mexico	200	1780 [b]	Sparse (four-winged saltbrush, creosote)	Flat-lying floodplain terrace	Fine sand, sandy loam	8.4 [4.2]	Phillips et al., 1988

Table A-2. Estimates of diffuse annual recharge by estimation technique

Estimation Technique	Geographic Region	Site	Site Hydrologic Conditions					Estimated Annual Recharge (mm/yr [%P])	Reference
			MAP [1]	E^2	Vegetation	Topography	Soil/Aquifer Type		
Water level fluctuations	Texas-Gulf Region	Ogallala Aquifer in Lea County, New Mexico	~355	No report	Sparse	High plains	Sand and gravel	6.4-10.7 [1.8-3.0]	Theis, 1937
Water level fluctuation	Ohio River Basin	Wright-Patterson AFB	965	No report	Abundant	Central lowlands	Glacial sand and gravels	315-401 [33-42]	Dumouchelle et al., 1993
Hybrid water-level fluctuation	Arkansas, White and Red River Basins	Great Bend Prairie, Arkansas Basin, central Kansas	585	47-60 [d]	Sparse	Low relief sand-dune prairie	Alluvial	56 [10]	Sophocleous, 1992

[1] Mean annual precipitation (mm/yr)
[2] Evaporation (mm/yr)
[3] Where basin outflow is limited to known pumping discharge
[4] Recharge, approximated by steady-state yield, may be overestimated
[5] Soil-water balance computations based on data provided for mean annual precipitation, evapotranspiration and runoff
[6] Assumes a 60-percent flow parameter equals baseflow
[7] Range of recharge estimated from average baseflow
[8] Average recharge estimated from average baseflow
[a] Potential evapotranspiration
[b] Lake evaporation
[c] Report of annual recharge estimate as 0% based on MAP reported by Zurawski (1978)
[d] Complimentary relationship areal evapotranspiration (CRAE) (Morton, 1983)

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