

Estimation of Infiltration and Recharge for Environmental Site Assessment

Health and Environmental Sciences Department
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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

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 estimation 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

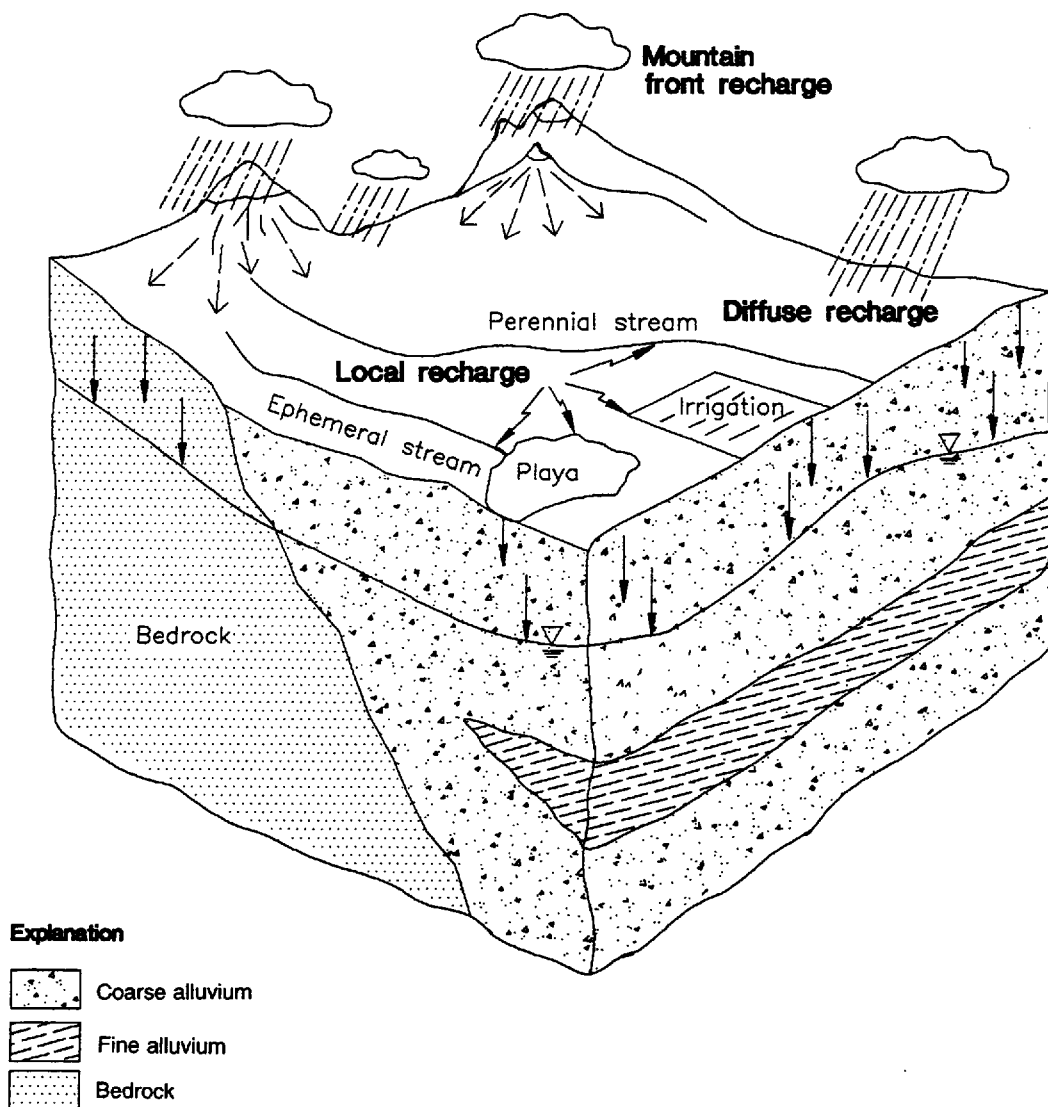


Section 2

RECHARGE CONCEPTS AND TERMINOLOGY

Recharge is simply the addition of water to an aquifer. Natural recharge to ground-water 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.

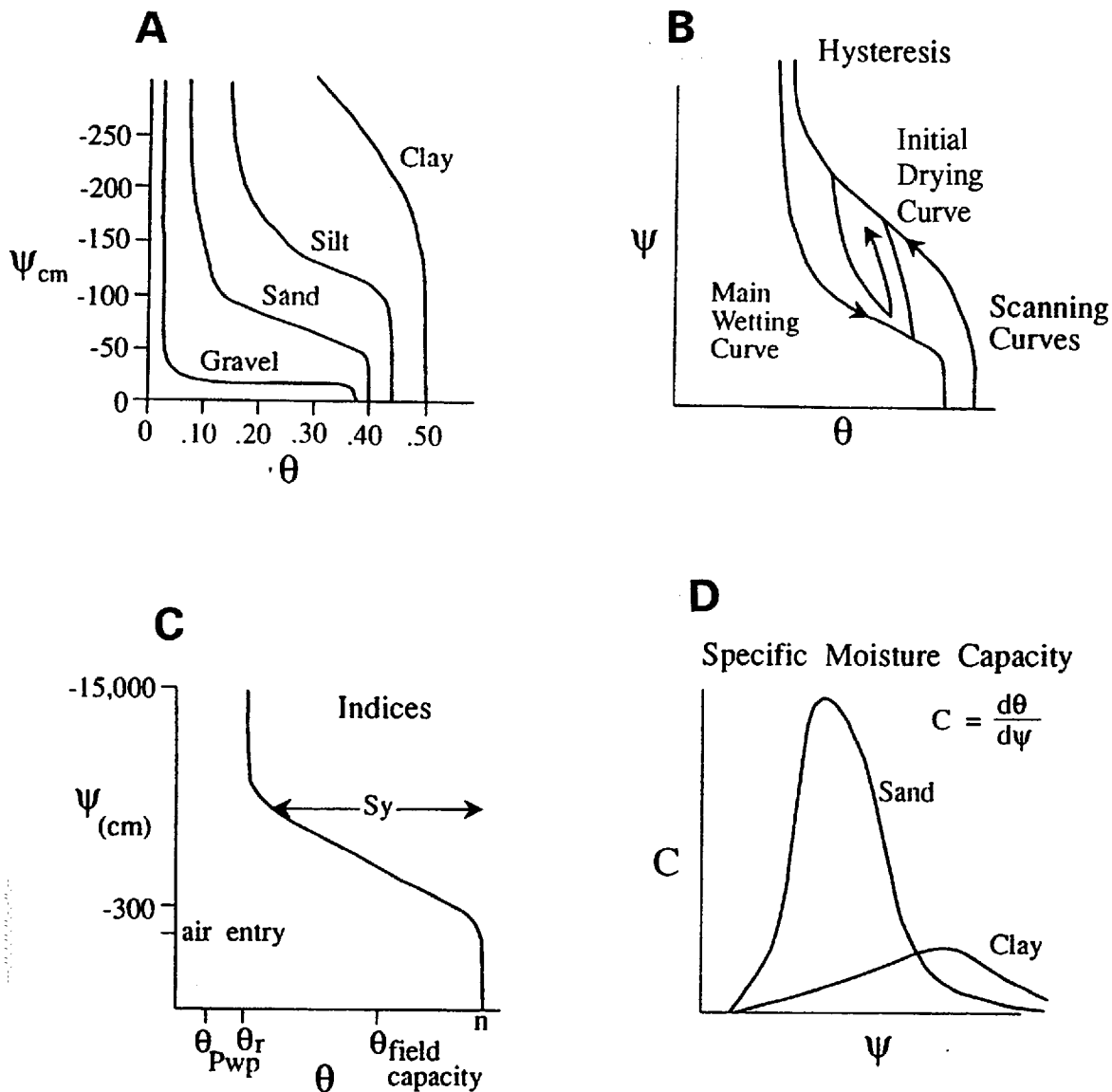


Figure 2-2. Soil-water retention: Pressure head, ψ , versus volumetric water content, θ , (A) for different soil textures, (B) showing the effect of hysteresis on a specific soil, (C) illustrating index parameters, and (D) expressed as specific moisture capacity.

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 saturated 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

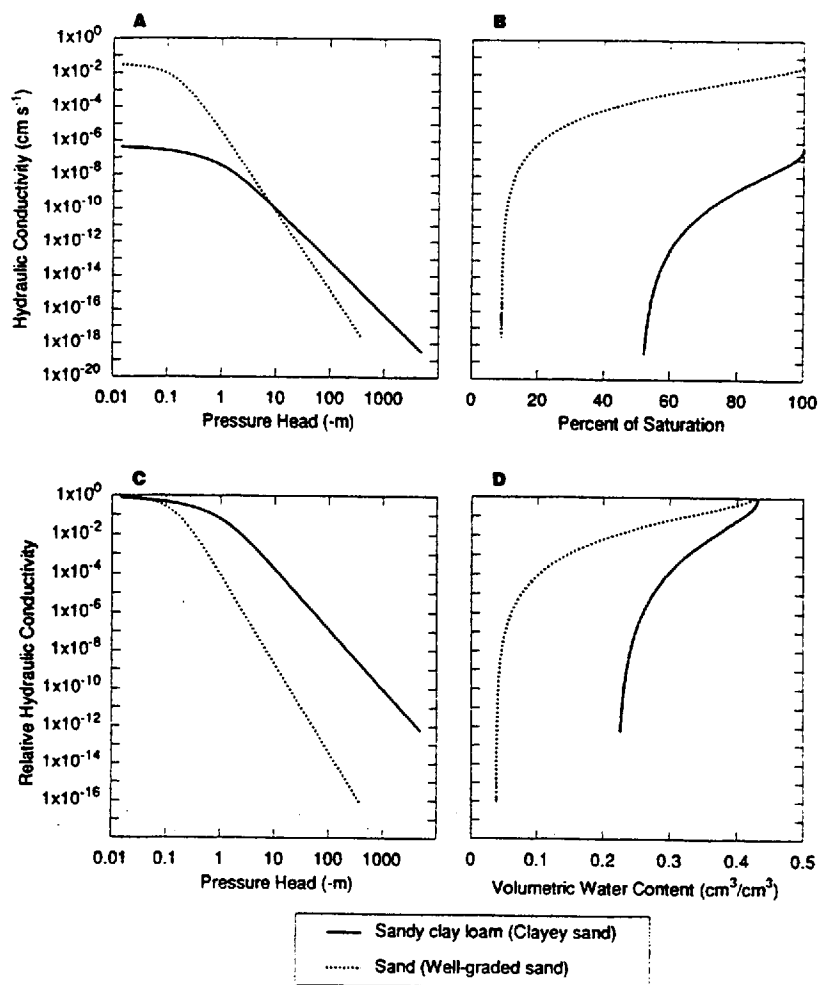


Figure 2-3. Unsaturated hydraulic properties of two soils

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.

Infiltration, Redistribution and Recharge

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.

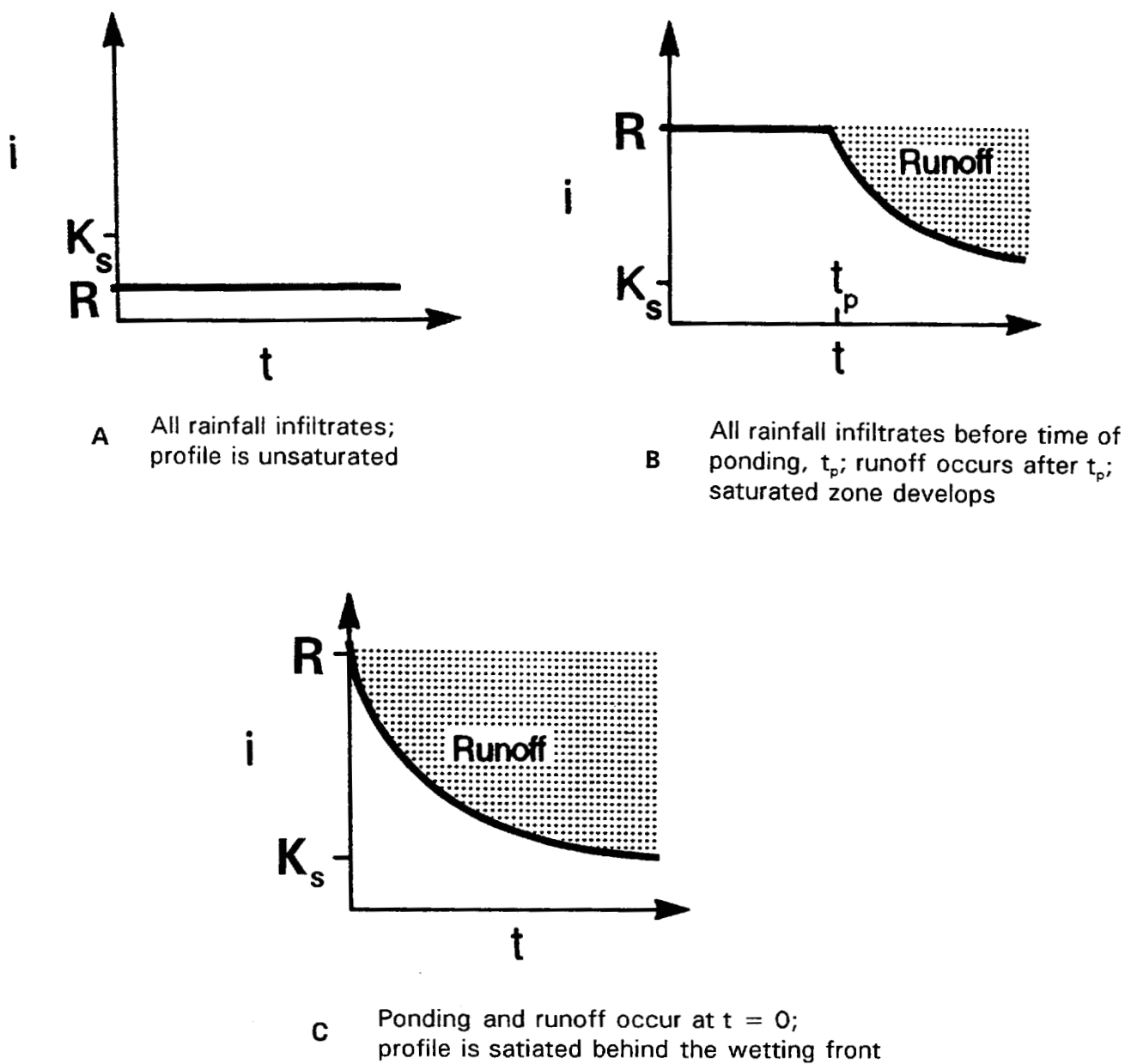


Figure 2-4. Infiltration rate (i) dependence on rain intensity, R

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.

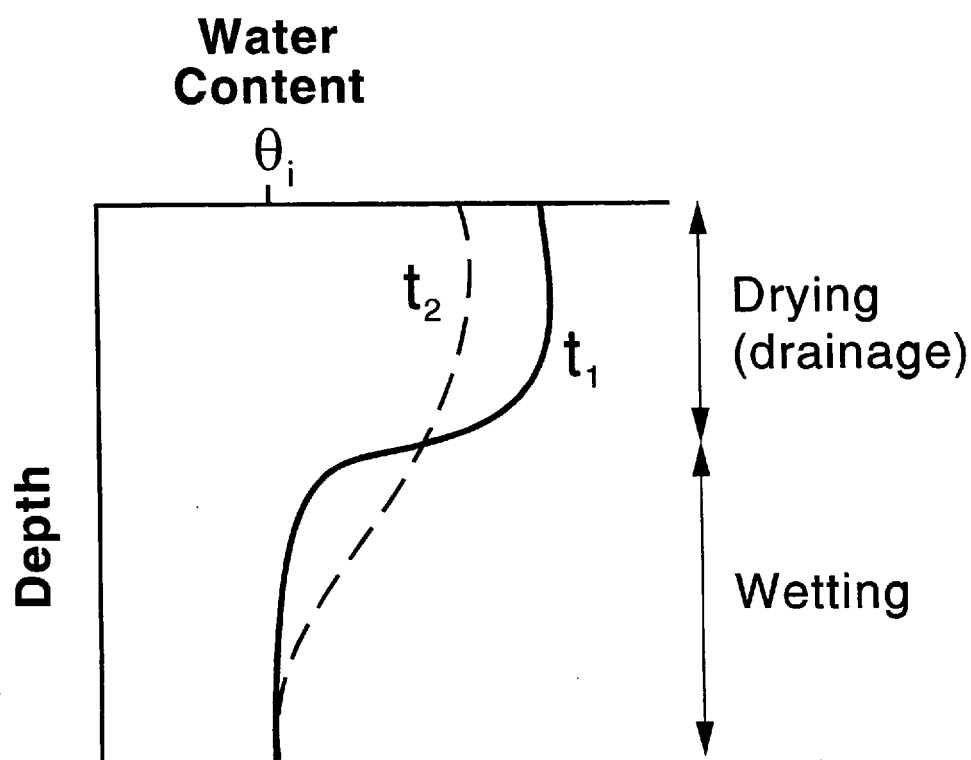


Figure 2-5. Water content profiles during infiltration (time t_1) and redistribution (time t_2). θ_i is the initial water content.

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:

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

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 :

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

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.

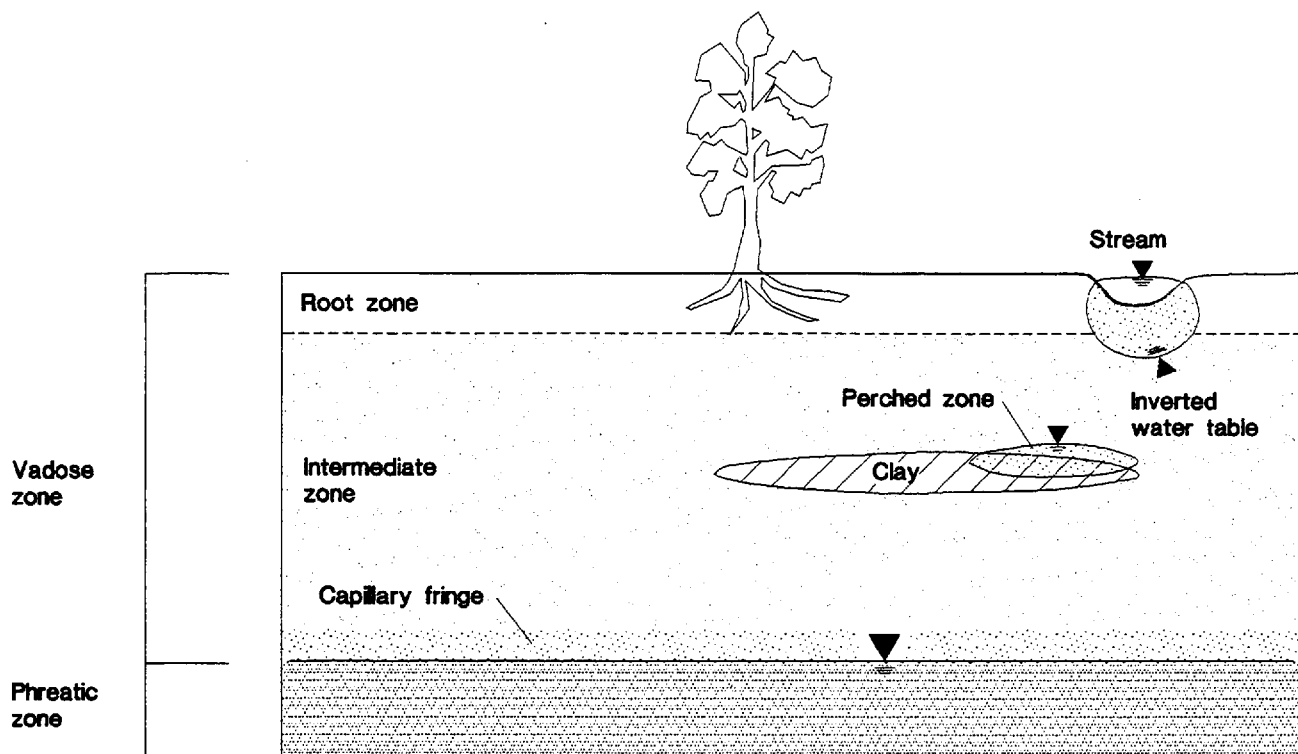


Figure 2-6. Conceptual model of the vadose zone

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.

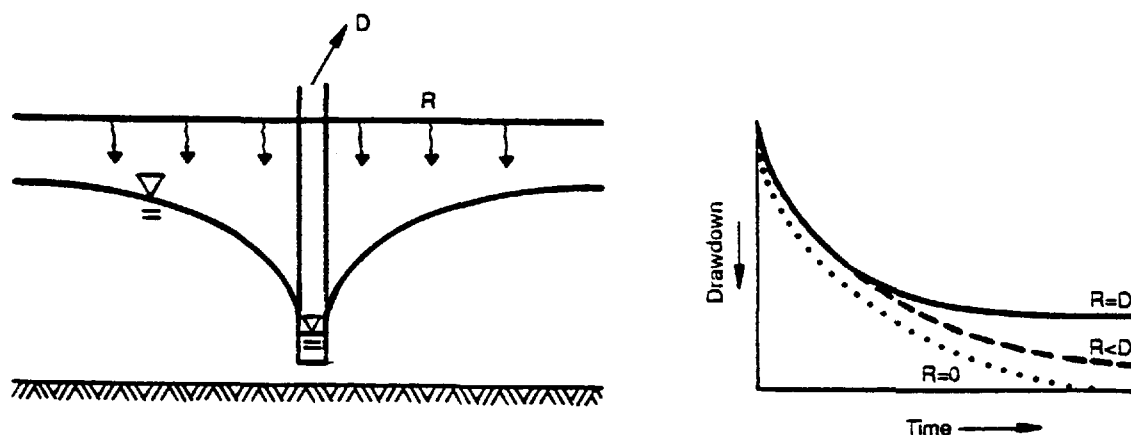
Flow through the vadose zone during an infiltration event may occur under both saturated and unsaturated conditions. Saturated flow is more likely in humid climates, with moderate- to low-permeability soils and shallow water table conditions, following a series of storms of high intensity or prolonged 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:

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

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_{is}	Multidimensional K_{is}	Multidimensional K_{is} and $K(\psi)$, sorptivity, pore geometry	Vertical K_{is} , $K(\psi)$, $K(\theta)$
Range of Use	$K_{is} = 10^{-2}$ to 10^{-6} cm/s	$K_{is} = 10^{-1}$ to 10^{-9} cm/s	$K_{is} = 10^{-1}$ to 10^{-8} cm/s	$K_{is} = 10^{-2}$ to 10^{-6} cm/s; tensions up to 15 cm of water	$K_{is} = 10^{-2}$ to 10^{-6} cm/s
Time to Complete	<4 h at $K_{is} = 10^{-5}$ cm/s	<4 h at $K_{is} = 10^{-5}$ cm/s	<4 h at $K_{is} = 10^{-5}$ 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_{is} 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 $\pm 1\%$ by volume.	Moderate; typically $\pm 2\%$ 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.