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# Vadose Water

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## Introduction

The term vadose is derived from the Latin *vadosus*, meaning 'shallow.' In this sense, however, it refers to shallow depths beneath the land surface, not shallow portions of surface water bodies. The vadose zone is frequently called the unsaturated zone, and sometimes the zone of aeration, as its pore space usually contains air as well as water. The vadose zone extends from the land surface to the water table (the lowest water table if there is more than one).

Prediction of the transport rates of water and other substances within the vadose zone is critical to infiltration, runoff, erosion, plant growth, microbiota, contaminant transport, aquifer recharge, and discharge to surface water. Vadose-zone flow is fundamentally complicated by nonlinearity and hysteresis of unsaturated hydraulic properties, and extreme sensitivity to materials and hydraulic conditions. There is much variety in its natural constituents: soils, rocks, water, air, plants, animals, and microorganisms. Modern hydrology must consider interactions not only among these constituents themselves, but also with a wide variety of contaminants, including pesticides, fertilizers, irrigation wastewater, manure, sewage, toxic chemicals, radioactive substances, bacteria, mine wastes, and organic liquids.

The porous medium of the vadose zone is typically soil (Figure 1), but may also be porous rock, or any other material that occurs near the earth's surface. In general, the forces of molecular attraction are greater between solid and water than between solid and air. Consequently, water behaves as the wetting phase, and air as the nonwetting phase. Within the pores, water tends to cling to solid surfaces in films and in partially filled pores with curved air–water interfaces (Figure 2).

## Fundamental Processes of Vadose Water

### Unsaturated Hydrostatics

**Water content** The most basic measure of the water is volumetric water content, often symbolized $\theta$, defined as the volume of water per bulk volume of the medium.

The most standard measurement of $\theta$ is the gravimetric method. The procedure is to dry a sample of the porous medium in an oven until the weight is constant, and then calculate how much water was in the soil from the difference between the dry weight and the initial wet weight. Other methods in widespread use have advantages, such as being less disruptive. Neutron scattering is commonly used for monitoring $\theta$ as a function of depth in the field. This method is based on the high effectiveness of water, among the various components of the wet soil, in slowing neutrons. Commercially available equipment has a neutron source and detector housed in a cylindrical probe that can be lowered to various depths in a lined hole. Another way to monitor $\theta$ in the lab or field is by measurement of the dielectric constant of the medium, usually by time-domain reflectometry (TDR). For most applications, TDR electrodes in the form of metal rods are inserted into the soil. Liquid water has a much higher dielectric constant than do other vadose zone constituents, so a measurement of this property can indicate the amount of water present within the volume sensed. A less common method is to measure electrical conductivity, which increases with $\theta$. This principle can be applied tomographically for observing two- or three-dimensional details of changing water distributions in the field. The reflective or absorptive behavior of ground penetrating radar can also be used directly or tomographically to indicate water content distributions.

**Water pressure and energy** Matric pressure, usually symbolized $\psi$, may be thought of as the pressure of the water in a pore of the medium relative to the pressure of the air, in other words, the pressure difference across an air–water interface. The liquid–solid attraction that curves the air–water interfaces also causes capillary rise (Figure 3). Water that has risen to higher levels in the tube, because of the attraction of the tube walls for water, is at a lower pressure than the bulk water outside the tube. The narrower the tube, the stronger is the capillary effect. Similarly, in an unsaturated porous medium, water is generally at lower pressure than the air, so $\psi$ is negative, the concave side of air–water interfaces is toward the air.

The most direct measurement of $\psi$ is by a tensiometer. In firm contact with the porous medium, this device allows for equilibration of pressure between the water in unsaturated pores and the water in a


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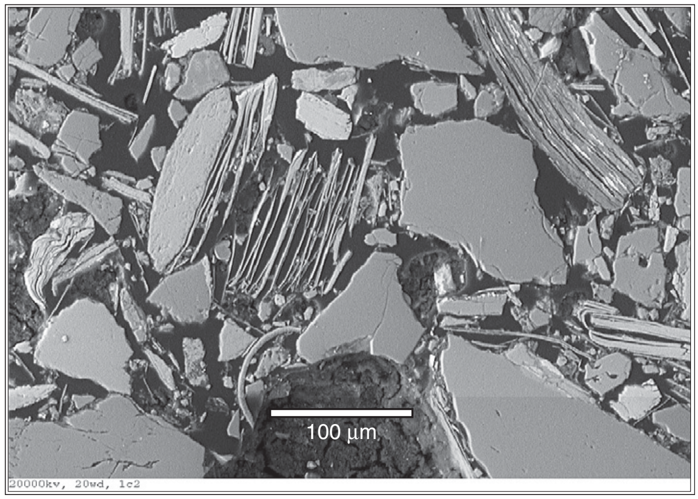


**Figure 1** Particles and pores of a silt loam soil, scanning electron micrograph of a thin section. Reproduced from Lebron I, Schaap MG and Suarez DL (1999) Saturated hydraulic conductivity prediction from microscopic pore geometry measurements and neural network analysis. *Water Resources Research* 35: 3149–3158, with permission from the American Geophysical Union (http://www.agu.org/pubs/copyright.html).

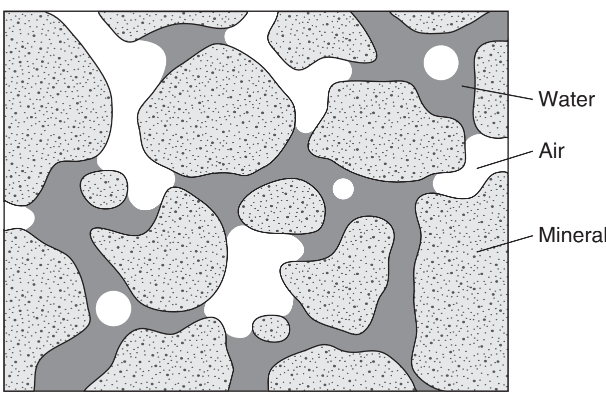


**Figure 2** Pore space of a hypothetical unsaturated medium, illustrating the arrangement of solid, liquid, and gaseous phases. Few intergrain contacts appear in these figures because such contacts are essentially points in three-dimensional space, and mostly do not lie in the two-dimensional plane of the image.

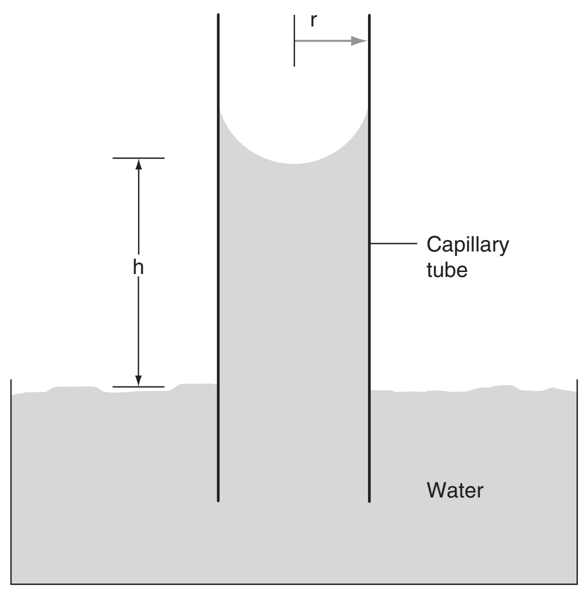


**Figure 3** Capillary rise in a tube.

small chamber where a gauge or transducer reads the pressure (Figure 4). Its use is limited to relatively wet soils.

Other methods are available for relatively dry media and for easier application when less accuracy is acceptable. Some of these are based on the humidity of the air in soil pores. A low (strongly negative) $\psi$ increases the pore water's effectiveness for absorbing water molecules out of the vapor in the soil air, resulting in a lower relative humidity. The effect is slight, however; a 0 to $-15$ atm range in $\psi$ corresponds to a 100 to 99% range in relative humidity. Another class of methods uses an intermediary porous medium of known retention properties, typically gypsum blocks, nylon fabric, or filter paper. This medium is placed in contact with the medium to be measured so that $\psi$ becomes equal in both. Then the water content of the intermediary medium is measured by other means (usually electrical conductivity, thermal diffusivity, or mass) and translated into a matric pressure using the known properties.

**Water retention** Analogously to capillary rise, the smaller pores of a medium hold water more strongly than the larger pores do. To extract water from a

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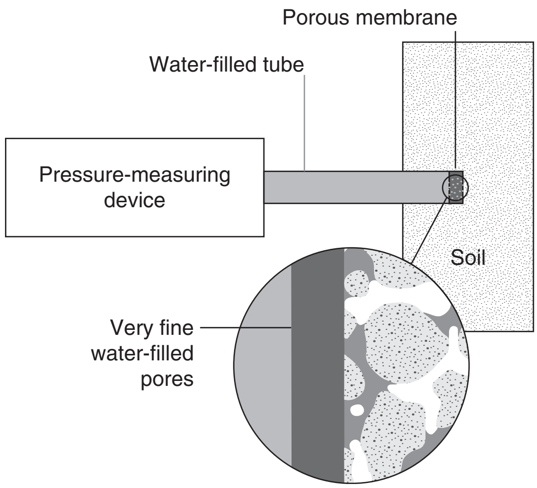


**Figure 4** A tensiometer establishes a continuous water phase from the soil pores to a chamber for pressure is measured to indicate matric pressure.

small pore requires application of a more highly negative matric pressure. The volume of water held in the soil at different matric pressures therefore depends on the pore-size distribution, and is a characteristic of a particular porous material. This property, known as soil–water retention, is expressible as a set of $\theta$ vs. $\psi$ curves for a given medium. When $\psi$ is close to zero, air–water interfaces are broadly curved, nearly all pores are filled, and $\theta$ is high. If $\psi$ is much less than zero, the interfaces are more tightly curved, they can no longer span the largest pores, and the pores have less water in them. Thus, greater $\theta$ goes with greater (less strongly negative) $\psi$.

Normally, during the process of wetting a porous medium, some air is trapped as the pores surrounding it become water-filled. In soil brought to $\psi = 0$, it is common for trapped air to occupy one-tenth or more of the total pore space. Trapped air will eventually dissolve and diffuse away, but vadose-zone moisture conditions commonly change rapidly enough that the pore space always contains some amount of air. Consequently, for most retention curves, when $\psi = 0$, $\theta$ has a value less than the total porosity.

In a granular medium, the particle-size distribution, or texture, relates in some way to the pore-size distribution. Larger particles may have larger pores between them. In addition to texture, the structure of the medium, especially as related to such features as aggregation, shrinkage cracks, and biologically generated holes, substantially influences the retention curve.

*Examples* Figure 5(a) shows a retention curve for a core sample of a silt loam soil from an apple orchard. Water retention is hysteretic; $\theta$ for a given

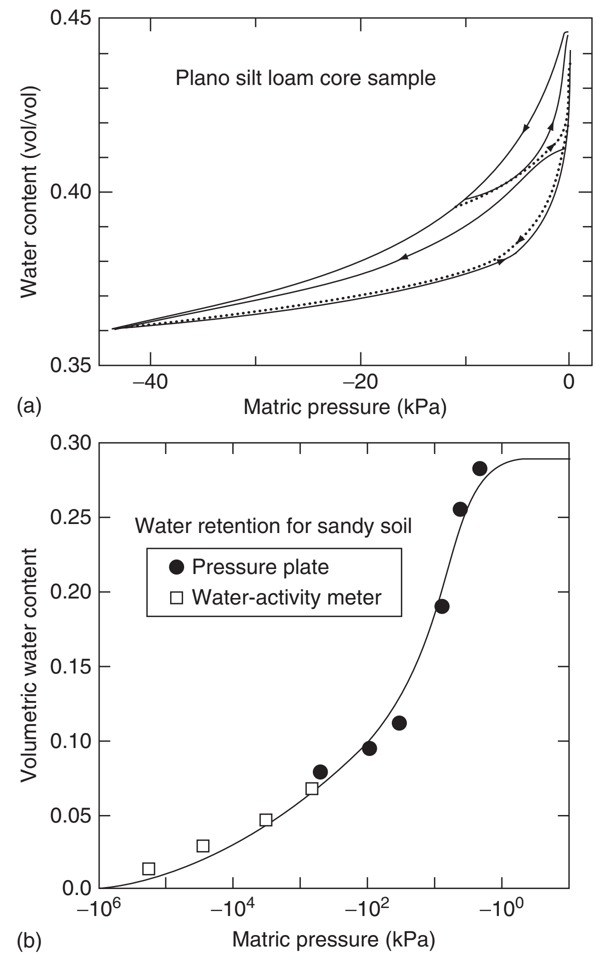


**Figure 5** Water retention relations. (a) Hysteretic water retention of a silt loam soil, with arrows indicating the direction of change on each curve. Reproduced from Nimmo JR and Miller EE (1986) The temperature dependence of isothermal moisture vs. potential characteristic of soils. *Soil Science Society of America Journal* 50: 1105–1113, with permission from the Soil Science Society of America. (b) Drying retention curve for a sandy soil from the Amargosa Desert Research Site. The points are measurements by two different methods and the smooth curve is a fit of the model of Rossi and Nimmo (1994). Reproduced from Andraski BJ (1996) Properties and variability of soil and trench fill at an arid waste-burial site. *Soil Science Society of America Journal* 60: 54–66, with permission from the Soil Science Society of America.

$\psi$ is different when measured for drying and wetting, and in general depends on the wetting/drying history of the medium. Thus, there is not a unique curve but a family of curves. The drying retention curve in Figure 5(b) is far from linear and covers five orders of magnitude in $\psi$. This enormous range requires multiple measurement techniques. In most cases investigators measure and plot only a single curve from the family of possible curves, usually a drying curve, and over only a portion of the range, usually at the wet end.

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Considering the drying of soil from saturation, $\theta$ in Figure 5(b) stays high until a particular $\psi$ value where it starts to decline. That $\psi$ is called the air-entry value. By the capillary hypothesis, it is assumed to have a nonzero value because the largest fully wet pore of the medium will stay filled until the air–water pressure difference exceeds in magnitude the equivalent $\psi$ value of capillary rise. In natural media, the air-entry value is usually poorly determined, as the decline in $\theta$ with $\psi$ starts gradually, beginning at $\psi$ nearly equal to zero. Artificial porous media, however, can be made in such a way that many pores are close to the size of the largest pore, so that air-entry is a sharp and sudden phenomenon.

*Practical significance* The water retention relation is important in quantifying soil moisture dynamics, as discussed later. Another area in which it is important is in soil–plant–water relations. Often termed matric potential because it is representative of the energy state of the soil water, $\psi$ indicates the work that must be done by plant roots to extract water from the unsaturated soil. Plants wilt from inadequate soil moisture not in direct response to low $\theta$, but rather because at low $\theta$, $\psi$ is low. In soil that is too dry, $\psi$ is so highly negative that a plant is incapable of overcoming this energy barrier. Typically, the minimum $\psi$ is about $-15$ atm for agricultural plants, though much lower in some plants, especially those native to arid regions.

*Measurement or estimation of water retention* Any system that makes independent simultaneous measurements of $\theta$ and $\psi$ can indicate the water retention relation. In addition, there are methods specifically intended to measure this property. Many of these methods use a porous membrane, often ceramic, to permit equilibration of water pressure between the porous medium on one side of the membrane with bulk water on the other, as in tensiometers (Figure 4). The pressure of this bulk water (and hence the pore-water pressure) is controlled, as is the air pressure in the medium, in order to control $\psi$. The pressure, or less commonly the volume of water, is adjusted through a planned sequence, and paired values of $\psi$ and $\theta$ (one of them controlled and the other measured) represent the retention curve.

Because various nonhydraulic properties of a medium, especially particle-size distribution, correlate in some way with water retention but are considerably easier to measure, property-transfer models have been developed for estimating water retention from other properties. One broad class of such models is based on theoretical relationships between pore sizes and particle sizes. Models of this type may work reasonably well for sandy media. Another class of such models uses statistically calibrated pedotransfer functions. The basis for this type of model is not a principle like the correlation of pore and particle size, but rather a database of measured water retention and other properties for a wide variety of media. Given a medium's particle-size distribution and other properties such as organic matter content, a pedotransfer function can estimate a retention curve with good statistical comparability to retention curves of other media in the database whose nonhydraulic properties may be similar. Whatever the choice of model, however, without any retention measurements for the medium in question, it is usually impossible to know whether the model result is a good representation of the retention curve.

*Empirical formulas for water retention* In general, a water retention curve can be represented by measured data, interpolated as needed. It is often convenient, however, to express the curve as a parametric empirical formula. Among the most widely used empirical formulas are that of Brooks and Corey

$$\theta = (\theta_{\max} - \theta_{\min})\left[\frac{\psi}{\psi_b}\right]^b + \theta_{\min} \qquad [1]$$

where $\psi_b$, $b$, and $\theta_{\min}$ are fitted empirical parameters and $\theta_{\max}$ is the maximum value of $\theta$, and that of van Genuchten

$$\theta = (\theta_{\max} - \theta_{\min})\left[\frac{1}{1+\left(\frac{\psi}{\psi_c}\right)^{\nu}}\right]^{\mu} + \theta_{\min} \qquad [2]$$

where $\psi_c$, $\nu$, $\mu$, and $\theta_{\min}$ are fitted empirical parameters. Fundamentally, $\theta_{\min}$ should equal zero, but a finite value is often used to improve the fit in the higher-$\theta$ portion of the curve. Equations as simplified as these cannot represent the precise form of the $\theta$–$\psi$ relation of a natural medium, though they may serve for various practical purposes.

### Diffuse Unsaturated Flow

Traditionally, unsaturated flow is considered as a continuum in which the average behavior of water in many pores within a compact region of space (a representative elementary volume (REV)) represents the characteristics of the medium point-by-point. In general, this leads to a conceptualization of moisture varying systematically throughout the medium (Figure 6).

In conventional unsaturated flow theory, two types of factors determine water flux: driving forces (chiefly gravity and matric pressure gradients) and properties of the medium. The matric forces sometimes greatly

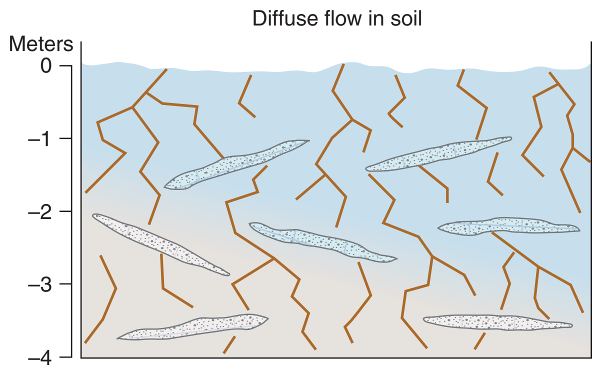


**Figure 6** A diffuse distribution of water in soil. Blue shading indicates wetter soil over most of the upper half of this profile, especially to the right. Drier soil is lower and to the left. The transition between wet and dry soil is gradual and spread over a considerable volume.

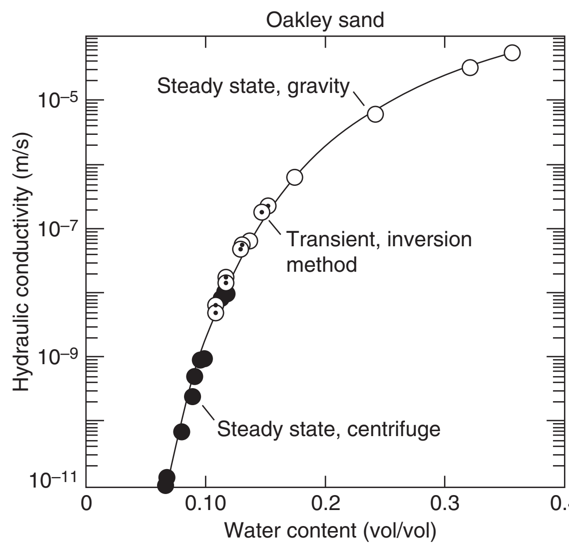


**Figure 7** Hydraulic conductivity for a sandy soil (Oakley sand), measured by three methods.

exceed the gravitational force. Other forces may also drive flow under some conditions, as when temperature or osmotic gradients are significant.

**Darcy's law for vadose water** Unsaturated flow has its basic mathematical expression in Darcy's law, in a form such as

$$q = -\frac{K(\theta)}{\rho g}\left[\frac{d\psi}{dz} + \rho g\right] \quad [3]$$

where $q$ is the flux density, $K$ is the unsaturated hydraulic conductivity, $\rho$ is the density of water, $g$ is the acceleration of gravity, and $z$ is upward distance. The conversion factor $1/\rho g$ is shown here explicitly so that this expression can be used directly with $\psi$ in SI pressure units (kPa), and $K$ in velocity units (m/s). In head units, $\psi$ takes dimensions of length.

**Unsteady diffuse flow** In the general case of transient (nonsteady) unsaturated flow, the flow itself causes $\theta$ to change throughout the medium, which leads to continuously changing hydraulic conductivity and driving forces. These interacting processes can be accommodated mathematically by combining the equation of continuity

$$\frac{\partial\theta}{\partial t} = -\frac{\partial q}{\partial z} \quad [4]$$

with Darcy's law [3] to get Richards' equation, which for one-dimensional vertical flow within a medium in earth gravity can be written as

$$C\frac{\partial\psi}{\partial t} = \frac{1}{\rho g}\frac{\partial}{\partial z}\left[K\frac{\partial\psi}{\partial z}\right] + \frac{\partial K}{\partial z} \quad [5]$$

where $C$ is the differential water capacity, a property of the medium defined as $d\theta/d\psi$. It is also possible to formulate this equation in terms of $\theta$ rather than $\psi$. In general the equation can be solved numerically.

**Unsaturated hydraulic conductivity** $K$ of the medium depends on the whole set of filled pores, especially on their size, shape, and connectedness. In unsaturated media, as illustrated by the measurements in Figure 7, $K$ depends very strongly on $\theta$. Because the large pores empty first as $\theta$ decreases, the result is not only that fewer pores are filled to conduct water, but the remaining filled pores are smaller and therefore less conductive. With fewer pores filled, the paths of water flowing though the medium also become more tortuous. When the soil is quite dry, very few pores are filled, and the water moves mainly through poorly conducting films adhering to particle surfaces. The net effect of these factors is to reduce hydraulic conductivity by several orders of magnitude as the soil goes from saturation to typical field-dry conditions.

*Measurement or estimation of unsaturated K* The most accurate measurements of hydraulic conductivity are by steady-state methods. One technique is to establish constant (though not necessarily equal) pressures of water at two opposing faces of a porous medium, measure the flux density, and calculate $K$ using Darcy's law. Another is to force water through the medium at a constant and known flux density, which lets $\psi$ become uniform in part of the sample, then to compute $K$ from the known flux density and force of gravity. With gravity as the main driving force, steady-state measurements are possible only for the high $K$ values of fairly wet soil. Centrifugal

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force makes possible the accurate measurement of $K$ at low $\theta$.

Many techniques for measuring unsaturated hydraulic conductivity use unsteady flow. One of these is the instantaneous-profile method, useful in both laboratory and field. This method uses measurements of $\theta$ and $\psi$ within a medium in which unsteady flow has been established, so that both the flux density and the $\psi$ gradient can be computed at one or more instants of time. Another alternative for laboratory applications uses flow driven by evaporation. There are various indirect and inverse methods – a wide variety of situations where data are available describing water flow over time can provide information for an estimation of $K$. The tension infiltrometer method is in widespread use for field applications. This method uses the measured infiltration rate as a function of time for water applied at controlled $\psi$ values to calculate the unsaturated hydraulic properties. It is often implemented as an inverse method.

Property-transfer models can be useful for estimating $K$. Usually these use water retention, not particle-size distribution, as the more easily measured type of data from which unsaturated $K$ is calculated. If a transfer from particle size to $K$ is needed, such a model may be combined with a water retention property-transfer model, though reliability is likely to be reduced because the particle-size distribution is less directly related to $K$.

Capillary theory provides an interpretation of the pores in the medium that relates to both retention and $K$. Models developed by Mualem and Burdine have become widely used for this purpose. A direct combination of an empirical formula for water retention, such as [1] or [2], into a capillary-theory formulation of unsaturated $K$ can yield a convenient analytical formula for $K(\theta)$, and facilitate the combined treatment of water retention and unsaturated $K$.

*Empirical formulas for unsaturated K* As in the case of water retention, completely empirical formulas can represent unsaturated $K$. Gardner, for example, used

$$K(\theta) = A \exp(-\alpha\theta) \quad [6]$$

where $A$ and $\alpha$ are fitted empirical parameters. Such formulas have greater simplicity and sometimes lead to more realistic curve shapes than formulas developed for combined representation of $K$ and water retention. The $\alpha$ parameter in [6] is used in developing and applying other models, such as analytical solutions of equations representing unsaturated flow.

**Effects of dissimilar materials** Layers that contrast in hydraulic properties impede vertical flow by various mechanisms. When water moves down from a coarse to a fine layer, as from coarse sand to silt, if both layers are near saturation, the fine layer has smaller hydraulic conductivity; therefore, flow slows when it reaches the fine layer. If, however, the coarse layer is nearly saturated but the fine layer is initially fairly dry, at first the flow may be temporarily accelerated while the flow is dominated by the sorptive nature of the fine medium, which tends to suck water out of the coarse material. When water moves down from a fine to a coarse layer it will also be impeded under many circumstances. Dry coarse material has an extremely small hydraulic conductivity; thus it tends not to admit water into the pores and exhibits a somewhat self-perpetuating resistance to flow. Water breaks into the coarse layer if the pressure at the layer contact builds to the point that the water-entry pressure (the minimum water pressure needed to fill an empty pore) of some of the large pores is exceeded. This can generate flow instabilities. Stable or not, water flow into the pores of the coarse medium increases that medium's hydraulic conductivity. With equal $\psi$ values across the layer boundary, unsaturated $K$ of the coarse layer is often less than that of the fine layer. In general, stable or diffuse flow through layers where fine overlies coarse is slower than it would be if both layers had the properties of the fine medium.

### Preferential Flow

In recent decades it has become increasingly clear that much unsaturated-zone transport of importance, especially when water is abundant, occurs through a small fraction of the medium along preferential paths such as wormholes, fractures, fingers of enhanced wetness, and regions near contacts between dissimilar portions of the medium. This flow, for which accepted theory applies less well, occurs at rates typically some orders of magnitude faster than flow through the remainder of the medium. In many applications, its importance is redoubled because preferentially transported substances are exposed to only a small fraction of the soil or rock and only for limited time, reducing opportunity for adsorption or reactions.

**Types of preferential flow** Three basic modes of preferential flow (Figure 8) are (1) macropore flow, through pores distinguished from other pores by their larger size, greater continuity, or other attributes that can enhance flow; (2) funneled (or deflected or focused) flow, caused by flow-impeding features of the medium that concentrate flow in adjacent zones that are highly wetted and conductive; and (3) unstable flow, which concentrates flow in wet, conductive fingers.

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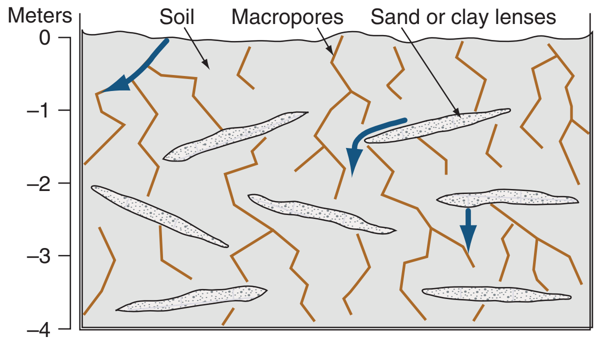


**Figure 8** Three basic types of preferential flow. Arrows indicate narrow regions of faster flow than their surroundings. Macropore flow occurs through channels created by aggregation, biotic activity, or similar causes. Funneled flow occurs when flow is deflected by heterogeneities of the medium so as to create zones of higher water content and greater *K*. Unstable flow can be generated at layer boundaries such as the bottom of a sand lens at right, where flow into the lower layer moves in the form of highly wetted fingers separated by regions of relatively dry soil.

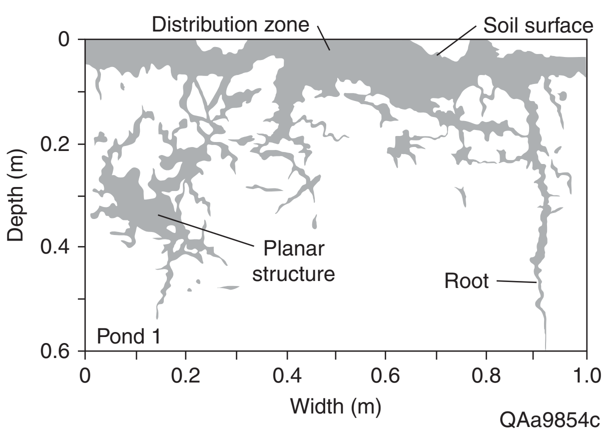


**Figure 9** Macropore flow paths highlighted using a dye tracer. Reproduced from Scanlon BR and Goldsmith RS (1997) Field study of spatial variability in unsaturated flow beneath and adjacent to playas. *Water Resources Research* 33: 2239–2252, with permission from the American Geophysical Union (http://www.agu.org/pubs/copyright.html).

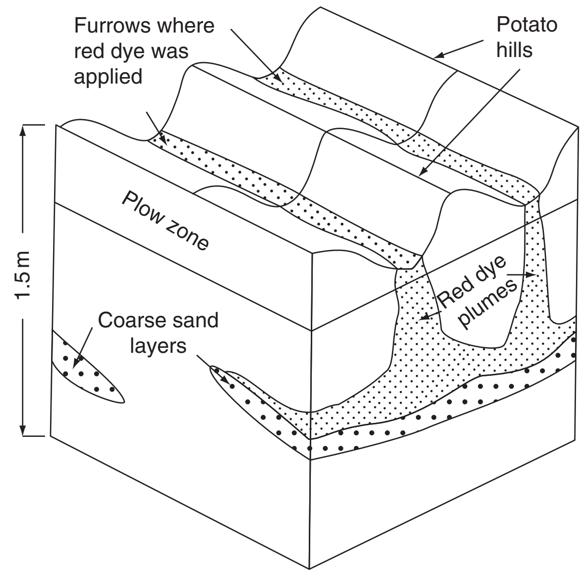


**Figure 10** Funneled flow identified using a dye tracer. Reproduced from Kung KJS (1990) Preferential flow in a sandy vadose zone: 1. Field observation. *Geoderma* 46: 51–58, with permission from Elsevier.

Common macropores include wormholes, root holes, and fractures (Figure 9). When macropores are filled with water, flow through them is fast. When they are empty, there may be essentially no flow through the macropores themselves though in some conditions film flow along macropore walls is significant. Macropores that are partly filled with water provide a variety of possibilities for the configuration and flow behavior of water.

Funneled flow commonly occurs with contrasting layers or lenses, where flow deflected in direction becomes spatially concentrated (Figure 10). The local increase in $\theta$ causes a corresponding increase in hydraulic conductivity and flux, and usually a change in the predominant direction of flow.

Unstable variations in flow and water content, even within a uniform portion of the medium, can increase flow rates considerably. A typical case has a layer of fine material above the coarse material. Downward-percolating water builds up significantly at the interface, and breaks through into the coarse medium at a few points. The material near individual points of breakthrough becomes wetter and hence much more conductive. For some time thereafter, additional flow into the coarse material moves in the few fingers that are already wet (Figure 11). Between fingers, the medium can be relatively dry. In addition to textural contrasts, hydrophobicity (water repellency) and air trapping may cause flow instability.

**Quantification of preferential flow** One straightforward quantitative treatment is to represent preferential flowpaths with discrete conduits whose geometry, with appropriate laminar-flow expressions, predict the flow rate through the part of the medium they occupy. Usually this requires a statistical characterization of the set of conducting pathways, because the position, number, shape, orientation, and connectedness of the individual pathways are unknown.

Perhaps more widely used are various forms of equivalent-medium approach. The key assumption is that the effective hydraulic properties of a large volume of the medium that includes preferential pathways are equivalent to the average properties of a

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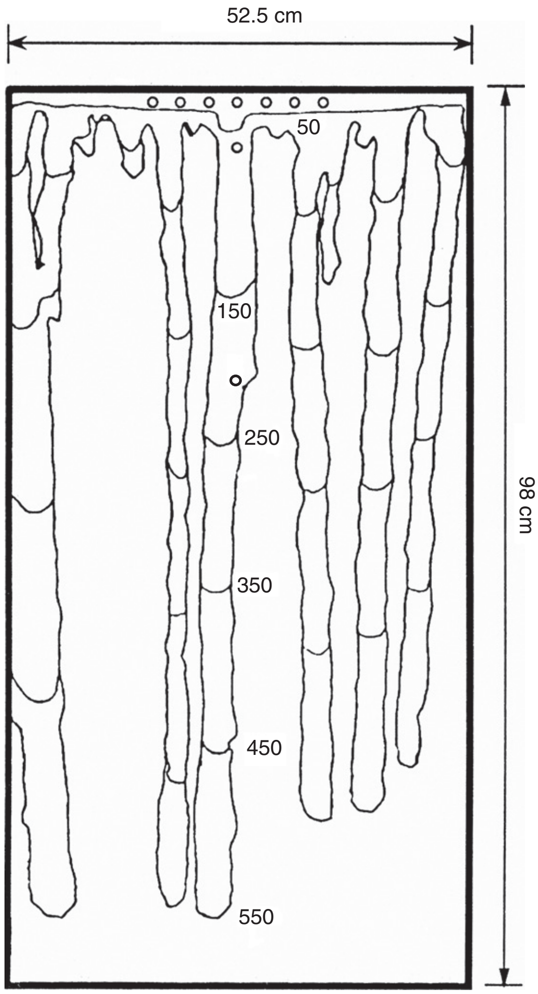


**Figure 11** Fingers generated in unstable flow in a laboratory investigation, showing the extent of wetted soil as a sequence in time. Numerals on the diagram indicate the number of seconds since flow began. Reproduced from Selker J, Leclerq P, Parlange JY, and Steenhuis T (1992) Fingered flow in two dimensions 1. Measurement of matric potential. *Water Resources Research* 28: 2513–2521, with permission from the American Geophysical Union (http://www.agu.org/pubs/copyright.html).

hypothetical homogeneous granular porous medium. The effective hydraulic properties then can be applied directly in numerical simulators using Darcy's law and Richards' equation. A major advantage of such an approach is that the many existing theories, models, and techniques developed for diffuse flow in granular media can be applied to preferential flow. A major drawback is that preferential flow may deviate significantly from behavior describable using this type of medium, precluding reliable results. It is also a common practice to treat preferential flow differently from nonpreferential flow (often combined with a conventional Richard's equation for the matrix flow).

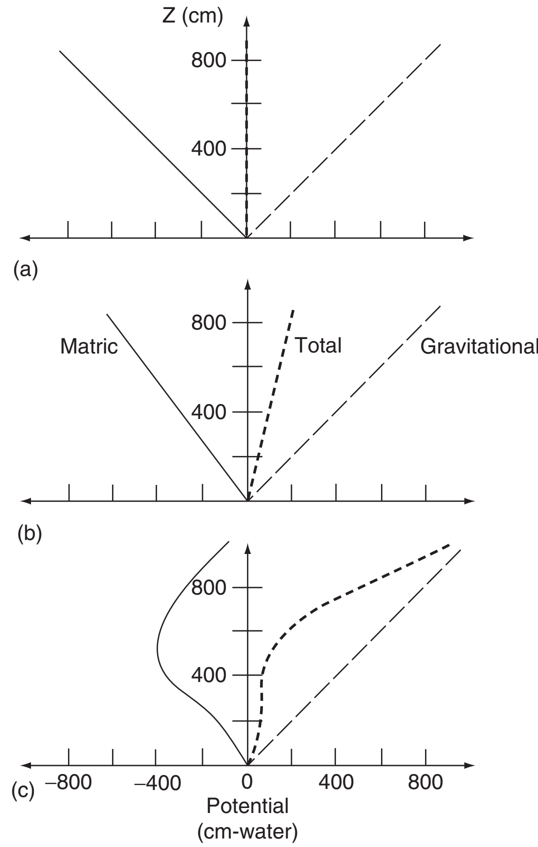


**Figure 12** Profiles of matric, gravitational, and total potential for idealized situations of (a) static water, (b) steady downward flow, and (c) unsteady flow. The water table is at $z = 0$.

## Vadose Water in the Hydrologic Cycle

### Moisture State in the Vadose Zone

The water content and matric pressure within the vadose zone influence and in turn are influenced by conditions in the saturated zone and atmosphere. The distribution of vadose water at a given time also depends on the energy state (whose components include matric and gravitational potential), wetting/drying history, and dynamics of the water itself. If there is no flow, one can infer that the gradient of total potential is zero, so if the matric and gravitational components are the only significant ones, they add to a constant total potential. Figure 12(a) shows this type of hydrostatic profile for the case where a water table is present. Since the matric pressure in this case is linear with depth and $\theta$ is controlled by the water retention properties of the medium, for a uniform vadose zone, the $\theta$ profile (not shown here) mimics the shape of the water retention curve. Given time, approximate versions of such a hydrostatic profile may develop in portions of a profile where

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water movement is negligible. If water flows vertically downward at a steady rate in a homogeneous medium, the total gradient must be constant, but the matric pressure does not cancel out the gravitational potential, as illustrated in Figure 12(b). In the general case of unsteady flow, the matric pressure profile cannot be determined so simply, and may take on an irregular form as in Figure 12(c).

The uppermost part of the water distribution profile is sometimes described in relation to field capacity, defined as the water content of a soil profile when the rate of downward flow has become negligible 2 or 3 days after a major infiltration. This concept is used in agriculture to indicate the wettest soil conditions to be considered for plant growth, and sometimes is mentioned in hydrologic investigations related to soil moisture storage. The definition of field capacity requires some subjective judgment, for example, in deciding what flow is negligible. Field capacity is implicitly associated with the entire soil profile through the root zone, including preferential-flow characteristics and, especially, flow-retarding layers that enable layers above them to retain a high water content.

In a portion of the vadose zone immediately above the water table, it may happen that all pores are filled with water, held by capillary forces. The depth interval that is saturated but above the water table is called a capillary fringe. In a hydrostatic profile, this corresponds to a flat portion of the retention curve between saturation and an air-entry pressure. Some media do not have a significant capillary fringe because their retention characteristics have the air-entry pressure at essentially zero. Where the water table fluctuates, the hydrostatic equilibrium needed for a capillary fringe may take considerable time to establish. Soil–water hysteresis would make for a different capillary fringe with a falling water table than with a rising water table.

## Moisture Dynamics in the Vadose Zone

### Interactions at the land surface

*Infiltration* Infiltration is the downward movement of water through the land surface. If the soil is initially dry, $\psi$ gradients may be the predominant downward driving force. When the soil is very wet to some depth, gravity may dominate instead. The usual case is that water infiltrates faster at the start and slows down as a zone of increased water content develops at the surface and expands. Figure 13 shows actual infiltration rates varying over time in three soils. Mathematically, the decline of infiltration rate as the soil gets wetter is frequently represented by an inverse proportionality to the square root of time, as predicted by several models of infiltration. If water at the surface is abundantly available, but not under significant pressure, infiltration occurs at the infiltration capacity, a rate determined by the soil rather than the rate of application or other factors. If water arrives at the land surface faster than the infiltration capacity, excess water ponds or runs off. Like hydraulic conductivity, infiltration capacity is not single-valued for a given medium but varies with water content and other conditions. Conditions that complicate the ideal conception of infiltration include: variation of application rate with time, spatial variability of soil and surface properties, water repellency of the soil, air trapping, and variations of temperature.

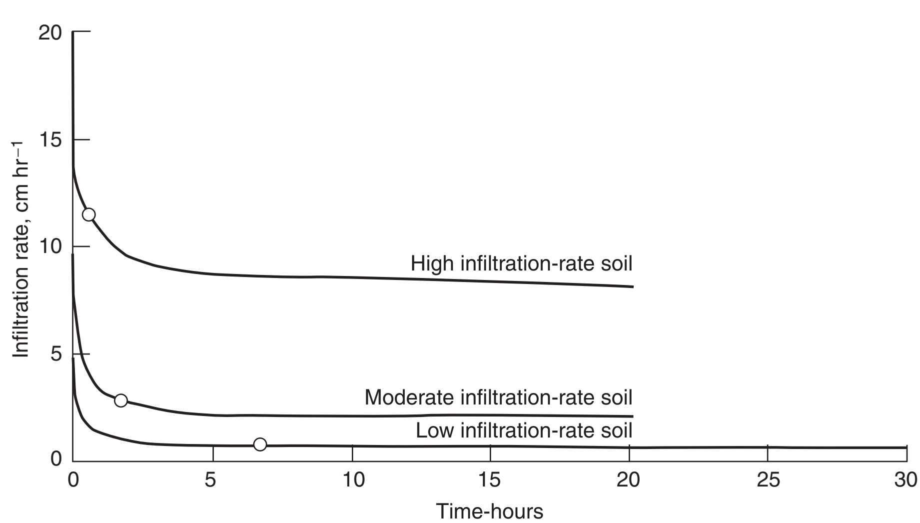


**Figure 13** Measured infiltration rates over time for three different soils. Reproduced from Swarner LR (1959) *Irrigation on western farms*. Agriculture Information Bulletin 199, U.S. Department of Interior and U.S. Department of Agriculture: Washington, DC.

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*Evapotranspiration* The transport of water from soil through plants to the atmosphere (known as transpiration) and the direct transport from soil to atmosphere (known as evaporation) together constitute evapotranspiration. When the soil is wet enough, atmospheric conditions control the evaporation rate. When the soil is too dry to supply water at the maximum rate the atmosphere can absorb, the soil properties will control the evaporation rate. Thus there are at least two cases to consider: the atmosphere-dominated 'constant-rate' phase during which the transport mechanisms of the soil are ignored, and the soil-dominated 'declining-rate' phase, during which atmospheric effects are ignored. On vegetated land, transpiration typically far exceeds evaporation. Capillary forces can draw water up from the water table to depths from which it supplies the process of evapotranspiration, which can be a substantial loss mechanism from a water-table aquifer, especially where the vadose zone is thin.

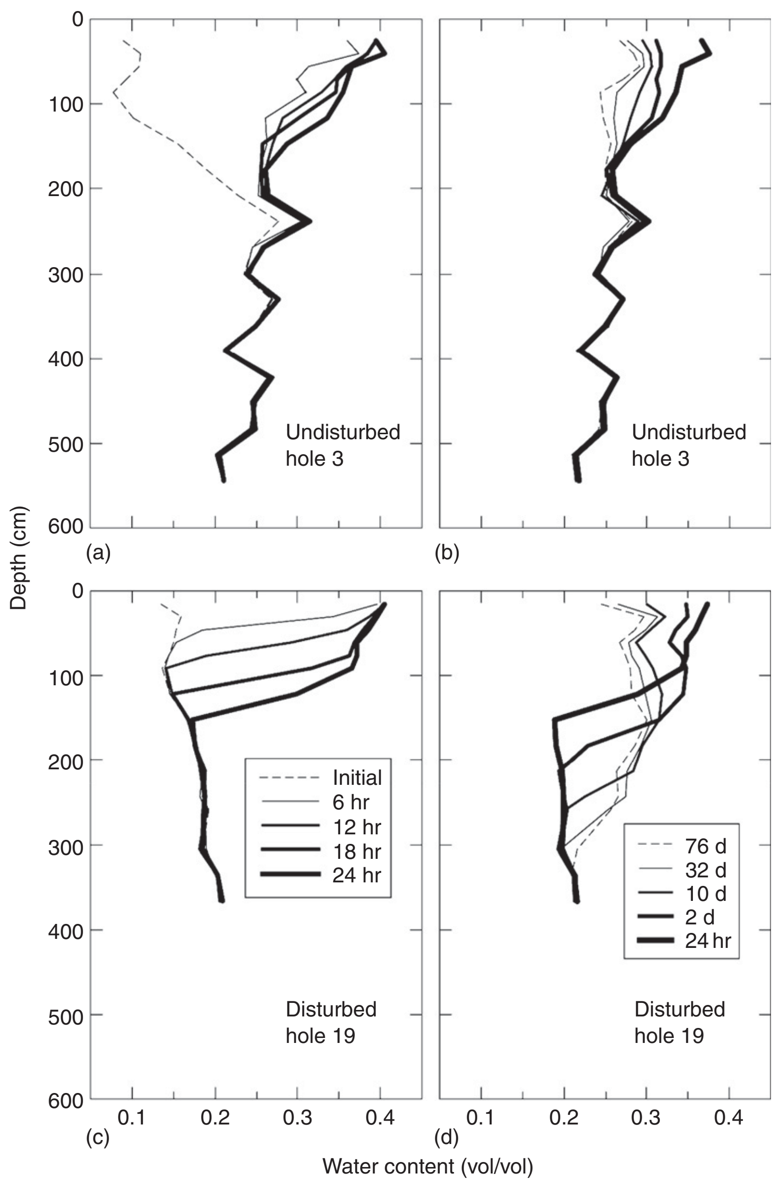


**Figure 14** Measured water distributions during and after 24 h of flood infiltration in (a, b) an undisturbed soil on the Snake River Plain in Idaho and (c, d) nearby soil that was disturbed by temporary removal and replacement. Evaporation was inhibited by an impermeable cover at the land surface. Reproduced from Nimmo JR, Shakofsky SM, Kaminsky JF, and Lords GS (1999) Laboratory and field hydrologic characterization of the shallow subsurface at an Idaho National Engineering and Environmental Laboratory waste-disposal site. Water-Resources Investigations Report 99–4263, U.S. Geological Survey: Idaho Falls, Idaho.

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**Redistribution of infiltrated water** After water has infiltrated, it redistributes, driven by gravity, matric pressure gradients, and possibly other forces. Figure 14 illustrates $\theta$ distributions at various times during and after infiltration, in a mechanically disturbed soil and in a soil with intact natural structure. Redistribution continues until all forces balance out. Equivalently, the water may be considered to progress toward a state of minimal (and uniform) total energy of the earth–water–air system, i.e., equilibrium.

Normally hysteresis strongly influences redistribution because a wetting front progresses downward according to the wetting curves of water retention and conductivity, whereas $\theta$ in the upper portions of the wetted zone decreases according to the drying curves. Because a drying retention curve has greater $\theta$ for a given $\psi$, water contents remain higher in the upper portions than they would if there were no hysteresis. Thus one important consequence of hysteresis is to hold more water near the land surface where it is accessible to plants.

Usually, the above considerations need to be adjusted or reinterpreted with attention to preferential flow. Qualitatively, a major effect of preferential flow is to permit more rapid movement of water to significant depths. This would occur primarily under very wet conditions, and would be followed in the redistribution process by a slower flow of water into the regions between preferential flow channels.

A common phenomenon in layered media is perching, the accumulation of water in a region of the vadose zone to the point where it becomes saturated even though there is unsaturated material between that region and the water table. The high water content of a perched zone causes greater hydraulic conductivity and potentially faster transport through the three-dimensional system. The main effect is not a direct increase in vertical flow, though possibly in horizontal flow. New and different conditions may affect biological and chemical processes in a perched zone, e.g., reduced aeration.

A situation comparable to perching exists when a body of surface water has a vadose zone underneath it (Figure 15). This may be caused by a flow-restricting layer at or beneath a lakebed or streambed. The situation may alternatively be thought of as a perched water body directly under the lake or stream. The key condition is that the impeding layer must reduce the downward flow rate to less than the saturated hydraulic conductivity of the layer immediately below it. Another way this can happen is as a transient response to ephemeral surface water, below which an unsaturated state may persist for some time after standing water has come into the depression or channel.

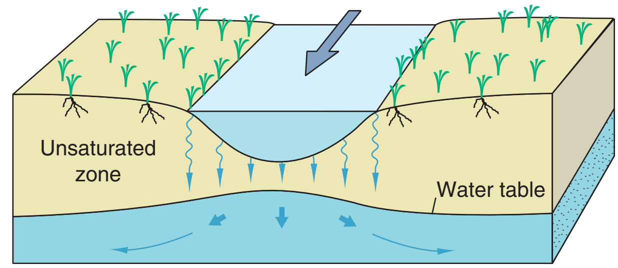


**Figure 15** A stream, disconnected from the water table, so that interaction between surface water and the aquifer occurs through the unsaturated zone. Adapted from Winter TC, Harvey JW, Franke LO, and Alley WM (1998) Ground water and surface water – A single resource. Circular 1139, U.S. Geological Survey.

### Aquifer Recharge

Aquifer recharge is water that moves from the land surface or unsaturated zone into the saturated zone. Quantitative estimation of recharge rate contributes to the understanding of large-scale hydrologic processes. It is important for evaluating the sustainability of groundwater supplies, though it does not equate with a sustainable rate of extraction. Where contamination of an aquifer is a concern, estimating the recharge rate is a first step toward predicting solute transport to the aquifer. Recharge may cause a short- or long-term rise of the water table. Artificial drainage, e.g., with horizontal porous pipes buried at a chosen depth, is sometimes used to maintain a minimal thickness of vadose zone for agricultural or other purposes.

Recharge rates vary considerably in time and space. Recharge often occurs episodically in response to storms and other short-term, high-intensity inputs. For a given amount of infiltration, temporal concentration enhances recharge because it entails shorter residence times for water in the portions of the soil from which evapotranspiration takes place. Similarly, a larger fraction will become recharge if it is concentrated in narrow channels such as fingers or macropores, not only because this tends to hasten its passage through the unsaturated zone, but also because the water then occupies less of the volume of soil from which evapotranspiration takes place.

## Conclusion

The state and dynamics of vadose water are complicated by the interaction of multiple phases. At least three drastically different substances – water, air, and solid mineral – are critical to its nature and quantification. Unsaturated flow phenomena are extremely sensitive to the proportions of those phases, especially the fluid phases, as natural variations in the relative amounts of water and air can cause a property like hydraulic conductivity to vary over many orders of magnitude. When the flow of vadose water is diffuse in character, it can be treated quantitatively with Darcy's law adapted for unsaturated flow, and with Richards' equation. When it occurs within preferential pathways, there are various models, none yet generally accepted, to quantify the flow. The state and dynamics of vadose water control or contribute to a wide variety of processes within the hydrologic cycle, including infiltration, evapotranspiration, infiltration and runoff, and aquifer recharge.

*See also:* Atmospheric Water and Precipitation; Evapotranspiration; Ground Water; Hydrological Cycle and Water Budgets.

## Further Reading

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