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150: Unsaturated Zone Flow Processes

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Water flow in the unsaturated zone is greatly influenced by unsaturated hydrostatics (water content, energy, pressure, and retention) and by unsaturated hydrodynamics (diffuse flow and preferential flow). Important multiphase processes include the transport of gases, nonaqueous liquids, and solid particles. Numerous means are available for determination of unsaturated conditions and properties, both measurement (of moisture state, water retention, and dynamic characteristics) and through various formulas and models that are mostly empirical in nature, but in some cases incorporating insight into unsaturated-zone physical processes. Applications to practical problems include models and techniques relating to distributions of water and energy, fluxes at the land surface, inputs, outputs, and fluxes within the unsaturated zone, all of which are frequently complicated by heterogeneity and preferential flow. Further scientific advance requires new measurement techniques and theoretical constructs that more adequately represent the important physical processes within practical modeling schemes.

INTRODUCTION

The unsaturated zone, sometimes called the *vadose zone* or *zone of aeration*, plays several critical hydrologic roles. As a storage medium, it is a zone in which water is immediately available to the biosphere. As a buffer zone between the land surface and aquifers below, the unsaturated zone is a controlling agent in the transmission of contaminants and aquifer-recharging water. As an accessible body of material in which physical and chemical processes may be relatively slow, it is a place where wastes are emplaced to isolate them from significant exchange with other environmental components. Thus, the flow processes that occur in the unsaturated zone substantially contribute to a wide variety of hydrologic processes.

Scientifically, the unsaturated zone is highly complex and must be studied with an interdisciplinary approach. 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.

This article first describes in a fundamental way the physical basis of phenomena that strongly relate to unsaturated flow. The next section presents techniques for obtaining quantitative values of properties that influence unsaturated flow, by direct measurement and by indirect means. The third major section describes some of the main hydrologic applications related to flow in the unsaturated zone.

PHENOMENA OF UNSATURATED ZONE FLOW

Water resides in an unsaturated porous medium along with air and solids, as Figure 1 illustrates. The usual tendency is for water to cling to solid surfaces, in films, and in curved air–water interfaces as shown. Hydrological processes involve movement of any of these materials. Thus, transport in an unsaturated medium is always a case of multiphase transport, though the term "multiphase" is used mainly for cases where gas, solid, or multiple-liquid phases are considered.

Various materials comprise the solid fabric of the unsaturated zone, including soil, stones, porous rock, and organic matter. A common distinction is between particulate, or granular, media in which particles are separate from each other, and consolidated or lithified media in which they are joined.

Basic features of soil and rock relevant to flow in the unsaturated zone include texture, structure, and

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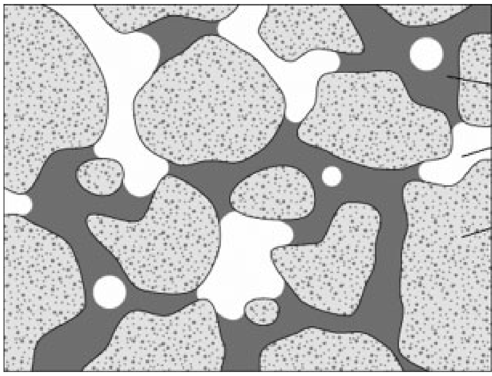

Figure 1 Microscopic cross-sectional view of a hypothetical unsaturated medium. Few intergrain contacts appear because these contacts are essentially points in three-dimensional space, and mostly do not lie in the two-dimensional plane of this figure

the content of mineral and nonmineral constituents (*see also* **Chapter 147, Characterization of Porous and Fractured Media, Volume 4**). Texture, applied to a granular medium, refers to particle-size distribution. Structure refers to the arrangement of the solid components of the medium. Essential structural considerations include porosity, aggregation, cementation, and macropores. Structure is often considered alongside texture as a primary determinant of the pores of the medium.

This section considers unsaturated-zone flow phenomena through progressive increases in complexity as elaborated in Table 1. It concludes with a discussion of multiphase flow, with explicit attention to the transport of materials other than liquid water.

Unsaturated Hydrostatics

Water Content, Energy, and Pressure

The most basic measure of the water is volumetric water content, symbolized θ or θ_w, defined as the volume of water per bulk volume of the medium. An alternative is the mass-basis (or gravimetric) water content θ_m, the mass of water per mass of solid. These are related by the formula

$$\theta_w = \frac{\rho_b}{\rho_w}\theta_m \tag{1}$$

where ρ_b is the bulk density and ρ_w is the density of water. In general, the volumetric water content is most useful because its range has a clearly defined maximum at the medium's porosity, ϕ, and it relates easily to visualizations of intrapore geometry as in Figure 1.

Water is held in an unsaturated medium by forces whose effect is expressed in terms of the energy state of the water, force being the negative gradient of energy. The energy is usually expressed as a potential, taken as energy per unit volume. For an incompressible bulk material like water, the energy per unit volume can equivalently be considered as a pressure. Gravitational energy can be treated essentially as it is for saturated modes of fluid transport. The matric potential or pressure, which often is the only other significant type of energy determining the chief water transport processes in an unsaturated medium, arises from the interaction of water with a rigid matrix ("matric" being the adjective form of "matrix"). Sometimes this quantity is called *capillary potential* or *pressure*. Matric pressure 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. When a medium is unsaturated, the water generally is at lower pressure than the air, so the matric pressure is negative. Another term is "suction", the negative of matric pressure. Many problems are simpler with the expression of matric pressure in head units.

Matric pressure is often thought of in relation to surface tension and capillary phenomena, especially in the range near saturation. A pore with water can be compared to a thin capillary tube with one end immersed, as in Figure 2.

Table 1 Levels of complexity considered in unsaturated flow

Type of flow	Phenomena	Mathematical description	Relevant features and properties of medium	Typical applications
Static (no flow)	All forces (on water) balance	Hydrostatic equation	Water retention	Available water for plants; basis for understanding
Steady	Flow driven by unchanging force field	Darcy's law	Hydraulic conductivity	Long-term averages
Unsteady – diffuse	Continuity, and force field affected by dynamic conditions	Darcy's law and continuity equation (combined as Richards' equation)	Hydraulic conductivity and water retention	Stable flow in media whose pore sizes approximate grain sizes
Preferential	Flow concentrated in pathways that differ in character from the bulk medium	Unknown; diverse alternatives in current use	Macropores, layer contrasts, propensity for unstable flow	Flow in media with significant nonuniformities; unstable flow

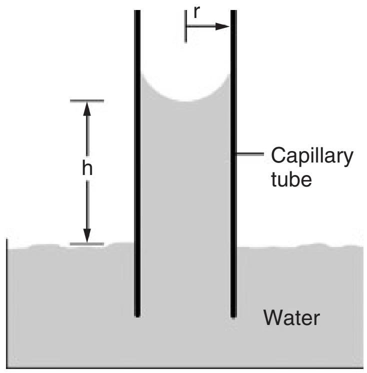

Figure 2 Capillary rise in a tube of circular cross section

The curvature of the air–water interface is inversely related to the water pressure: tighter curvature is associated with smaller pores and with more negative matric pressure. To quantify the relation between the tube radius and matric pressure ψ, consider the pressure in the water just below the curved air–water interface, which equals ψ. For the static situation, upward and downward forces balance. The downward force is the pressure acting on the area of the air–water interface, $\psi\pi r^2$. The upward force results from surface tension σ, the force per unit length of the air–water interface, acting on a circumference of the inner wall of the tube. For the case where the air–water interface is perfectly tangent to the tube wall, this force is $2\pi r\sigma$. Equating these forces gives

$$\psi = -\frac{2\sigma}{r} \tag{2}$$

Pores in natural media are not perfectly cylindrical, but the same relation applies if it is understood that r represents an effective pore radius. The relation given in equation (2) is useful in picturing what happens within a pore and in modeling unsaturated hydraulic properties.

Where the three phases come together, the angle between the air–water interface and the air–solid interface, measured through the water, is the contact angle. This angle depends on the materials. If the solid has much greater attraction to the liquid than to the air, it is highly wettable and the contact angle can be essentially zero, as assumed in the derivation of equation (2). If the solid has much greater attraction to the air, the material is nonwettable and would have a contact angle near 180°. Intermediate cases have intermediate values of contact angle, and some materials are difficult to classify as wettable or nonwettable. For a nonzero contact angle, formulas like (2) are modified by multiplying the force associated with surface tension by the cosine of the contact angle.

When the medium is so dry that water does not fill pores, but adheres in thin films to the solid matrix, the concepts of surface tension and capillarity are not so directly applicable, and the forces of adhesion establish the matric pressure. One can think of a capillary component of matric pressure that dominates when the medium is wet, and an adhesive component that dominates when it is dry.

Water Retention

If the matric pressure is close to zero, air–water interfaces are broadly curved, nearly all pores are filled, and the water content is high. If matric pressure is much less than zero, the interfaces are more tightly curved, they can no longer go across the largest pores, and the pores have less water in them. Thus, greater water content goes with greater (less strongly negative) matric pressure.

The relation between matric pressure and water content, called a *water retention* curve, depends on the medium. Larger pores empty first as the water content decreases. A medium with many large pores will have a retention curve that drops rapidly to low water content at high matric pressures. Conversely, a fine-pored medium will retain much water even at low matric pressures, and so will have a flatter retention curve. Figure 3 shows a typical retention curve for a sandy soil. The curve is far from linear and covers 5 orders of magnitude in ψ. This enormous range is difficult to work with and requires multiple measurement techniques. In most cases, investigators measure and plot only a portion of the range, usually at the wet end.

Considering the drying of soil from saturation, θ in Figure 3 stays high until a particular ψ value where it starts to decline. This ψ is called the *air-entry value*. By the capillary hypothesis, it exists because the largest

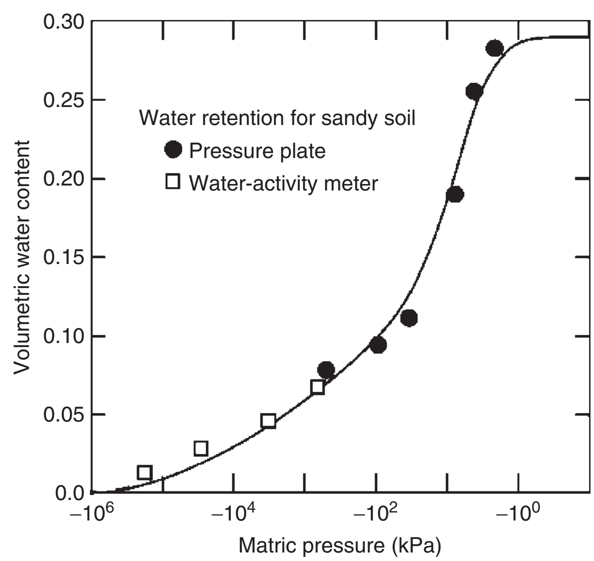

Figure 3 A retention curve for a sandy soil from the amargosa desert research site (Andraski, 1996). The points are measurements by two different methods and the smooth curve is a fit of the model of Rossi and Nimmo (1994)

fully wet pore of the medium will stay filled until the air–water pressure difference exceeds the equivalent ψ value of capillary rise. In natural media the air-entry value is usually poorly determined, as the decline in θ with ψ starts gradually, beginning at ψ 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.

The maximum θ value of the curve usually is not equal to ϕ but rather something less, because at $\psi = 0$, trapped air occupies some of the pore space. At the other end, the curve goes to $\theta = 0$ at a ψ value of about -10^6 kPa.

In a granular medium, the particle-size distribution or texture relates in a general way to the pore-size distribution, as larger particles may have larger pores between them. Texture thus is a major influence on the retention curve. Additionally, the structure of the medium, especially as related to such features as aggregation, shrinkage cracks, and biologically generated holes, substantially influences the retention curve.

At low matric pressures, few pores are filled and a large fraction of the total water is in thin films. These films are thinner at lower matric pressures, with less energy for holding water onto the solid medium. At high matric pressures thicker films can be important, as when the medium is nearly saturated except for fractures and other large pores. If the films are thick enough, water in them may be free to flow. Tokunaga and Wan (1997) have measured film thicknesses of tens of microns at matric pressures between -0.1 kPa and 0.

The retention relation is strongly hysteretic: when measured as the medium wets, water content is less for a given matric pressure than it is when measured as the medium dries. Figure 4 shows a typical example. The outer curves, starting from extreme wet or extreme dry conditions, are called *main drying* and *main wetting* curves. The curves starting from intermediate water contents are called *drying* and *wetting scanning* curves. Of course, there are whole families of possible scanning curves, starting from different points. Scanning curves that start on main curves are called *primary scanning* curves, those that start on primary scanning curves are secondary scanning curves, and so forth. Unsaturated zone investigations frequently neglect hysteresis, not always justifiably.

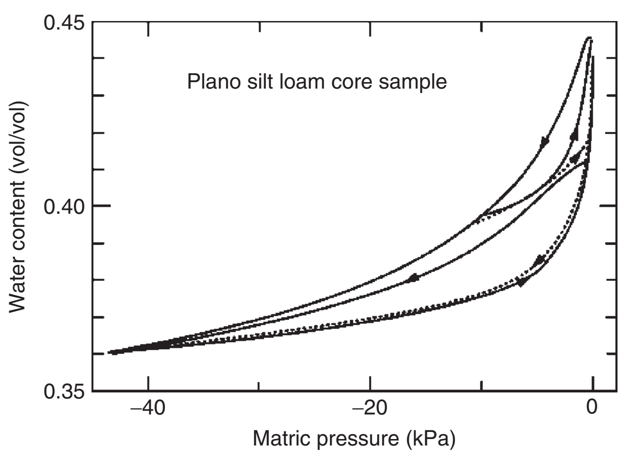

Figure 4 Hysteretic water retention in a soil core sample, data of Nimmo and Miller (1986)

Several mechanisms cause $\theta - \psi$ hysteresis, the main one in fairly wet media being the Haines jump mechanism, illustrated in Figure 5. Unlike capillary tubes, pores in natural media are not uniform in effective diameter. The pore throats, which control the ψ at which pores empty and therefore determine the drying curve, are smaller than the pore bodies, which control the ψ at which pores fill and determine the wetting curve. As the medium dries and ψ decreases, water retreats gradually as the air–water interface becomes more curved. At the narrowest part of the pore throat, this surface can no longer increase curvature by gradual amounts, so in what is called a *Haines* jump, it retreats suddenly to narrower channels elsewhere in the nearby pore space. An analogous phenomenon occurs during wetting, when the decreasing interface curvature cannot be supported by the diameter of the pore at its maximum. Hysteresis occurs because the drying and wetting Haines jumps occur at different ψ values. Some of the pore space, where the movement of the interface is always gradual, is not subject to hysteresis. The amount of such nonhysteretic pore space varies with the pore geometry of the medium, and thus is one phenomenon that makes the degree of hysteresis vary among media. Other mechanisms include contact angle hysteresis, which is not well understood because contact angles within pores are difficult to measure, and adsorptive hysteresis, which may be quite significant near the dry end of the moisture range.

As explained below in connection with gas transport, natural media do not usually become fully saturated, even at $\psi = 0$. Some amount of air normally gets trapped in the form of bubbles enclosed by water, typically occupying 10 to 30% of the pore space. Thus, the main drying curves in Figures 3 and 4 have a maximum θ that is less than ϕ. Sometimes, though, the medium may be saturated long enough for all trapped bubbles to dissolve. On drying, the retention curve would then start from $\theta = \phi$, as illustrated by the data in Figure 6. This is called a *first* (or initial or primary) *drying* curve. One natural phenomenon likely to involve a first drying curve is the decline of a water table after a long time at a high level.

Temperature, because it affects surface tension and other relevant properties, significantly affects the retention relation. Increasing temperature means that less water will be held at a given matric pressure. This effect can be quantified using the gain-factor model of Nimmo and

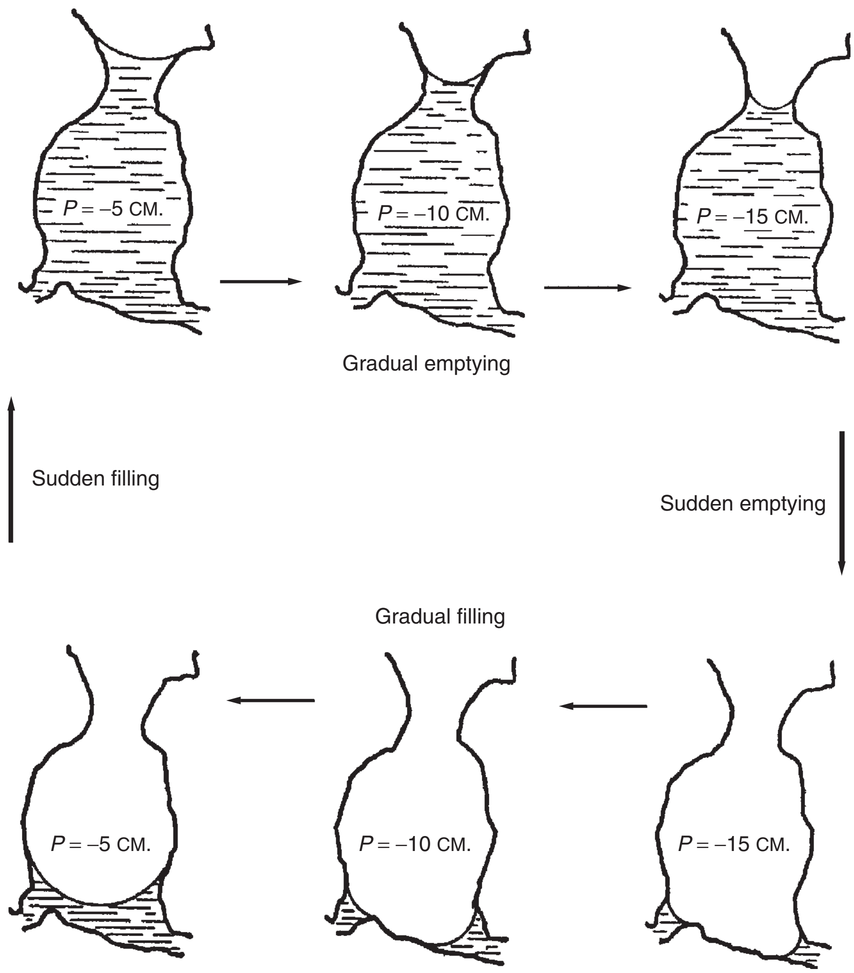

Figure 5 Haines jumps in a natural pore, illustration from Miller and Miller (1956) (Reprinted with permission from Miller, E. E. and Miller, R. D. (1956). Physical theory for capillary flow phenomena. *Journal of Applied Physics*, **27**, 324–332. © 1956 American Institute of Physics.)

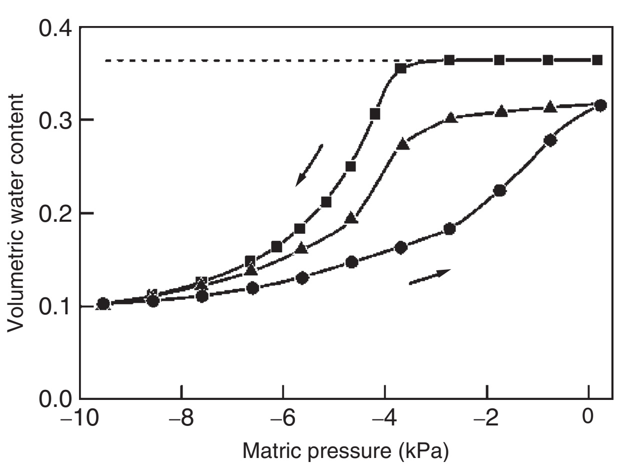

Figure 6 A first drying curve, shown with main wetting and drying curves, measured for a sandy soil (Oakley sand). The dashed line is at the value of θ equal to the porosity. Data of Stonestrom and Rubin (1989a)

Miller (1986) or the enthalpy-based model of Grant and Salehzadeh (1996).

Basic Unsaturated Flow

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 exceed the gravitational force. Other forces may be significant for driving flow under some conditions, as when temperature gradients are significant, or when there are physical barriers to the movement of solutes. The medium properties of chief importance are the water retention relation and the hydraulic conductivity.

Unsaturated flow has its basic mathematical expression in Darcy's law, which states that the flux density q is proportional to the driving force. The proportionality constant is the hydraulic conductivity K. For the case of one-dimensional flow driven by gravity and matric pressure gradients, Darcy's law can be expressed as

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

where ρ 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 ψ in SI pressure units such as kPa, and K in velocity units such as m/s. In head units, ρg (the weight of water per unit volume) equals dimensionless unity and ψ takes dimensions of length.

K of the medium depends on the whole set of filled pores, especially the size, shape, and connectedness of filled channels. The retention relation and the history of the moisture state determine what pores are filled. In unsaturated media, as illustrated by the measurements in Figure 7, K depends very strongly on the water content. As water content decreases, the large pores, which make by far the greatest contribution to K, empty first. Then, not only are there fewer filled pores to conduct water, but they are smaller and therefore less conductive because there is more viscous friction. With fewer pores filled, the paths of water flowing though the medium become more tortuous. When the soil is quite dry, few pores are filled, and the water moves mainly through poorly conducting films adhering to particle surfaces. These factors combine to reduce hydraulic conductivity by several orders of magnitude as the soil goes from saturation to typical field-dry conditions.

Other factors also can influence hydraulic conductivity. Matric pressure is relevant though its main effect is indirect, through influence on θ. Temperature affects K through its influence on viscosity and other factors. Microorganisms can reduce K by constricting or obstructing channels though the pores. Chemical activity can similarly reduce K if precipitates form. It can also influence K by affecting the cohesion of particles and hence the structure of the medium.

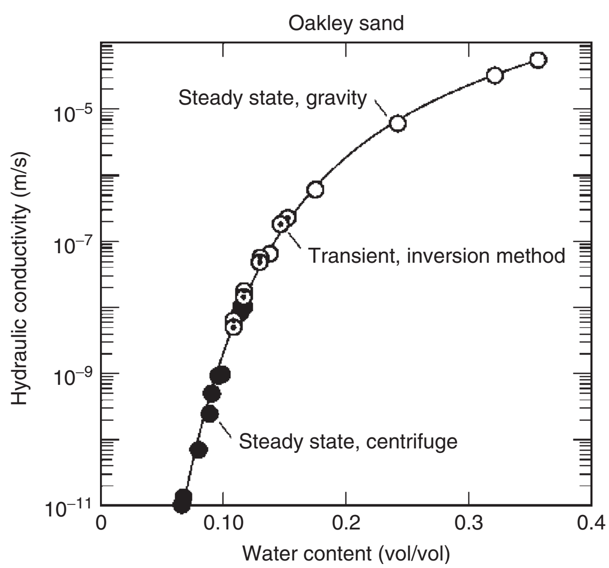

Figure 7 Hydraulic conductivity for a sandy soil (Oakley sand), measured by three methods

Unsteady Diffuse Flow

When unsaturated flow is transient (nonsteady), as it generally is, the flow itself causes the water content to change throughout the medium, which leads to continuously changing hydraulic conductivity and driving forces. These interacting processes, first described by Buckingham (1907), can be accommodated mathematically by combining the equation of continuity

$$\frac{\partial\theta}{\partial t} = -\frac{\partial q}{\partial z} \tag{4}$$

with Darcy's law (3) to get Richards' (1931) equation, which for one-dimensional vertical flow within a medium in earth gravity can be written

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

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 θ (Sposito, 1986). The equation can be solved numerically in the general case (e.g. Lappala *et al.*, 1987). Analytical solutions have also been developed (e.g. Salvucci, 1996), though these require simplifying assumptions that are not directly applicable to most situations of unsaturated flow. Richards' equation does not adequately represent all circumstances of unsaturated flow, for example, at wetting fronts and flow instabilities (Stonestrom and Akstin, 1994; DiCarlo, 2004). Some of these are discussed below in connection with unstable flow.

An alternative formulation of unsteady unsaturated flow depends on a property called *hydraulic diffusivity* or *soil-water diffusivity*. This is based on the fictional but useful assumption that the flow is driven by gradients of water content rather than potential. Richards' equation (5) without the gravitational term can be transformed into

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[D(\theta)\frac{\partial}{\partial z}(\theta)\right] \tag{6}$$

where

$$D(\theta) = K(\theta)\frac{d\psi}{d\theta} \tag{7}$$

is the hydraulic diffusivity (Childs and Collis-George, 1950). $D(\theta)$ is a property of the medium dependent on θ and its history. Advantages of this formulation include that: (i) $D(\theta)$ can be easier to measure than $K(\theta)$ and

$\theta(\psi)$; (ii) the typical increase of $d\psi/d\theta$ with decreasing θ compensates in part for the decline in K with decreasing θ, so that in the field D usually varies less than K; and (iii) many mathematical techniques for solving equation (6) have previously been developed for diffusion applications.

Preferential Flow

Preferential Paths

Flowpaths that permit fast movement, whether because of their character (e.g. large pore diameter) or their present state (e.g. high water content) are called *preferential paths*. Such a path may be a single pore (i.e. a macropore), a connected series of pores, or a group of adjacent pores acting in parallel. These paths are common in natural porous media with significant heterogeneity, such as surface soils and fractured rock. Flow in preferential paths transports water and contaminants much faster than would be predicted from bulk medium properties and Richards' equation. Another important effect of preferential flow is that a relatively small fraction of the subsurface medium interacts with contaminants, which limits adsorption and other attenuating processes. Three basic types of preferential flow (Figure 8) are (i) macropore flow, caused by flow-enhancing features of the medium; (ii) funneled (or deflected or focused) flow, caused by flow-impeding features of the medium; and (iii) unstable (or fingered) flow, caused by temporary flow-enhancing conditions of parts of the medium.

Macropores, distinguished from other pores by their larger size, greater continuity, or other attributes, conduct preferential flow under conditions such as extreme wetness. Common macropores include wormholes, root holes, and fractures. Where macropore flow occurs, flow through the remainder of the medium may be called *matrix flow*. When macropores are filled, flow through them is fast. When macropores are empty, they constitute a barrier to matrix flow and there may be essentially no flow through the macropores themselves. In some conditions, however, there may be significant film flow along macropore walls (Su *et al.*, 2003). Macropores that are partly filled provide a variety of possibilities for the configuration and flow behavior of water.

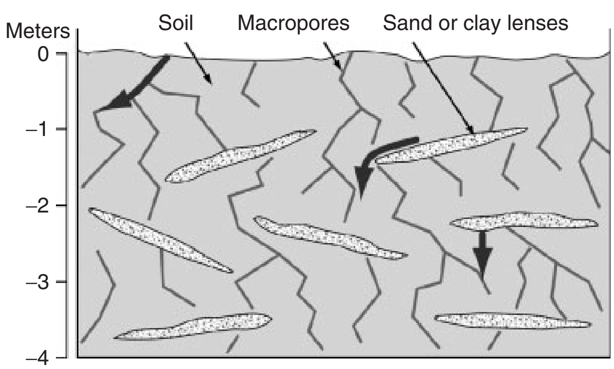

Figure 8 Three basic types of preferential flow. Arrows indicate narrow regions of faster flow than their surroundings. Macropore flow occurs through channels created by aggregation, biotic activity, or similar causes. Funneled flow occurs when flow is deflected by heterogeneities of the medium so as 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. A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

Funneled flow commonly occurs with contrasting layers or lenses, where flow deflected in direction becomes spatially concentrated. The local increase in water content causes a corresponding increase in hydraulic conductivity and flux. In field experiments in a sandy soil, Kung (1990) found that the flow became more preferential with depth (Figure 9). At about 6-m depth, the flow was moving through less than 1% of the whole soil matrix. Although this medium had no significant observable macropores, preferential flowpaths were the dominant pattern. The main feature causing this preferential flow was "... an interbedded soil structure with textural discontinuities and inclined bedding planes". Considering Kung's work and others', Pruess (1998) noted that because funneling results from horizontal impediments and can produce very rapid flow "... we have the remarkable situation that unsaturated seepage can actually proceed faster in a medium with lower average permeability".

Unstable variations in flow and water content, even within a uniform portion of the medium, can increase flow rates considerably (Wang *et al.*, 2003). A typical case has a layer of fine material above coarse material. Downward-percolating water does not immediately cross

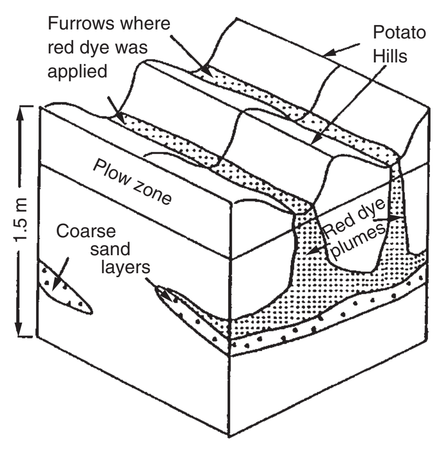

Figure 9 Funneled flow investigated using a dye tracer (Reprinted from Kung *et al.* 1990, © 1990, with permission from Elsevier)

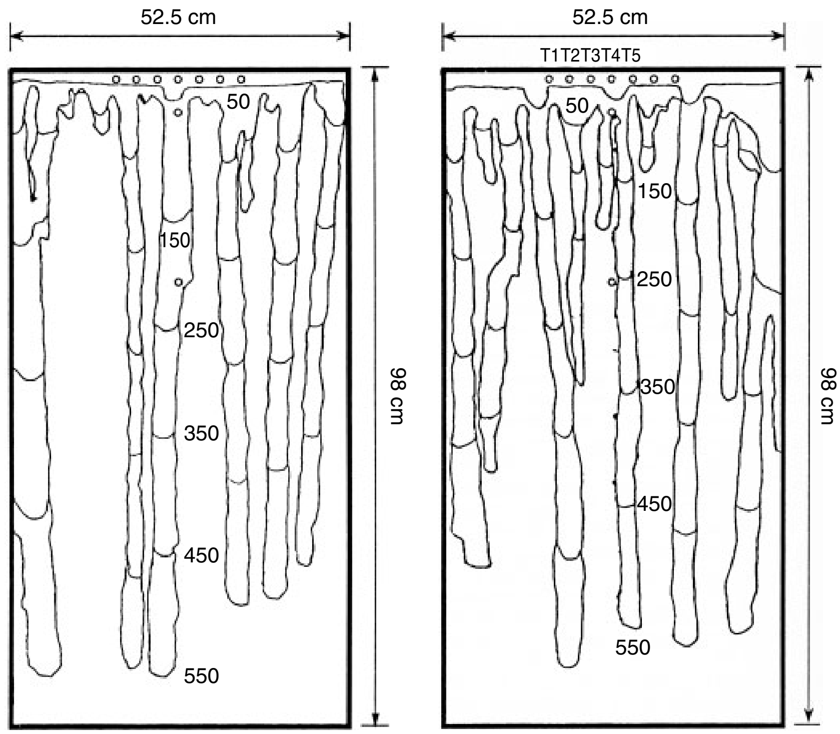

Figure 10 Fingers generated in unstable flow in a laboratory investigation (Reproduced from Selker *et al.*, 1992, by permission of American Geophysical Union)

the interface into the coarse material. When water pressure builds up significantly at the interface, water may break through into the coarse medium at only 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 10). Between fingers, the medium can be relatively dry. In addition to textural contrasts, hydrophobicity (water repellency) and air trapping may contribute to flow instability. Layer boundaries may interrupt preferential flowpaths. There may also be processes that homogenize preferential flow, though such effects have been little studied.

In virtually every unsaturated-zone transport problem, it is essential to assess the prevalence of preferential flow and the flow mechanisms that might be active. One approach is to evaluate the features of a particular site that might cause macropore, funneled, or unstable flow. Another is to collect and evaluate evidence from observed water or solute distributions that cannot easily be explained without hypothesizing preferential flow.

Quantification of Preferential Flow

One straightforward way of quantitatively treating preferential flow is by representing preferential flowpaths with discrete pathways whose geometry and water content, with appropriate laminar-flow expressions, predict the flow rate through the part of the medium they occupy. Usually this is impossible because the position, number, shape, orientation, and connectedness of pathways are unknown.

Conceptually opposite to the discrete pathway approach is the widely used equivalent-medium approach. The key assumption is that the effective hydraulic properties of a large volume of the medium are equivalent to the average properties of a homogeneous granular porous medium. The effective hydraulic properties then can be applied directly in numerical simulators employing Darcy's law and Richards' equation. The main advantage of the equivalent-medium approach is that the many existing theories, models, and techniques developed for diffuse flow in granular media can be applied to preferential flow. This approach is only valid if certain conditions are met. The medium must be representable as a continuum (Bear, 1979, p. 28–31), but this requirement is difficult to satisfy in macroporous media with flow inhomogeneities at relatively large scales. Representative volumes may necessarily be so large that only a small fraction of the assumed volume is participating in the flow. Modeled transport velocities then may greatly underestimate the actual velocities. In practice, an adjustment of the effective porosity of the equivalent granular medium is commonly done to compensate for such effects, though this adjustment will not produce accurate predictions if, as expected, the degree of preferential flow depends strongly on the degree of saturation. The equivalent-medium approach also is inadequate where it is essential to have knowledge of the individual flowpaths.

Unstable flow complicates the quantification in at least two ways that do not apply to macropore or funneled flow: (i) unstable flow is not tied to particular permanent features of the medium; and (ii) the preferentiality of unstable flow changes dynamically (e.g. unstable flowpaths commonly grow wider as flow progresses through them). Theories of unstable flow in terms of scaling and other concepts, have been developed for example, by Raats (1973), Parlange and Hillel (1976), Glass *et al.* (1989), Selker *et al.* (1992), Wang *et al.* (1998), Eliassi and Glass (2002), and Jury *et al.* (2003).

Multimodality

A family of approaches that rely on a conceptual partitioning of water or pore space into portions with different flow rates and behaviors may more realistically represent flow that includes preferential paths than do traditional unsaturated-flow models. Models used in these approaches have names such as "dual-porosity" and "dual-modality". The concept of mobility, that is, how easily certain system components (in particular, water and contaminants) move within different parts of the medium, is frequently used in defining and characterizing these approaches.

The simplest of these models assume matrix flow to be negligible, so that all flow is preferential flow. Given the nonlinear nature of unsaturated flow, the difference in conductance between, say, a wormhole, and an interparticle space between clay or silt grains, may be several orders of magnitude. For practical purposes this may constitute a mobile–immobile distinction. Other models assume the matrix to be permeable but with different properties and possibly different modes of flow than the portion of the medium that has preferential flow.

The degrees of possible mobility cover a continuum, and truly immobile water is unlikely. Some models postulate three degrees of effective mobility, for example, mesopores in addition to micropores and macropores (Luxmoore, 1981). A closer approach to a continuum of mobility is that of Griffioen *et al.* (1998).

Solute transport occurs in both diffuse and preferential modes. **Chapter 152, Modeling Solute Transport Phenomena, Volume 4** describes important phenomena, many of which are as relevant in unsaturated as saturated media.

Multiphase Flow

The transport of water and solutes is often considered independently of other fluids or solids. But normally, more than one phase is transported in an unsaturated medium. Besides the liquid water, air and other gases, there may be nonaqueous liquids that need to be considered as separate phases, and there may be solid particles that are free to move. Multiphase flow is a common element of contamination problems because many contaminants are nonaqueous, for example, volatile vapors, organic liquids (oils, solvents, etc.), colloids, and microorganisms.

For gases and nonaqueous liquids, it is often helpful to use the concept of permeability to represent the flow-influencing property of the solid matrix independently of the fluid. This relates to K according to

$$k = \frac{K\mu}{\rho g} \tag{8}$$

where k (dimensions L^2) is the permeability, μ is the viscosity (dimensions $M L^{-1} T^{-1}$), and ρ ($M L^{-3}$) is the density of the fluid. Darcy's law (3) takes the form

$$q = -\frac{k k_r(\theta)}{\mu}\left[\frac{d\psi}{dz} + \rho g\right] \tag{9}$$

where k_r is relative permeability, a dimensionless factor needed to account for the θ dependence in an unsaturated medium. Ideally, a medium has the same k for any fluid. This independence is not absolute because fluid properties other than μ and ρ, such as those related to slip effects (described below) or other non-Newtonic phenomena, can affect the flow, and some fluids cause structural changes that affect k.

Transport of Gases

Lacking the strong intermolecular forces of a liquid, individual gas molecules move nearly independently. Gas in the unsaturated zone commonly extends over large distances without being blocked by liquid, forming a continuous gas phase. Unlike solutes in liquid, the minority components within a gaseous mixture tend to mix completely at the molecular level, and to move in the same way as any other gas molecules. The more independent motion of the molecules also makes gases less viscous than liquids. Gas is highly compressible, which complicates some transport phenomena while making possible some useful measurement techniques. Gas flow exhibits slip, also called the *Klinkenberg* effect. Even the gas molecules closest to surfaces are free to move, and, unlike the essentially stationary boundary molecules of liquids, will have a net motion in the direction of the bulk flow of gas. This makes the permeability to gas somewhat greater than one would expect from measurements of the liquid permeability. Gas molecules dissolve in liquids. Like other solutes, gases differ in solubility. Carbon dioxide is much more soluble than nitrogen or oxygen. Gases also interact in chemical or biological reactions, of great importance in some contamination and restoration problems.

Strictly speaking, there is viscous friction between gas and adjacent liquid, so moving liquid tends to drag gas with it, and vice versa. Usually this friction is so much smaller than other driving forces that it can be ignored.

Gas flow is then independent of liquid flow, and often described by a separate implementation of Darcy's law. The gas pressure takes the place of matric pressure. Gas flow is often slow because the driving forces are small. When components of the gas are of comparable density, gravity is not a major driving force, and, unlike water driven by differences in matric pressure, in gases there normally is no mechanism to sustain large pressure gradients. Therefore convection is often not as important as other mechanisms, especially diffusion. In a gas, the molecules of a minority component move more easily than they would in a liquid, because of the lesser intermolecular cohesion. This causes greater diffusion. Dispersion, on the other hand, still depends largely on convective flow rates, and so is not necessarily greater in a gas. In contrast to solute transport, therefore, diffusion of gases is normally more important than dispersion. The diffusion coefficient for two gases in a porous medium is directly related to the diffusion coefficient for the gases in a free space, but also depends on such factors as pore size, tortuosity, and continuity.

Gas is trapped if it resides in bubbles or pockets from which every possible path to the outside of the medium goes through liquid or solid. Air easily becomes trapped during wetting, wherever a pore is slower than its neighbors to fill with water. Once the pore is surrounded by water, air within it is trapped. Stonestrom and Rubin (1989b) found experimentally during the drying and wetting of two different soils that some amount of air was trapped whenever θ was greater than about 70% of the porosity. A medium that has been "saturated" by ordinary means usually has a significant fraction of its pore space occupied by trapped bubbles of air within certain pores. Sometimes the word "satiated" is used for this condition, to reserve "saturated" for the case when there is strictly no gas in the pores. In general, though, "saturation" means that the medium has finished absorbing water by the wetting process it has been subjected to. This normally results in a matric pressure of zero, with trapped air. Trapped air bubbles change size with matric pressure in accordance with Boyle's law. They also shrink as gas dissolves or expand as it comes out of solution. The quantity of trapped air is easy to determine at saturation because all air present is trapped; measurement of volumetric water content subtracted from the porosity will indicate volumetric trapped air content. For the amount of air trapped at saturation, Mualem (1974) proposed a rule of thumb based on empirical observations, that 10% of the pore space will be occupied by trapped air. In general, this will vary with the medium, rate of wetting, water content before wetting, and other factors, and can far exceed the 10% guideline.

Transport of Nonaqueous Liquids

Liquids such as oils and organic solvents that do not easily dissolve in water are retained and transported within a porous medium in some of the same ways as water, but with crucial differences. Being a phase separate from water, they are often called *nonaqueous phase liquids*, or NAPLs. Hess *et al.*, (1992) and Essaid *et al.*, (1993) give examples of how fluid contents can be observed and measured and their distributions simulated at a site of oil contamination of groundwater.

To a first approximation, nonaqueous liquids obey the laws of surface tension and viscous flow, just like water. Liquids such as oils typically have a weaker tendency to cling to solid surfaces than does water. Thus, they are a relatively nonwetting phase, and they exist mainly in blobs or interconnected shapes within the pores of the medium, separated from the particle surfaces by a layer of water (Figure 11). With enough nonaqueous liquid present, it also can form a continuous phase across significant distances within the medium. Then its transport may be described using the permeability form of Darcy's law (9). Otherwise, the nonaqueous liquid is present mainly as isolated blobs, around which water flows as it flows around solid particles. Dillard *et al.*, (1997) show how such factors as the nature of the K distribution can affect oil transport in the subsurface.

Nonaqueous liquid transport often must be considered in combination with other modes. Many liquids volatilize significantly, so the resulting vapor moves in gaseous form. Nonaqueous liquids normally dissolve significantly and undergo transport also as a solute. These modes of transport are less affected by the degree of continuity of the nonaqueous phase than is the liquid transport mode.

Transport of Particles

Small solid particles can move through the subsurface in response to a complex set of factors. These particles include bacteria and other microorganisms as well as nonliving colloids such as rock fragments, mineral precipitates such

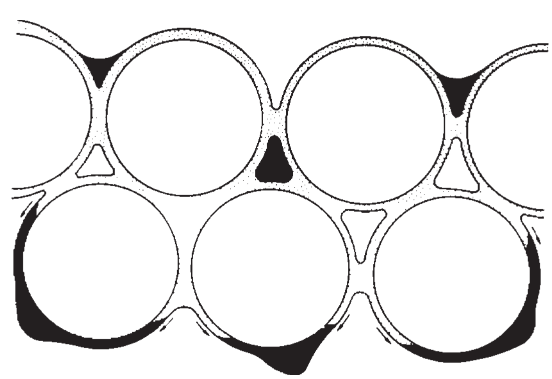

Figure 11 Four phases present in an artificial medium made of glass spheres. In this cross-sectional diagram, stippled areas are water, black areas are nonaqueous liquid, and white areas within the circles are solid glass and elsewhere are air (Reproduced from Schwille, 1988 by permission of Lewis Publishers (Chelsea, MI, USA))

as iron oxides and carbonates, weathering products such as clay minerals, macromolecular components of dissolved organic carbon such as humic substances, and microemulsions of nonaqueous liquids.

Mobile particles play several roles in practical unsaturated zone problems. (i) They themselves are sometimes contaminants. (ii) They act as carriers of any contaminants that sorb onto them, overriding the intrinsic transport characteristics of those contaminants. This complicates solute transport problems, especially when the solutes of interest adsorb onto the colloids and the colloids are more mobile than the solute. (iii) They can be naturally present or artificially introduced agents (typically bacteria) that break down organic contaminants and other compounds into other substances whose presence may be more desirable. (iv) Their presence and possible redistribution within the pore space can affect hydraulic conductivity or other transport properties.

Mechanisms of particulate transport include convection, advection, and adsorption, as for solutes, and other mechanisms (Harvey and Garabedian, 1991). Adsorption depends strongly on the composition of the medium, generally increasing with additional clay and decreasing with additional organic matter. Another mechanism of great importance is straining, the blocking of further motion by pores that are smaller than the particles themselves. For microbiological organisms such factors as death and reproduction also must normally be considered.

In unsaturated media, there are additional mechanisms that can make particle transport more complex than in saturated media (DeNovio *et al.*, 2004). The movement is strongly retarded as water content decreases, leading to a sometimes useful assumption that through the soil there is little motion except when the medium is nearly saturated. Particles can be removed from mobile fluid through mechanical filtration by films as well as by small pore spaces. Besides the reasons for water itself and solutes to move much more slowly at lower water contents, there is also a strong tendency for solid particles to adsorb onto the air–water interface (Wan and Wilson, 1994), as illustrated in Figure 12. This adsorption may be stronger than the adsorption onto the soil matrix itself. This effect also means that when the pores are all filled with water except for some amount of trapped air, that air may significantly retard the transport.

The mobility of inorganic colloids in the subsurface is controlled by chemical interactions between colloids and immobile matrix surfaces, and by hydrological and physical factors. Changes in aqueous chemistry can cause colloids to aggregate or to disaggregate. Higher ionic strengths, for example, usually favor colloid aggregation. This affects mobility because in general larger, aggregated particles are less mobile. Particles can be released into subsurface water as a result of mechanical grinding of mineral surfaces. Hydrodynamic forces associated with increases in flow rate can dislodge colloids.

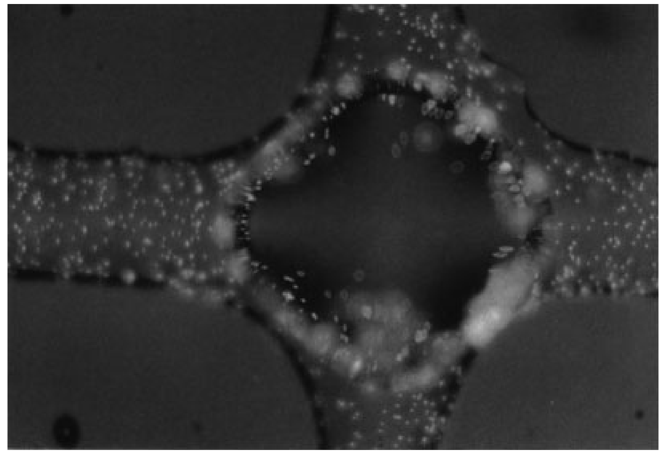

Figure 12 Microscopic cross-sectional view of hydrophobic latex particles in water in artificial pores (Wan and Wilson, 1994). The main pore body, the intersection between the two perpendicular pores, is about 200 μm across. The latex particles (light-colored dots) are 1 μm in diameter. A large air bubble occupies most of the main pore. Some of the particles are adsorbed on the air-water and water-solid interfaces (Reproduced from Wan and Wilson, 1994 by permission of American Geophysical Union). A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

DETERMINATION OF UNSATURATED CONDITIONS AND PROPERTIES

Moisture State

Water has many distinctive properties – volatility, density, molecular and nuclear structure, and electrical and thermal properties – that lend themselves to measurements indicating the amount of water present in a porous medium. The basic gravimetric method is to dry soil in an oven until the weight is constant, so that the difference between that weight and the initial wet weight indicates how much water was in the soil. This is often used as the standard for calibrating other methods.

Several methods in widespread use are minimally disruptive. X-ray or γ-ray attenuation can indicate water content in space and time: the more water, the less the beam intensity coming out of the medium. This effect is sometimes used tomographically to produce a two-dimensional map of water content in a cross section of the medium. Neutron scattering is commonly used for monitoring water content as a function of depth in the field. This method is based on the relative effectiveness of the various components of the wet soil in slowing neutrons. Because of the conservation of energy and momentum, a neutron passing through matter is slowed down most effectively by collisions with particles that are about the same size as itself. In soil, essentially the only particles that are the size of a neutron are the hydrogen nuclei in water molecules. A probe that includes both a source of fast neutrons and a detector of slow neutrons registers more counts in wetter soil. 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, to obtain measurements as a function of depth. Another way to monitor water content over a period of time in the lab or field is by measurement of the dielectric constant of the medium, usually by time-domain reflectometry (TDR) (Topp *et al.*, 1980). Liquid water has a much greater dielectric constant than other constituents of soil or rock, so this effect can indicate the amount of water present within the volume sensed. For most applications, TDR electrodes in the form of metal rods are inserted into the soil. Various geometries of electrodes are possible, including a coaxial cylinder for laboratory use with core samples. A less common method is to measure electrical conductivity (e.g. Sheets and Hendricks, 1995), which increases with water content. This principle can be applied tomographically (e.g. Daily *et al.*, 1992) for observing two- or three-dimensional details of changing water distributions in the field. Ground penetrating radar (e.g. Eppstein and Dougherty, 1998) can also be applied directly or tomographically.

The most direct measurement of matric pressure 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 larger chamber where a gauge or transducer reads the pressure (Figure 13). The key feature is the porous membrane that contacts the soil. In order to assure a continuous liquid water pathway from the pore water to the chamber water, this membrane must remain totally saturated. Thus, it must have an air-entry value beyond any ψ value to be measured, and must not be allowed to dry out. There is a basic ψ limit of about -80 kPa, below which ordinary tensiometers fail because of runaway bubble formation, though Miller and Salehzadeh (1993) have shown that devices that remove dissolved air can extend this range.

Other methods are available for media drier than -80 kPa 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) matric pressure 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 -1500 kPa matric pressure range corresponds to a 100 to 99% range in relative humidity. Psychrometers (Andraski and Scanlon, 2002) and chilled-mirror devices (Gee *et al.*, 1992) are both used to measure humidity for this purpose. Another class of methods uses an intermediary porous medium of known retention properties. Examples include

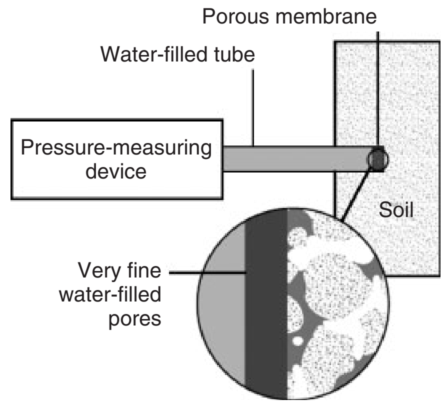

Figure 13 Schematic diagram of a tensiometer in contact with soil for matric pressure measurements. The pressure measuring device can be an electrical transducer, a manometer, or other pressure gauge. The porous membrane is often porous ceramic or sintered metal, or one of various other materials intended for use as filters. Especially in field applications, the membrane is often shaped as a tube or cup rather than a disk, to have greater contact area with the soil

gypsum blocks, nylon fabric, and filter paper (Scanlon *et al.*, 2002). This medium is placed in contact with the medium to be measured so that the matric pressure 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. Often the retention curve of the intermediary medium is not known directly, but rather a calibration relation between matric pressure and the measured quantity (e.g. electrical resistance) is known.

Water Retention

Measurement

Any system that makes independent simultaneous measurements of water content and matric pressure can indicate water retention relations. Additionally, there are methods specifically intended to measure this property. Many of these use a large porous membrane, often ceramic, to permit equilibration of water pressure between the porous medium on one side of the membrane and bulk water on the other. 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 its matric pressure. Pressure or suction chambers that do this use the principle of water-equilibration through a membrane of the sort used in tensiometers as in Figure 13. The pressure, or less commonly the volume of water, is adjusted through a planned sequence, and paired values of matric pressure and water content (one of them controlled and the other measured) represent the retention curve.

Estimation

Because various nonhydraulic properties of a medium, especially particle-size distribution, correlate in some way with water retention but are considerably easier to measure, models have been developed for estimating water retention from these. The basis is that the water retention curve of a medium depends directly on its pore-size distribution, which is related to the particle-size distribution. In a randomly packed medium, intergrain pores are expected to be larger where the particles are larger. Arya and Paris (1981) developed a widely used model based on this principle, using capillary theory and particular assumptions about the effective radii of capillaries to be associated with particular sizes. Models of this type often work reasonably well for sandy media. In general, however, the correlation between particle and pore size does not hold. For example, increasing fine-particle content, especially with clay particles, usually correlates in soils with increasing aggregation and more numerous macropores. Adding clay to sand can increase rather than decrease the number of large pores. More recent models (e.g. Rieu and Sposito, 1991; Nimmo, 1997) extend their applicability to more types of media with the use of additional information such as aggregate-size distributions.

Another way of estimating water retention without measuring it is with statistically calibrated pedotransfer functions such as the Rosetta model (Schaap, 1998). The basis for these 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 model of this type can estimate a retention curve with good statistical comparability to known retention curves of other media with similar nonhydraulic properties. The foundation in empirical data helps to compensate for such effects as the increase in large pores with increasing clay. The accuracy and reliability of these models are limited, however. They sometimes, but not always, produce acceptable estimates. Without any retention measurements for the medium in question, it is usually impossible to know if the model result is a good representation of the retention curve or not.

Hysteresis models have been developed to represent the entire set of hysteretic moisture relations from a partial data set. The most widely used is that of Mualem (1974), which yields a complete set of scanning curves given both drying and wetting main curves. Other models require greater or lesser amounts of data. The model of Nimmo (1992), for example, requires the main drying curve and two points of the main wetting curve. In general, less stringent data requirements lead to modeled results of less reliability.

Empirical Formulas for Water Retention

Whatever method is used to determine a water retention curve, it is frequently convenient to express it as a parametric empirical formula. This helps in providing interpolation or extrapolation of data, in giving a smooth, continuous form that is easy to work with mathematically, and in representation of the curve with a few (typically 2 to 5) parameter values. Among the most widely used empirical formulas are that of Brooks and Corey (1964)

$$\theta = (\theta_{\max} - \theta_{\min}) \left[\frac{\psi}{\psi_b} \right]^b + \theta_{\min} \tag{10}$$

where ψ_b, b, and $\theta_{\min}$ are fitted empirical parameters and $\theta_{\max}$ is the maximum value of θ; and that of van Genuchten (1980)

$$\theta = (\theta_{\max} - \theta_{\min}) \left[\frac{1}{1 + \left(\dfrac{\psi}{\psi_c} \right)^{\nu}} \right]^{\mu} + \theta_{\min} \tag{11}$$

where ψ_c, ν, μ, and $\theta_{\min}$ are fitted empirical parameters. Representations as simplified as these have shortcomings. For example, the Brooks and Corey curve works poorly in the wettest portion of the retention curve, and the van Genuchten in the driest. Numerous alternatives have been published, with advantages for particular applications or types of media. Some of these are based on different mathematical formulas, for example, the lognormal distribution (Kosugi, 1994), which has advantages in understandability of the significance of its parameter values. Others are formed by combining formulas by addition (Ross and Smettem, 1993) or by joining different functions that each apply in a different portion of the range. For example, Ross *et al.* (1991) developed an equation for realistic representation in the dry range, Rossi and Nimmo (1994) for the whole range from oven-dryness to saturation, and Durner (1994) for bimodality of pore-size distribution.

Dynamic Characteristics

Measurement of K and D

The most accurate measurements of hydraulic conductivity are by steady-state methods. One way 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 (Mualem and Klute, 1984). Another is to force water through at a constant and known flux density, which lets the matric pressure become uniform in part of the sample, then to compute K from the known flux density and force of gravity (Childs and Collis-George, 1950). With gravity as the main driving force, steady-state measurements are possible only for the high K values of fairly wet soil. Centrifugal force makes possible the accurate measurement of K at low water contents (Nimmo *et al.*, 2002b).

There are many techniques for measuring unsaturated hydraulic conductivity using unsteady flow. One of these is the instantaneous-profile or unsteady drainage flux method (Hamilton *et al.*, 1981). It is useful in both the laboratory and the field. This method is based on a determination, within a medium in which unsteady flow has been established, of both the flux density and the matric pressure gradient, from water contents in space and time, and from matric pressure measurements at a given instant of time after the application of water. Another alternative for laboratory applications uses flow driven by evaporation. There are various indirect and inverse methods – a wide variety of situations where available data describing water flow over time can provide information for an estimation of K or D (Hopmans *et al.*, 2002). 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 ψ values as data to calculate the unsaturated hydraulic properties. It is often implemented as an inverse method, as discussed further below. Garnier *et al.* (1997) have developed a transient flow method for the more difficult case of soils that swell as water content increases. Infiltrometer methods are useful in field applications (e.g. Smettem *et al.*, 1995; Vandervaere *et al.*, 2000a; Vandervaere *et al.*, 2000b). Water is supplied at a known or measured rate, generally through a flow-restricting membrane or crust to establish unsaturated flow, and measurements of the changing moisture state in the soil are used with the flow rate to estimate unsaturated K.

Diffusivity methods also give K if the water retention relation is known. Typically, these methods are easier but less accurate than direct K measurements. One family of techniques examines the water content distribution within a medium whose water content is changing with inflow. Another popular technique for diffusivity is the one-step outflow method (Gardner, 1956; Doering, 1965) in which a sample in a pressure chamber with an outflow membrane is suddenly exposed to a step increase in air pressure. This forces water out through the membrane as the matric pressure and water content decrease. Measurements of the rate of outflow permit an estimation of $D(\theta)$, normally to within about 1 order of magnitude. Various refinements of the method and alternative methods of calculation have been developed (e.g. Passioura, 1977; Valiantzas and Kerkides, 1990; Etching and Hopmans, 1993). Multistep techniques have also been developed in recent years (van Dam *et al.*, 1994). These resemble the one-step outflow method, but are conducted through a series of small pressure-change steps instead of one large one.

Determination of saturated K can be important in the unsaturated zone because frequently there are localized zones of saturation and also because this property is sometimes used in estimating other hydraulic properties. A straightforward method uses the flow rate measured with a constant applied head gradient, much like the analogous method for unsaturated K. It is also possible to supply water to a soil column from a source with a level that falls as the water is used up, the falling-head method for saturated K. A falling head, where the water level can be related to the cumulative flux, is more useful for saturated than for unsaturated properties because the water content does not change with the applied pressure. Saturated K is easier to measure than unsaturated K but it has unique difficulties. One is the fact that gaps between a soil sample and a retainer wall will fill with water, possibly becoming rapid-flow channels that lead to an indication of K much greater than the actual K of the soil. This is not usually a problem for unsaturated measurements because large gaps desaturate when the matric pressure is even slightly less than zero. Another problem in a saturated K measurement is that because of air trapping the soil is not really saturated and its water content may be poorly determined. For these and other reasons, measured values of saturated K frequently have poor reliability and may have little relation to observed field phenomena that would seem to occur at saturation.

To determine permeability for a gas, a straightforward flow-measuring apparatus can be used with soil samples in a laboratory for accurate results but with considerable effort (Stonestrom and Rubin, 1989a). Field methods (Weeks, 1978; Baehr and Hult, 1991; Shan, 1995) normally rely on measurements of pressure variations at points within the medium in response to pressure changes imposed elsewhere in the medium.

Estimation of K

Property transfer models can be useful for estimating K, as for water retention. 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 like that of Arya and Paris (1981), though reliability is likely to be markedly 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 K and retention. Purcell (1949) was the first to use it to quantitatively relate these two properties, using the assumption that pores are equivalent to a bundle of capillary tubes of different sizes, with a particular size distribution for a given medium. The drying retention curve relates to the effective radius at which pores empty and the capillary formula (2) quantifies this radius in terms of ψ. Because larger pores empty first, for a given ψ, pores smaller than the radius corresponding to ψ are considered filled. Poiseuille's law gives the effective conductance of a filled capillary, and an integration of such conductances for all filled pores leads to an estimate of K for that ψ value (and its corresponding θ). In effect, this is a summation of contributions from each size of filled pore, weighted by the abundance of that size, to give a number assumed proportional to K. Multiplication by a matching factor, quantified using a single known value of K, gives the $K(\theta)$ or $K(\psi)$ curve. In practice, the matching factor is often based on saturated K, though in general this is a poor choice because saturated K depends primarily on the very largest pores of the medium, which have little relevance to $K(\theta)$ over most of the θ or ψ range (Nimmo and Akstin, 1988). Choices of how the model handles issues such as pore length, connectedness, and tortuosity lead to different versions of this sort of model (e.g. Childs and Collis-George, 1950; Burdine, 1953). The version of Mualem (1976) has become widely used because it is mathematically easy to work with and gives estimates as good as or better than others. Mualem and Dagan (1978) and Hoffmann-Riem *et al.*, (1999) have developed schemes for generalizing this class of models. More recent developments include the incorporation of angular pore geometry (Tuller *et al.*, 1999), which allows a pore of given size to form a continuum of sizes of effective water flow conduit, rather than restricting it to the circular-capillary states of complete fullness or complete emptiness. Network and percolation-theory models have also been developed but have not come into widespread use.

As in the case of water retention, completely empirical formulas can represent unsaturated K. Gardner (1958), for example, used a formula of the form

$$K(\theta) = A\exp(-\alpha\theta) \tag{12}$$

where A and α 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 as described below. The α parameter in (12) is widely used in developing and applying other models, such as analytical solutions of Richards' equation.

Combined Estimation of K and Water Retention

A direct combination of an empirical formula for water retention 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. For example, the Brooks and Corey (1964), van Genuchten (1980), or lognormal (Kosugi, 1999) model can be inserted into the Burdine (1953) or Mualem (1976) model. The combination of the van Genuchten and Mualem models results in the formula

$$K(\psi) = K_{\mathrm{m}}\frac{\left\{1-\left(\dfrac{\psi}{\psi_c}\right)^{\nu-1}\left[1+\left(\dfrac{\psi}{\psi_c}\right)^{\nu}\right]^{-\mu}\right\}^2}{\left[1+\left(\dfrac{\psi}{\psi_c}\right)^{\nu}\right]^{\mu/2}} \tag{13}$$

where K_{m} is the matching factor computed from a known $(\psi,\ K)$ point. Leij *et al.* (1997) give a tabulation and evaluation of formulas that result from such combinations. Use of such formulas is central to pedotransfer-function models for coordinated estimation of water retention and K. As in the water-retention application of pedotransfer functions described above, these provide a particularly easy way to estimate the hydraulic properties needed for a Richards-equation analysis, though likewise with problems of poor and unknown reliability.

Translation of parameter values from one model to another, that is, for a given medium finding the values for one model's parameters directly from the parameter values previously estimated for a different model (Morel-Seytoux *et al.*, 1996; Morel-Seytoux and Nimmo, 1999), is a convenience where the retention curve is known only in terms of one particular parameterization, and an application requires another. In order to give a unique conversion, a specific equivalence criterion must be chosen. This can, for example, be the invariance of a relevant property like capillary drive (a single-number combination of retention and conductive properties related especially to infiltration rates).

Inverse approaches can yield hydraulic property values in a variety of situations and experiments, as explained in detail in **Chapter 156, Inverse Methods for Parameter Estimations, Volume 4**. The basic idea is that a forward model (e g. Richards' Equation), which can compute output conditions (θ or ψ as a function of space and time) using given properties (water retention and unsaturated K), can be used in an algorithm that does the inverse of this. That is, given a set of measured output conditions, the algorithm determines values of the properties that would most closely simulate the measured conditions. Those values then have effectively been determined, and can be used in the forward model. Conditions such as water content as a function of space and time frequently are easier to measure than the properties of the medium. One way of inverse modeling is to use a forward model with manual or automated trial-and-error: guess the property values, compute outputs, compare with observation, revise estimated property values, and try again. Many particular field or lab situations, for example, the one-step or multistep outflow method, are suitable for unsaturated hydraulic property estimation by inverse approaches (Hopmans *et al.*, 2002). Frequently, the transient ψ values from one or more tensiometers installed in the sample are used as data in addition to the outflow rate as a function of time.

Scaling provides another way of determining unsaturated properties from limited data, relating them from place to place or from one medium to another (Miller, 1980). In a wet medium, surface tension is the main phenomenon associated with retention, and viscosity with conductivity. These are drawn together in the surface-tension viscous-flow

(STVF) similitude model of Miller and Miller, (1956). One way for two media to be similar is for one of them to be geometrically a direct magnification of the other by a scale factor λ. For natural media, the point-to-point correspondence is not exact; the particles of Miller-similar media can be repositioned, and they do not have to be exact scale replicates, as long as they are statistically equivalent for the purpose at hand. If two media are Miller-similar, if surface tension controls water retention, and if viscous flow laws determine K for a given geometrical configuration, the unsaturated hydraulic properties of one medium can be directly determined from those of the other using the known value of λ. The value of ψ in one medium differs by a factor of λ from that in the other. Water content, being a ratio of volumes, is the same in each. Thus, the retention curve $\theta(\psi)$ of one similar medium equals $\theta(\lambda\psi)$ of the other. Because of the radius-squared factor for conductivity in Poiseuille's law, $K(\theta)$ for similar media is different by a factor of λ^2. Warrick *et al.*, (1977) relaxed some of the Miller–Miller criteria, opening up new field applications by showing the relaxed criteria to be useful for relating properties from one sample to another. Porosity, for example, does not have to be identical for Warrick-similar soils. Whenever a set of media can be considered similar with known values of λ, knowledge of hydraulic properties for one of the media permits calculation of the properties of all. Recent applications of this sort include those of Rockhold *et al.*, (1996) and Nimmo *et al.*, (2002a).

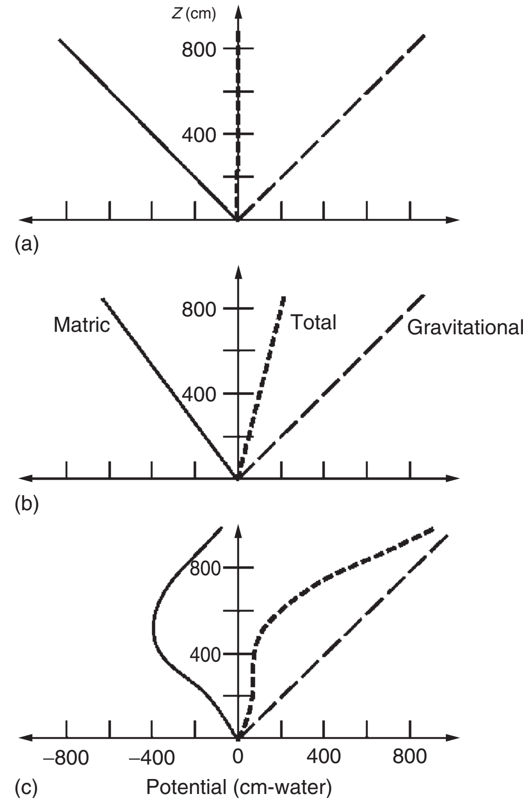

Figure 14 Profiles of matric, gravitational, and total potential for idealized situations of (a) static water, (b) steady downward flow, and (c) unsteady flow. The water table is at $z = 0$

APPLICATIONS TO DISTRIBUTIONS AND FLUXES

This article emphasizes the presence and movement of water within the unsaturated zone. **Chapter 146, Aquifer Recharge, Volume 4** considers the fluxes from the unsaturated zone into the saturated zone. **Chapter 152, Modeling Solute Transport Phenomena, Volume 4** and **Chapter 153, Groundwater Pollution and Remediation, Volume 4** consider the fluxes of other substances, for the evaluation of which the flux of water is usually the first consideration.

Distributions of Water and Energy

The distribution of water with depth at a given time depends on the energy state (based on components including 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 14(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 the water content is controlled by the water retention properties of the medium, for a uniform unsaturated zone, the water content profile (not shown in Figure 14) then mimics the shape of the water retention curve. The previous history of the unsaturated zone would dictate whether a wetting or drying retention curve would apply. In general, there is significant trapped air when the medium is close to saturation; thus, it frequently is appropriate to consider a repeat-cycle curve rather than a first drying curve. Given time, approximate versions of such a hydrostatic profile may develop in portions of a profile where 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 14(b). At many locations this situation probably never develops, but some locations, especially where the unsaturated zone is thick, may commonly have constant potential or an approximation to it. Darcy's law applies directly in field situations where flow is steady, though steadiness in nature is often temporary or approximate. In the general case of unsteady flow, the matric pressure profile cannot be determined so simply, and may take on an irregular form like the example in Figure 14(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 two or three days after a major infiltration (SSSA, 1997). An essential assumption for use of the field capacity concept is that soil, when wetted to a high water content, loses water by drainage at a declining rate that does become negligible. This concept is used in agriculture to indicate the wettest soil conditions that need to be considered for plant growth, and sometimes is mentioned in hydrologic investigations, for example, as related to soil moisture storage. Elements of the definition of field capacity are imprecise and require subjective judgment, for example, in deciding what 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. Thus, it is not appropriate to consider the field capacity of an individual soil sample or a point within the unsaturated zone. Because it is based on distinct intervals of major infiltration and major drainage, the concept of field capacity is most applicable in field plots that are periodically wetted to a high degree, as by annual snowmelt, flood, or monsoons; especially in applications where only approximate quantification or rules of thumb are required. This concept does not work well in rocky or fracture-dominated media that retain significant water only briefly, on land where water application is consistently erratic, or in applications requiring exactness, repeatability, or validity at multiple scales of observation.

In a portion of the unsaturated 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. As in other cases of saturation within the unsaturated zone, there is likely to be significant trapped air, though not an air phase that is continuous through the medium. Some media will not have a significant capillary fringe because their retention characteristics have the air-entry pressure at essentially zero. These media, including materials with significant fractures or other macropores, are more common than is apparent from databases and literature surveys of measured water retention, because, historically, most measurements have been done on samples that have been repacked from mechanically disturbed material. Repacking often destroys the large pores that would otherwise lead to a zero air-entry pressure. Caution must also be invoked in applying the capillary fringe concept where the water table fluctuates. The hydrostatic equilibrium required for a capillary fringe may take considerable time to establish, given typically small values of unsaturated K. Furthermore, soil-water hysteresis would make for a different capillary fringe with a falling than with a rising water table.

Fluxes at the Land Surface

Input

Infiltration is the downward movement of water through the land surface. Because osmotic, thermal, and other modes of driving water flow are usually negligible, water is driven into the soil mainly by gravity and ψ gradients. If the soil is dry near the surface the ψ gradients usually dominate gravity. When the soil is very wet to some depth, gravity may become the major driving force. 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 grows. Figure 15 shows actual infiltration rates i $[\mathrm{L\,T^{-1}}]$ varying over time in four soil columns (Hillel and Gardner, 1970). If water at the surface is abundantly available but not under significant pressure, infiltration occurs at the infiltration capacity, a rate determined only by the soil, not by the rate of application or

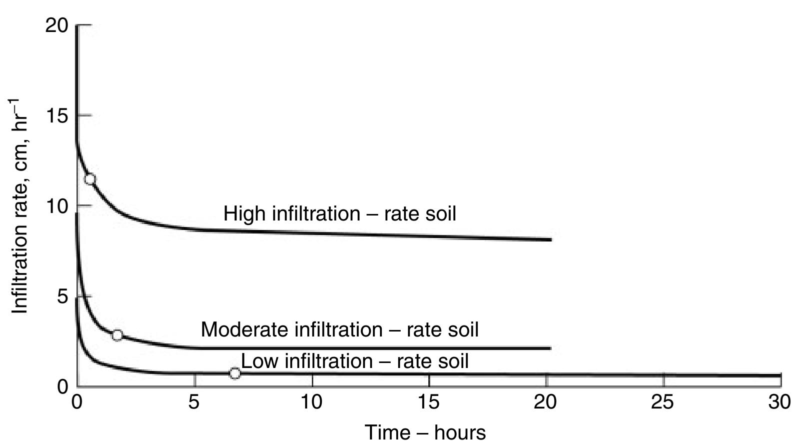

Figure 15 Measured infiltration rates over time for three soil columns

other factors. If water arrives at the 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. If water arrives at a rate less than the infiltration capacity, the water may immediately infiltrate. The infiltration rate then is determined by the rate of application. This article addresses the issue of what happens when the soil limits the infiltration rate, requiring attention to the infiltration capacity not only as the key element for quantifying i under that condition but also as the criterion that determines whether that condition exists. 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.

Richards' equation (5) can be applied to infiltration processes. But the phenomena involved, especially with water contents, pressures, and fluxes sharply changing in time and space, make this difficult or inappropriate. Most studies use instead a formulation specifically tailored for infiltration, generally with more specific assumptions. Some of these formulations are based on a conceptualization of the physics of infiltration while others are more purely empirical.

The Green and Ampt model of infiltration assumes uniform properties of the medium, a perfectly sharp wetting front at depth L, a constant matric pressure ψ_o at the land surface, a uniform initial water content θ_i, a uniform water content θ_w behind the wetting front, and a value of matric pressure ψ_{wf} at the wetting front that remains fixed even as the front moves downward. Gravity may be included, or, in a simplified form of the model, assumed negligible. The mathematical formulation of the model comes from expressing the infiltration rate in two different ways – by Darcy's law and by the computed change in water storage – and setting them equal to each other. This leads to the depth of the wetting front being

$$L(t) = \left(2K\frac{\Delta\psi}{\Delta\theta}t\right)^{1/2} \tag{14}$$

and the infiltration rate

$$i = \Delta\theta\frac{\mathrm{d}L}{\mathrm{d}t} = \Delta\theta\left(\frac{K\dfrac{\Delta\psi}{\Delta\theta}}{2t}\right)^{1/2} \tag{15}$$

where $\Delta\theta = \theta_w - \theta_i$ and $\Delta\psi = \psi_o - \psi_{wf}$. Analogous formulas can be derived for the case where gravity is not negligible. The inverse proportionality of i to the square root of time appears frequently in infiltration formulas. The natural decline of the infiltration rate as the soil gets wetter correlates in the model with the declining driving force as the constant $\Delta\psi$ is stretched over an increasing depth interval. Mathematically, this works out to square-root-of-time behavior. The Green–Ampt model gives a direct quantification of infiltration with simple formulas for the basic character of infiltration under common circumstances. But, of course, it does not give specific details of the infiltration process. Generally, one must infer a value of ψ_{wf} from other known facts concerning the infiltration; the fact that ψ_{wf} cannot be measured severely limits the use of this model in a predictive manner. The Green–Ampt assumptions are most likely to be satisfied in uniform, coarse-textured soil with simple structure.

Philip (1957) presented mathematical solutions to the problem of infiltration under fewer restrictions than required by Green and Ampt. Philip assumed that the medium properties and the initial water content are uniform, and that the water content at the land surface is held at a fixed water content greater than the initial water content. Starting with a form of Richards' equation, he derived

$$i = \frac{1}{2}\mathrm{S}t^{-1/2} \tag{16}$$

where S is a soil hydraulic property called the *sorptivity* (dimensions $\mathrm{L\,T^{-1/2}}$), which indicates the relative tendency for the medium to absorb infiltrating water. For the case where gravity is significant, Philip developed a series solution that can be approximated for short times by the truncation

$$i = \frac{1}{2}\mathrm{S}t^{-1/2} + A \tag{17}$$

where A is an empirical parameter whose value approaches saturated K after a long time. Philip's model has been tested against measurements, notably by Davidson *et al.* (1963), who found good agreement with the model in terms of both preservation of the shape of the wetted profile, and of the rate of its downward movement. Yet, it would be exactly true only for certain unlikely restrictions on the unsaturated hydraulic conductivity and the ponding of infiltrating water. Furthermore, most cases of infiltration are in heterogeneous soils with nonuniform water content and variable water application, in which the Philip model would give approximate results at best.

Empirical formulas for infiltration have been used where the relevant soil hydraulic properties are not necessarily known. Gardner and Widtsoe (1921) used an exponential formula expressed by Horton (1940) as

$$i = i_c + (i_o - i_c)\exp(-\alpha t) \tag{18}$$

where i_o is the initial infiltration rate, i_c is the rate of infiltration after it becomes steady, and α is an empirical constant that depends on the soil. Holtan (1961) proposed an empirical formula useful for the early stages of infiltration, based on a power law function of the effective yet-unused water-holding capacity of the upper part of the soil profile:

$$i = i_c + a\left[ML_l - \int_0^t i\,dt\right]^n \quad (19)$$

where i_c, a, and n are constants, and M is the air-filled soil porosity down to a specified depth L_l (ideally to a known impeding layer).

Output

The term exfiltration is occasionally used to indicate water fluxes out of the soil at the land surface, though mostly these processes are discussed in terms of evapotranspiration. As a whole, evapotranspiration is usually treated as a surface or micrometeorological problem, though certain critical unsaturated-zone processes are involved, especially for the evaporation, or bare-soil, component.

When the soil is wet enough, plenty of water is available at the surface so that atmospheric conditions control the evaporation rate. When the soil is too dry to supply water at the maximum rate the atmosphere will absorb it, 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. If the second phase persists long enough, the soil may become so dry that water transport within it is primarily in the form of vapor rather than liquid. This defines a third phase, in which the evaporation rate is slow and may be nearly constant. The initial atmosphere-dominated phase typically lasts a few hours to a few days after irrigation or rainfall. As with infiltration, there is a moisture state criterion based on local conditions and properties that separates atmosphere-dominated and soil-dominated flow. Gardner and Hillel (1962) have quantitatively treated the question of when the atmosphere-dominated phase ends. The overall reduction of evaporation rate with time can be considered in terms of self-mulching; after a bare-surface soil has dried to a certain extent, a layer with very small θ forms at the surface. The low hydraulic conductivity of this layer limits the rate of flow of water from the soil below it to the land surface, thus limiting evaporation (Buckingham, 1907).

During the soil-dominated phase, to roughly estimate the average evaporation rate, the problem can be approached with Richards' equation, assuming that the system is isothermal, vapor does not flow, gravity is negligible, diffusivity is constant, and the soil is a semi-infinite slab uniformly wet at the beginning. The problem then falls into the same class as the gravity-free infiltration problem. The solution for evaporative flux density goes as the inverse square root of time, just as in typical assessments of infiltration rate.

Water resource applications often require values of total evapotranspiration, including treatment of atmospheric conditions in addition to the unsaturated-flow phenomena considered here. Infiltration minus evapotranspiration is often taken as the net input of water to the subsurface. A second major hydrologic application is the evaporation of water from the saturated zone. Capillary forces can draw water up from the water table to depths from which it supplies the evaporative process. This can be a substantial loss mechanism from a water-table aquifer, especially where the unsaturated zone is thin. Various quantitative treatments of these effects have been developed, for example, by Ripple *et al.* (1972).

Fluxes Within the Unsaturated Zone

Redistribution

After water has infiltrated, it redistributes, driven by gravity, matric pressure gradients, and possibly other forces. Figure 16 illustrates water content distributions at various times during and after infiltration, in a mechanically disturbed soil and in a soil with intact natural structure. Redistribution continues until conditions are such that 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, that is, equilibrium.

In an idealized case of homogeneous soil and perfectly sharp wetting fronts, the redistributing water content profile may take the shape of an elongating rectangle. If there are no losses to evaporation or anything else, the rectangle will maintain the same area as it evolves. This picture of redistribution may be approximated to various degrees in real media, for example, as in Figure 16(d). Many factors cause deviations from the ideal, including layers that retard the flow, and preferential flow and lateral heterogeneity that advance the downward movement of a portion of the water, thereby blunting the rectangular sharpness of the wetting front. Greater K above than below the wetting front may cause water to accumulate in the lower portions of the wetted zone, distorting the rectangle laterally.

Normally, hysteresis strongly influences redistribution because the wetting front progresses downward according to the wetting curves of water retention and conductivity, whereas θ in the upper portions of the wetted zone decreases according to the drying curves. Because a drying retention curve has greater θ for a given ψ, 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.

After unsteady infiltration, a pattern of variations in water content with depth becomes established as alternations of surface wetness and surface dryness move downward by gravity. Matric pressure gradients move water from wet to dry, both upward and downward, so as the pattern of θ variation moves downward, wet zones become drier and dry zones wetter. Thus, variations are damped out with depth.

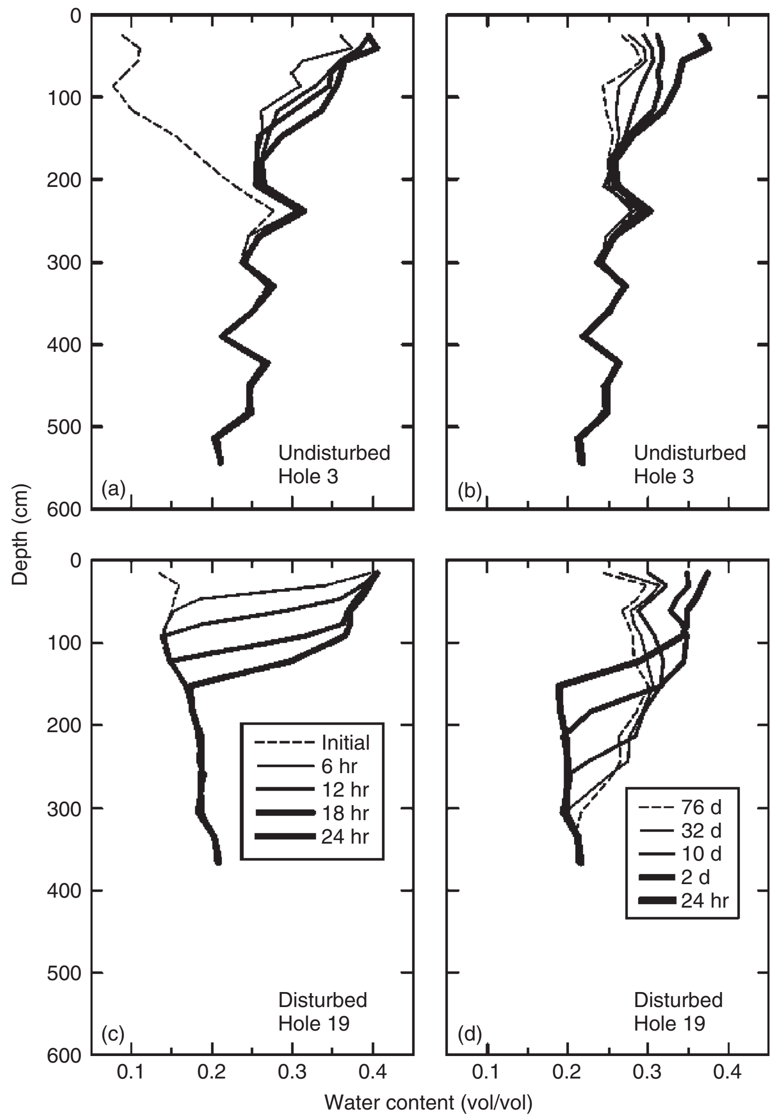

Figure 16 Measured water distributions during and after 24 h of flood infiltration in (c,d) a disturbed, relatively homogeneous soil in a landfill and (a,b) an adjacent undisturbed soil, on the snake river plain in Idaho (Nimmo *et al.*, 1999). Evaporation was inhibited by an impermeable cover at the land surface

Gravity takes on greater relative importance as the variations decrease. In a thick unsaturated zone, a region may develop in which gravity is the only significant driving force, as illustrated in Figure 2 of **Chapter 146, Aquifer Recharge, Volume 4**. Approximating the periodic infiltration of water with a sinusoidally varying flux, Gardner (1964) estimated the depth at which θ variations would become negligible. For a periodicity of one year and typical values of D, the profile would become damped within a few meters of the surface. An important application of this is in Darcian methods for estimating aquifer recharge rates as discussed in **Chapter 146, Aquifer Recharge, Volume 4**.

Usually the above considerations would need to be adjusted or reinterpreted with attention to preferential flow. There is not yet a widely accepted quantitative theory for this. 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.

Effects of Layering

Layers that contrast in hydraulic properties impede vertical flow by various mechanisms. (i) 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. (ii) Where a fine layer overlies a coarse layer, water moving downward is impeded under many conditions. When coarse material is dry, it 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, as discussed previously. Stable or not, water flow into the pores of the coarse medium increases that medium's hydraulic conductivity. With equal ψ values across the layer boundary, which is necessary for water not to accumulate at the boundary, unsaturated K of the coarse layer is often less than that of the fine layer. 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. Miller and Gardner (1962) demonstrated this effect experimentally.

Layer thickness can influence the long-term tendency of downward flow in other ways. An important case is where a granular medium overlies fractured bedrock. If the fractures are not microscopically narrow and the rock is otherwise impermeable, the bedrock admits water only under nearly saturated conditions. The thinner the layer of granular material, the more easily it becomes saturated down to the depth of the bedrock and hence the more frequently it generates deep percolation. This more frequent saturation at the layer boundary increases the fraction of average precipitation that flows into the bedrock and possibly into the aquifer as recharge.

A common phenomenon in layered media is the accumulation of water in a region of the unsaturated zone to the point where it becomes saturated, even though there is unsaturated material between that region and the saturated zone. Because this phenomenon usually results from an impeding layer on which excess water is perched, it is called *perching*. It may be a temporary or a nearly permanent feature, depending on the nature of the medium, the prevailing hydrologic conditions, and the effect of artificial modifications. 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 because the increase in effective vertical hydraulic conductivity is offset by a diminished vertical hydraulic gradient within the perched water. (Vertical flow within and below the perched water cannot be faster than the vertical flow above the perched water or the perched water would have drained.) Horizontally, however, there may be greatly increased flow (e.g. Nimmo *et al.*, 2002c). New and different processes may significantly affect contaminant transport in a perched zone. Reduced aeration, for example, may affect biochemical processes. At the scale of the entire stratified vadose zone, perching may significantly increase anisotropy. For considerable horizontal distances the hydraulic conductivity might be as great as $1\,\mathrm{cm\,s^{-1}}$ or more, as for a saturated gravel. At the same time, however, vertical flow might be limited by an unsaturated layer having vertical hydraulic conductivity 10 or more orders of magnitude less, as for rock with unfilled fractures.

Layering strongly influences the direction of flow within the unsaturated zone. The flow is often assumed to be predominantly vertical because (i) with a continuous air phase in the pores, buoyancy does not counteract gravity as it does in the saturated zone, and (ii) whatever the effect of other forces, gravity acts vertically. Some vertical flow continues even where impeding layers cause substantial perching and horizontal diversion. Horizontal flow can be especially important in sloping or laterally heterogeneous media. Horizontal flow under nearly saturated conditions, called *interflow*, may be substantial near the land surface during storms. In thick unsaturated zones where adjacent layers contrast sharply, horizontal flow can have great importance even in arid climates.

CONCLUSION

Unsaturated flow is complicated by the significance 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 proportion of water and air can cause a property like hydraulic conductivity to vary over many orders of magnitude.

Many of the features that make saturated zone hydrology difficult – for example, opacity of the subsurface, heterogeneity, complex geometry on small and large scales – are all the more influential when properties vary so drastically, for example, K at a given point in space varying by orders of magnitude. As a result of these complications, unsaturated zone hydrology is subject to a greater degree of inexactness than most fields of quantitative physical science. Often, even factor-of-10 accuracy is difficult to achieve. This means that in many presentations of analyses and model predictions, the precision of the mathematical results goes far beyond what is justified by sparseness or poor quality of the data that support those results, and also beyond what is justified by applicability of the models and concepts. Evaluation of uncertainty depends much on hard-to-quantify conceptual uncertainty.

Given only 100 years since quantitative physical theory was first applied to unsaturated flow, understanding has developed considerably. In the early twenty-first century, it is still developing rapidly and maturing. Further advancement in response to scientific and societal needs requires new measurement techniques to obtain more and higher quality data, and new theoretical constructs that more adequately represent the important physical processes within practical modeling schemes.

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