The accurate knowledge of hydraulic properties for unsaturated soils is critical in addressing problems in a variety of disciplines such as hydrology, ecology, environmental sciences, soil science, and agriculture. The purpose of this paper is to review the characterization of unsaturated soil hydraulic properties for their applicability in models simulating unsaturated water transport. In a theoretical section, we present the fundamentals for the definitions of the hydraulic properties in the framework of the continuum theory as well as provide the common parameterizations of the hydraulic functions. The characterization and parameterization of hysteresis and the phenomenon of dynamic effects in hydraulic properties are addressed in subsequent sections. Finally we discuss issues related to the handling of spatial and temporal variability, including geostatistical characterization, upscaling, and scale dependency of effective properties. The review closes with a summary on limits and opportunities for modeling water transport with Richards’ equation.

INTRODUCTION

The accurate characterization of the hydraulic properties for unsaturated soils is critical in addressing many problems in hydrology, ecology, environmental sciences, soil science, agriculture, and other disciplines. Knowledge of the hydraulic properties is required in nearly all basic and applied aspects of soil, water, nutrient, and salinity management research (van Genuchten et al., 1999). This chapter describes how hydraulic functions are defined in the framework of the continuum theory by averaging the microscopic phase properties (see Section “What are soil hydraulic properties?”). The hydraulic properties must be parameterized in closed-form expressions for their later application in simulation models. In Section “Parameterization of hydraulic functions”, we discuss common parameterizations and illustrate in particular some problems and pitfalls associated with the widely used van Genuchten/Mualem model (van Genuchten, 1980). It is common knowledge that the hydraulic properties of soils are hysteretic in nature, which is especially significant with regard to transient water and solute transport. Nevertheless, there still exists very little standard representation of hysteresis in numerical models and it continues to remain a poorly modeled phenomenon. Section “Hysteresis” briefly reviews the current state of knowledge with respect to the parameterization and incorporation of hysteresis into numerical models.

Frequent observations on water flow phenomena in unsaturated soils under transient conditions show that the concept of uniquely occurring soil hydraulic properties must be questioned, since the relationship between matric potential and water content may depend on boundary conditions. Section “Dynamic effects” reviews the history of empirical evidence for these so-called dynamic effects and presents approaches on how to deal with them in numerical modeling. The final part of the paper, Section “Variability”, discusses the impact of soil variability on the definition and measurement of effective hydraulic properties for large-scale applications. We show how geostatistical methods in combination with the concept of “scaling” are used to handle the variability of unsaturated hydraulic properties and we further illustrate the difference between mean and effective hydraulic conductivity in an example.
WHAT ARE SOIL HYDRAULIC PROPERTIES?

Unsaturated soils contain two fluid phases, namely, soil water and soil air, and an immobile phase, the soil matrix. In an universal sense, “soil hydraulic properties” describe macroscopic relations between the chemical potential, the phase concentration, and the transmission behavior of water and gases in soil. These relations depend on a multitude of factors, including temperature, pore space geometry, surface properties of the soil matrix, chemical composition of the soil solution, and properties of the wetting and non-wetting fluids that occupy the complementary parts of the pore space. Two-phase flow systems have historically been described using fluid pressures and volumetric saturations as primary variables. The capillary pressure, defined as the difference between fluid phase pressures under equilibrium conditions, is related algebraically to the saturation by a nonlinear relationship, which is usually highly hysteretic. Fluxes of water and air are related to the potential gradients of the respective phases. The coefficient of proportionality between flux and gradient involves the relative permeability, which is a nonlinear function of saturation. Macroscopic mass balance equations, augmented by these constitutive relationships, represent the state of the art in mathematical descriptions of two-phase flow in porous media (Held and Celia, 2001).

Representative Elementary Volume and Measurement Window

On a microscopic scale, water content is either one (in the water phase) or zero (else). To come to a reasonable definition of water content, we must average the phase density over a certain soil volume. A practical choice is to choose the smallest volume that contains all structural elements of the pore system in sufficient abundance. We call this minimum volume representative elementary volume, REV (Hubbert, 1956; Bear, 1972). In fine-textured nonstructured soils, such as fine sands or loess, volumes of 1 cm³ or less might be sufficiently large to yield a reliable average for soil porosity. Buchter et al. (1994) estimated the REV of the porosity of a Rendzina with >50% stones to be 1.5 dm³. In soils with shrinkage cracks, such as vertisols, the REV is as large as the whole soil profile. Since large structures will often extend preferentially into one spatial direction, it follows that the size of the REV will vary in different spatial directions. In general, the REV for repacked samples, which contain the soil fraction with particle diameters less than 2 mm, will be much smaller than the REV of undisturbed samples, since stones, worm channels, roots, cracks, and lenses are large structures and must be contained in an REV in sufficient numbers. The question as to how well and accurate we can measure hydraulic properties will depend on the relation of the REV scale to the size of the measuring window.

If we extend the REV concept to characterize structural properties of natural soil, we see that no single REV exists. Enlarging the averaging volume will lead again and again to the inclusion of new structural elements of larger size (Figure 1). If, on the other hand, we extend the sampling region keeping the sample volume constant, we would observe that the average variability increases continually. This shows that observed variability of a measured soil hydraulic property will depend on the size of the measuring window, on the size of structural elements of the system, and on the extent of the region of sampling. In other words: the question whether a property is “homogeneous” or “heterogeneous” depends on one hand on the extent and the structure of the system under consideration, and on the other hand, on the averaging volume. This implies that an adequate sampling volume must be adapted to the state property to be measured, and also to the system to be described.

Continuum Approach, Darcy’s Law, Richards’ Equation

Averaging the local phase fractions or phase properties in an REV, and mapping the resulting value to the central

![Figure 1](http://www.mrw.interscience.wiley.com/ehs)
point of the averaging volume leads to the definition of spatially continuous variables such as water content, air content, soil density, porosity, or water potential. The state variable “water content” thus represents the hydraulic condition of the soil at any given time and location as a result of hydraulic processes, which in turn are governed by soil hydraulic properties. This “continuum approach” (Cushman, 1984) enables us to relate in a mass balance the temporal changes of water content at a point to spatial gradients of the water flux, here noted in a one-dimensional form

\[
\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} + s \tag{1}
\]

where \(\theta\) is volumetric water content \([L^3 L^{-3}]\), \(q\) is volumetric water flux \([L^3 L^{-2} T^{-1}]\), \(t\) is time \([T]\), \(z\) is a spatial coordinate \([L]\), and \(s\) is a source/sink term \([T L^{-1}]\). The water flux is driven by a gradient of the water pressure, \(p_w\),

\[
q = -K^*(\theta) \frac{\partial (p_w - \rho g)}{\partial z} \tag{2}
\]

where \(K^*\) is the hydraulic conductivity \([L^3 T M^{-2}]\) and \(p_w\) is water pressure \([ML^{-1} T^{-2}]\), \(\rho\) is the density of water \([ML^{-3}]\), \(g\) is the gravitational acceleration \([LT^{-2}]\). Water pressure is an expression of water potential in units of energy per volume (Durner and Or, 2005; see Chapter 73, Soil Water Potential Measurement, Volume 2). The total water potential, \(\psi_{total}\), is affected by the position in the gravity field (gravitational potential, \(\psi_g\)), by forces due to water sorption on soil particle surfaces and capillary forces in the soil pore system (matric potential, \(\psi_m\)), by dissolved substances in the soil solution (osmotic potential, \(\psi_o\)), by forces due to soil swelling and overburden (overburden potential, \(\psi_\Omega\)), and by forces that result from air pressure acting on water–air interfaces (pneumatic potential \(\psi_p\)). If a pore system is rigid, nonswelling, isotropic, and osmotic gradients and flow resistance of the nonwetting fluid are negligible, isothermal soil water movement is governed just by gradients in the matric, and the gravitational potential (Jury et al., 1991; Kutilek and Nielsen, 1994).

In soil physics, it is customary to express the water potentials in units of energy per weight, that is, as heads [L]. Neglecting the pneumatic, osmotic, and overburden components of the water potential, and considering that the gradient of the gravitational potential is equal to \(-1\) (the direction of the spatial coordinate taken positive downward) leads to

\[
q = -K(\theta) \left( \frac{\partial h}{\partial z} - 1 \right) \tag{3}
\]

where \(h\) is the pressure head \([L]\) and \(K(\theta)\) is the hydraulic conductivity \([L^3 T^{-1}]\). Equation (3) expresses the Darcy–Buckingham law (Darcy, 1856; Buckingham, 1907) and defines the saturated/unsaturated hydraulic conductivity. It implies that the hydraulic conductivity is a system property that can be determined only by an inverse approach, that is, by matching its value to be consistent with an observed dynamic system behavior (Durner and Lipsius, 2005; see Chapter 75, Determining Soil Hydraulic Properties, Volume 2). For saturated water flow, \(K\) will not depend on \(h\), and equation (3) is linear. This is contrary to the unsaturated case, and makes it simple to invert the equation analytically by simple rearrangement.

Inserting equation (3) in equation (1), and replacing the dependency of \(K\) on \(\theta\) by \(h\) leads to the pressure-form of the one-dimensional Richards’ equation

\[
C(h) \frac{\partial h}{\partial t} = \frac{\partial}{\partial z} \left[ K(h) \left( \frac{\partial h}{\partial z} - 1 \right) \right] + s \tag{4}
\]

where \(C(h)\) is the specific water capacity \([L^{-1}]\), defined by the change of water content with pressure head, \(C = \partial \theta / \partial h\). Equation (4) was first derived by Richards (1931) and is the fundamental model for describing water flow in the unsaturated zone on the macroscopic scale (i.e. the scale where we usually measure soil hydraulic properties). The model is completed by appropriate initial and boundary conditions, and knowledge of the coefficients \(C(h)\) and \(K(h)\). Since \(C\) and \(K\) are nonlinearly dependent on \(h\), solution of equation (4) requires generally numerical methods. The neglecting of the pneumatic potential implies that equation (4) is valid only if gradients in the pressure of the nonwetting fluid are negligible, or – in other words – if air is free to move in the soil at any system state. This is not always the case, neither in nature nor in measurement experiments. Equation (4) is frequently used as a process model for water transport at much larger scales. The coefficients \(C\) and \(K\) are then used as effective properties, which have the same names as for the local scale. However, their values are no longer consistent with the local, REV-based definition, as shown in Section “Variability”.

In analogy to a diffusion process, water transport can also be described by

\[
\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[ D(\theta) \left( \frac{\partial \theta}{\partial z} \right) \right] - \frac{\partial K(\theta)}{\partial z} + s \tag{5}
\]

where \(D\) is the water diffusivity \([L^2 T^{-1}]\). For horizontal water transport and without sinks or sources, equation (5) is formally equal to Fick’s second law of diffusion. Comparison of equation (4) and (5) shows that the water diffusivity is given by the ratio of conductivity and specific water capacity,

\[
D(\theta) = \frac{K(\theta)}{C(\theta)} \tag{6}
\]

The coefficient \(D\) is highly water content-dependent and expresses a propagation velocity for water content changes.
It is highest for low damping (low capacity) and high hydraulic conductivity, which is the case for conditions near water saturation. Equation (5) is less frequently applied than the pressure head form of Richards’ equation, (4), because its use is restricted to purely unsaturated soils, and, for vertical transport, the knowledge of the hydraulic conductivity function is also required. Therefore, our discussion will be restricted to the water retention curve (WRC) and the conductivity curve.

**Retention Curve and Hydraulic Conductivity Curve**

In the remainder of this chapter, we will use the term “soil hydraulic properties” for the constitutive relationships \( \theta(h) \) and \( K(\theta) \), or \( K(h) \), as defined by equations (3) and (4) in the framework of the continuum approach. The relationship \( \theta(h) \) is called water retention curve, WRC (Also, “water retention characteristic”, “soil moisture characteristic”, “capillary pressure – saturation relationship”, “water characteristic curve”, “water content – matric potential curve”, “pF curve”). In principle, it can be determined by monitoring simultaneously the state variables \( h \) and \( \theta \) at an identical point in space during a hydraulic process that changes the systems state. The dependence of the hydraulic conductivity, \( K \), on water content is called hydraulic conductivity curve. It can be determined from simultaneous flux and head gradient measurements by inverting equation (4), or more simply, equation (3).

A multitude of empirical investigations lead to general expectations about the shapes of the relationships \( \theta(h) \), \( K(h) \) and \( K(\theta) \) for different soils (Carsel and Parrish, 1988; Schaap, 2005; see Chapter 76, Models for Indirect Estimation of Soil Hydraulic Properties, Volume 2). Figure 2 shows examples of hydraulic properties for a sand, a silt, and a clay. The depicted functions have been derived by a neural network prediction with the program Rosetta (Schaap et al., 2001). They are based on many measurements documented in the UNSODA soil properties database (Nemes et al., 2001) and follow the parameterization of van Genuchten/Mualem (see Section “Parameterization of hydraulic functions”). It is important to note that these “typical” shapes of hydraulic properties are largely based on measurements on sieved, repacked samples. Compared to undisturbed soils, significant differences must be expected, in particular in the range near saturation, because of the structural pore system (Durner, 1994). For an illustration, see Figure 1 in Durner and Lipsius (2005); see Chapter 75, Determining Soil Hydraulic Properties, Volume 2.

**PARAMETERIZATION OF HYDRAULIC FUNCTIONS**

**Water Retention Functions**

For use in simulation models of unsaturated water transport, the constitutive relationships must be expressed in a continuous way over the whole moisture range, from dryness to saturation. This requires interpolation and smoothing of measured data, and it is common to express these interpolations by parametric functions. Numerous expressions have been proposed to describe the water retention characteristic (Brooks and Corey, 1964; King, 1965; Brutsaert, 1967; Visser, 1968; Taylor and Luthin, 1969; Laliberte, 1969; Farrel and Larson, 1972; Ahuja and Swarzendruber, 1972; Rogowski, 1972; Campbell, 1974; Su and Brooks, 1975; D’Hollander, 1979; Simmons et al., 1979; van Genuchten, 1980; Kovacs, 1981; Russo, 1988; Kosugi, 1996; Assouline et al., 1998). All these curves describe a continuous change of water content from a maximum value, \( \theta_s \), which is called saturated water content, toward a minimum value, \( \theta_r \), which

![Figure 2](image-url)  Typical hydraulic properties of differently textured soils (a) Water retention curves. (b) Unsaturated conductivity curves
is called **residual water content**. The transition from full to partial saturation takes place at some characteristic pressure head (often called **bubbling pressure** or **air entry pressure**). This is related to a characteristic width of the largest pores in the porous medium, and thereby depends on the soil texture and structure. The slope of the WRC at pressures lower than the air entry pressure depends on the width of the pore-size distribution. Thus, all parametric expressions need, in general, at least four adjustable parameters, and differences in their ability to describe observed data are generally small (Buchan and Grewal, 1990).

Empirical studies (van Genuchten and Nielsen, 1985) showed that out of the functions listed above, the expression proposed by van Genuchten (1980) is particularly suited,

\[ S_e = (1 + (\alpha |h|)^n)^{-m} \tag{7} \]

where \( S_e = (\theta - \theta_r) / (\theta_i - \theta_r) \) is a scaled water content, called **effective saturation**, and \( \alpha > 0 \) [L^{-1}], \( n > 1 \) [-] and \( m > 0 \) [-] are empirical curve shape parameters. The inverse of \( \alpha \) is related to the air entry pressure head, \( n \) is related to the width of the pore-size distribution between saturation and air entry pressure, and the product \( nm \) is related to the width of the pore-size distribution between air entry pressure and dryness. In practical use, the van Genuchten equation (7) is mostly applied with the constraint \( m + 1/n = i \) (\( i = 1, 2, \ldots \)). This reduces the flexibility of the function, but has the advantage that it yields closed-form expressions for the hydraulic conductivity functions when inserted into the conductivity prediction models of Mualem (1976), Burdine (1953), or Childs and Collis-George (1950). This makes it particularly easy to use the hydraulic relationships in numerical models (Luckner et al., 1989), and thus the constrained van Genuchten/Mualem model has become de facto a standard in numerical modeling of water transport in unsaturated porous media.

When hydraulic properties are expressed by parametric functions, the process of measuring hydraulic properties is reduced to the process of estimating function parameters. In doing this, it is of crucial importance to be aware that there is no a priori evidence that the selected empirical relationship is indeed suited for a given soil. In particular, any extension beyond the moisture range covered by the actual measurement method is (sometimes reasonable) speculation. Accordingly, the determination of optimal model parameters should be combined with a measure that tests the adequacy of the chosen functional model approach (Finsterle, 1999). Unfortunately, this is rather seldom done in practice, because it requires sufficient measurements over the range of interest to be statistically significant.

The approximate nature of the van Genuchten function and other low-parameterized functions as models for the effective pore-size distribution becomes evident in the asymptotic behavior toward saturation and dryness. From physical reasoning, finite values for both the water capacity function and the slope of the conductivity function are to be expected over the whole moisture range, but for values of \( n < 2 \), the slope of the van Genuchten/Mualem conductivity function becomes infinity close to saturation. Constraining the \( n \) parameter to values greater than 2 is impractical since empirical investigations for all textural soil groups, except for sands, yield optimal fits for \( n < 2 \) (Carsel and Parrish, 1988). Also, toward dryness, the residual water content must be regarded as a pure empirical fitting parameter, and analyses of both, water retention data (Rossi and Nimmo, 1994) and conductivity data, (Tuller and Or, 2001) show that the usual functions are not capable in describing the measured values in the dry range appropriately. On the basis of thermodynamic reasoning, the true equilibrium water contents at pressure head ranges beyond \(-10^5 \) cm are clearly smaller than those given by the usual parameterizations, approaching oven dryness at pressure heads of \( h = -10^5 \) cm. For the practice of simulating water transport in the liquid phase, however, the use of residual water concept appears to yield reasonable results.

The above-cited parametric expressions for WRCs describe measured retention data well if the soils do not exhibit a distinct secondary pore system. To describe naturally structured soils, more flexible expressions have been introduced. These are either constructed by superposition of basic shape functions (Othmer et al., 1991; Durner, 1992; Rossi and Nimmo, 1994) or by piecewise combinations of local shape functions, such as splines or Hermitean basis functions (Erh, 1972; Rossi and Nimmo, 1994; Mohanty et al., 1997; Kastanek and Nielsen, 2001; Prunty and Casey, 2002; Bitterlich et al., 2004). To determine these types of functions accurately, the demand on the precision and range of measurement methods is high (Zumthüll and Durner, 1998).

The consequences of using oversimplified hydraulic functions in practical applications depends on the extent and nature of the deviations between true and parameterized curves, on the range and type of applied boundary conditions, and on the purpose of the measurements. In general, even small changes in hydraulic descriptions can cause significant changes in hydraulic behavior, which is a prerequisite for the identification of hydraulic properties from transient-flow experiments (Vogel et al., 2001; Lambot et al., 2004). But for soil classification purposes, such as the estimate of plant available water, simplified descriptions may perfectly serve its purpose. For use in models of water and solute transport, spatial variability of soil hydraulic functions may be so large that seemingly small errors in the representation of the properties at a point scale appear irrelevant. Deviations of systematic nature, however, can have systematic and significant consequences for the predicted hydraulic conductivity curves.
(Durner, 1994) and thus lead to a different effective system behavior, regardless of the superposed local variability. For very demanding applications, for example, tests of conductivity prediction models, investigations on dynamic flow behavior, or derivations of pedotransfer functions (Schaap, 2005; see Chapter 76, Models for Indirect Estimation of Soil Hydraulic Properties, Volume 2), systematic biases induced by oversimplified functions are intolerable since they lead to misinterpretations of comparisons between observations and models.

**Hydraulic Conductivity Functions**

Similar to the water retention characteristic, the dependence of the hydraulic conductivity on water content, water saturation, or pressure head, is generally expressed by a simple parametric expression. Common to all expressions is the scaling of a shape function, the so-called *relative conductivity function* $K_r(\theta)$, to the saturated conductivity, $K_s$:

$$K(\theta) = K_s \cdot K_r(\theta)$$ (8)

Parametric expressions for the relative conductivity function are less manifold, as compared to the water retention function, because the conductivities are often coupled to the retention model. Simple empirical expressions for the $K(\theta)$ and $K(h)$ relation have been proposed, amongst others, by Gardner (1958), Brooks and Corey (1964), King (1965), and Rijtema (1965). Most frequently used are the exponential model (Gardner, 1958)

$$K(\theta) = K_s \cdot e^{\alpha h}$$ (9)

and the power function (e.g. Averjanov, 1950; Irmay, 1954, Brooks and Corey, 1964; Ahuja and Swarzendruber, 1972),

$$K(\theta) = K_s \cdot S_{\theta}^\beta$$ (10)

where $\alpha$ and $\beta$ are empirical coefficients. The combination of statistical pore-bundle conductivity models, for example, of Burdine (1953) or Mualem (1976), with the retention function of van Genuchten (1980) leads to closed-form expressions, which are sometimes used as empirical “stand-alone” expressions of the conductivity curve. The van Genuchten/Mualem model (van Genuchten, 1980) is most frequently used,

$$K_r(S_c) = S_c^\beta \left[1 - (1 - S_c^{1/m})^m\right]^2$$ (11)

where the parameter $l$ [–] is called *tortuosity factor*, and $m$ is related to the retention curve parameter $n$ in equation (7) by $m = 1 - 1/n$. When applying the van Genuchten/Mualem conductivity model (11), the user should be aware of a model artifact. For small values of the parameter $n$, an unrealistic steep decrease of the predicted conductivity near saturation occurs, because the function considers no finite air entry pressure. Modifications of the van Genuchten/Mualem conductivity function have been proposed that avoid this artifact (Vogel and Cislerova, 1988; Vogel et al., 1999, 2001).

In analogy to the WRC, piecewise combined functions are used to express the conductivity function with enhanced flexibility (e.g. Poulsen et al., 2002; Schwartz and Evett, 2002; Bitterlich et al., 2004). Recently, analytical expressions for retention and conductivity functions have been proposed which are not fully empirical, but depend on hydrodynamic considerations about the topology of the pore space of porous media. Notable are the functions of Assouline et al. (1998), Assouline (2001), Tuller and Or (Tuller et al., 1999; Or and Tuller, 1999; Tuller and Or, 2001), and Zhu and Mohanty (2003).

**HYSTERESIS**

The dependencies of $\theta$ and $K$ on the pressure head $h$ are hysteretic. This implies that the relationships, as shown in Figure 2, are valid only for a certain history of wetting and drying, and for one particular saturation or desaturation path. Most measurement experiments therefore rely on a uniquely defined initial hydraulic state, from which the system is changed continually in one direction. Drainage experiments in the laboratory start usually from a maximum achievable saturation. Rewetting the soil from complete dryness will yield different relationships. Further, drying and wetting cycles from intermediate states between full saturation and oven dryness will again give different relationships. Hysteresis of the water retention characteristic is schematized as shown in Figure 3. Drainage of a full

![Figure 3 Schematization of hysteresis of the water retention characteristic](image-url)
saturated medium yields the primary drying curve, imbibition from oven dryness yields the primary wetting curve, drainage from saturation yields the main drying curve, imbibition from the residual water content yields the main wetting curve, and any changes between imbibition and drainage at intermediate pressure values yield secondary scanning curves.

Hysteresis has been recognized as an important component of soil water redistribution since the early work of Haines (1930). The maximum water saturation of soils in the unsaturated zone under natural conditions is always significantly lower than the porosity, due to air entrapment, with typical saturation values of 80 to 90% of pore space (Klute, 1986; Dane and Hopmans, 2002). Full saturation of the pore space can be obtained only by specific experimental manipulations (slow wetting from the bottom, flushing with carbon dioxide, flushing with de-aired water, and application of vacuum). To distinguish the “saturated water content” from the typical maximum saturation in the field, Hillel (1980) proposed the term “satiation” for the latter.

The hysteresis of $\theta(h)$ leads to hysteresis of $K(h)$, since conductivity is controlled by the water-filled pore space. The hysteresis of the $K(\theta)$ relation of soils is generally regarded as negligible (Mualem, 1986; Kool and Parker, 1987). Ignoring hysteresis in the $K-\theta-h$ relations may be partly based on an assumption that its effects are negligible, partly on the absence of a well-validated hysteretic model that is easy to calibrate (Lenhard et al., 1991), and partly on difficulties of performing measurement experiments that are suitable to identify parameters of hysteresis models. Neglecting hysteresis leads to significant errors in the water redistribution under transient boundary conditions (Dane and Wierenga, 1975; Gillham et al., 1979, Hoa et al., 1977; Viaene et al., 1994; Lehmann et al., 1999; Si and Kachanoski, 2000). Russo et al. (1989) and Mitchell and Mayer (1998) investigated the influence of hysteresis on solute transport and found that the magnitude of the deviations between hysteretic and nonhysteretic simulations was not a simple function of single parameters, but rather depended on the combined values of many or all of the hydraulic parameters. Hysteresis is also seen as one of the major obstacles for comparing different measurement methods for the $K(\theta)$ (Stolte et al., 1994; Basile et al., 2003).

Hysteresis has been incorporated into simulations of unsaturated flow since Rubin (1965). Jaynes (1992) showed that soil water redistribution is retarded by hysteresis, and its magnitude depends on the rate at which the soil-water content varies. This supports the consensus that highly transient soil water conditions enhance hysteresis effects (Topp et al., 1967). Hysteresis is typically assumed to be caused by the inkbottle effect, but other processes such as contact angle hysteresis, shrink-swell effects, or entrapped air play a significant role (Hillel, 1980). Contact angle hysteresis is affected by numerous factors, like surface roughness, chemical heterogeneity, fluid dynamics, particle shape, and gas adsorption, and all these relations depend on the composition of the soil solution (Dury et al., 1998; Henry et al., 2001). Therefore, hysteresis effects on soil-water redistribution may result from a single process, or interactions between different processes, and a treatise on the role of individual factors controlling hysteresis is rather speculative (Kutilek and Nielsen, 1994, p. 73).

Considerable effort has been put into the analysis and description of hysteretic soil hydraulic properties. This has led to numerous models for describing hysteresis in $\theta(h)$. Both, empirical models (Scott et al., 1983; Hogarth et al., 1988) and physically based models (Poulovassilis and Childs, 1971; Parlange, 1976; Mualem, 1974, 1984; Poulovassilis and Kargas, 2000) have been proposed. Kool and Parker (1987) combined the Scott hysteresis model with the retention function of van Genuchten (1980), which leads to an easily usable and consistent set of constitutive relationships (Luckner et al., 1989). These models provide a simple means for determining scanning curves from a limited amount of data, such as the main wetting and drying hysteresis curves. The models of Parlange (1976) and Mualem (1984) need only one branch of the loop to predict all scanning curves. However, when comparing different models of hysteresis using 10 measured scanning curves, Viaene et al. (1994) concluded that the best models were the conceptual models needing two branches for calibration. Jaynes (1992) showed with simulation studies that none of the models were consistently better than the others. Other investigations found the empirical model of Kool and Parker (1987) to be only approximately valid (Schultze et al., 1999; Basile et al., 2003). Incorporating a physically based hysteresis model in numerical codes is demanding, since bookkeeping of the wetting and drying history at any point in the soil must be provided.

**DYNAMIC EFFECTS**

Hydraulic properties of porous media are commonly assumed to be of “static” nature, that is, to depend only on the size distribution and geometrical arrangement of pores in the porous medium and on the wetting/drying history (hysteresis), provided the solid matrix is rigid, fluid properties are constant, and external conditions do not change with time. This implies that the relationship between water content and pressure head during monotonous draining or imbibition processes is not affected by the dynamics of the water flow. As a consequence, $\theta(h)$ can be expressed by a unique characteristic, which can be determined by any available measuring technique (Durner and Lipsius, 2005; see Chapter 75, Determining Soil Hydraulic Properties, Volume 2). However, investigations by various authors...
indicate that hydraulic properties may depend not only on the wetting and drying history, but also on the dynamics of water flow. This implies that hydraulic functions, which are measured under static equilibrium conditions, are not applicable to simulate water flow under transient conditions.

Davidson et al. (1966) found in imbibition and drainage studies that equilibrium water contents were affected by the dynamics of the wetting history. More water was removed from samples by one single large pressure decrease than by a sequence of small decreases. In contrast, more water was sorbed by the soils when a series of small pressure steps was applied in the imbibition process. Topp et al. (1967) compared water retention characteristics of vertical sand columns under static equilibrium conditions, under steady state flow, and under transient-flow conditions. They found that for transient experiments, at a given potential, water contents at drainage were significantly higher than water contents determined under static equilibrium or steady state conditions. Smiles et al. (1971) performed experiments where a series of imbibition/drainage cycles were applied to horizontal sand columns by imposing stepwise changes in the water pressure at one column end. They found that the retention characteristic depended on the size of the imposed pressure. Vachaud et al. (1972) confirmed these results for vertical soil columns. Stauffer (1977) investigated drainage in vertical columns packed with sand. Under dynamic flow conditions, he measured higher water contents at a given pressure as compared to the static \( \theta(h) \) relationship. Similar to Smiles et al. (1971), gas phase pressure was additionally measured and no deviation from atmospheric air pressure was observed. Plagge (1991) and Plagge et al. (1999) confirmed these findings with evaporation experiments on silty soils. Tensiometric pressures and water contents were measured with microtensiometers and TDR (time domain reflector) probes, which were installed at multiple depth levels in soil columns. Dependent on the distance from the boundary of the soil column, the locally measured retention curves differed considerably. The differences were systematic and reproducible, and could not be attributed to faults in packing technique or to measurement errors. Lennartz (1992) investigated dynamic nonequilibrium systematically by performing evaporation experiments on repacked soil samples of four different substrates. He found significant differences between static and dynamic \( \theta(h) \) characteristics, but could not identify a unique and simple tendency.

Wildenschild et al. (2001) investigated the flow-rate dependence of unsaturated hydraulic properties for disturbed soils in short laboratory columns by performing one-step and multistep outflow experiments. Retention characteristics were obtained from measured tensiometric pressures in the soil column and the average water content. For a sandy soil, they found pressure head differences of 10 to 15 cm (for a given saturation) and water content differences of up to 7% (for a given potential) between the slowest and the fastest outflow experiments. At a given pressure head, more water was retained with greater applied pressure steps. Conversely, Constantz (1993) reported higher water contents in slow multistep drainage experiments as compared to fast one-step drainage experiments. He attributes this to variations in pore water salt concentrations that induce differences in the pore water surface tension. Simunek et al. (2001) performed upward infiltration experiments and observed ongoing infiltration despite apparent equilibrium in the pressure heads. Also, during a 5-day redistribution phase where no flux across the boundaries occurred, they found an increase in tension in the whole soil sample by \( \Delta h = 50 \) cm. All the above-mentioned experiments indicate that the dynamic effect depends on the size of the pressure changes, being larger for big changes. Further, there is some indication that dynamic effects are larger for soils with a wide pore-size distribution. Without relating the observed phenomena to specific processes, we may categorize them in a general way as “dynamic nonequilibrium” (Schultze et al., 1999). According to Klute (1986, p. 660), the reasons for and implication of this observation continue to be a subject of investigation.

On the basis of a theoretical framework for multiphase flow in porous media, Hassanizadeh and Gray (1993) and Hassanizadeh et al. (2002) postulated the existence of a dynamic component in unsaturated water flow. Their analysis yields an approximate capillary pressure equation with a dynamic term, which depends linearly on the rate of change in water saturation

\[
p_a - p_w = p_c - \frac{1}{L} \frac{\partial S}{\partial t}
\]  

(12)

where \( p_w \) [\( ML^{-1} T^{-2} \)] is the macroscopic pressure in the water phase, \( p_a \) [\( ML^{-1} T^{-2} \)] is the pressure in the air phase, \( p_c \) [\( ML^{-1} T^{-2} \)] is the capillary pressure, \( L \) [\( TL^{-1} M^{-1} \)] is a nonnegative material coefficient, \( S = \theta/\theta_s \) [-] expresses the soil water saturation, and \( t \) [\( T \)] indicates the time. Equation 12 states that the pressure difference between air and water pressure is larger than the (equilibrium) capillary pressure under drainage conditions, and smaller when imbibition occurs.

Macroscopic simulation of water transport including dynamic effects is still in its infancy. Stauffer (1977) simulated the dynamic process with Richard’s equation by using a dynamic retention characteristic, where the difference to the static characteristic depended on the rate of change of the local water content with time. Schultze et al. (1999) showed that a two-phase model explains much of the observed dynamic flow behavior in a soil column. Simunek et al. (2001) simulated water transport with a dual-continuum model (Gerke and van Genuchten, 1993), and found good agreement with measurements, when
interpreting measured tensiometric measurements as being representative for the interaggregate pore space.

Summarizing, evaluations of transient-flow experiments, thermodynamic reasoning, pore-network modeling (Held and Celia, 2001), and continuum percolation theory (Hunt, 2004) all lead to the conclusion that hydraulic relations are not only dependent on the wetting/drying history and the actual system state, but also on the rate of change of the system state. This is a problem, because hydraulic functions are primarily used in model applications under transient conditions, whereas most measurement experiments are based on equilibrium conditions (Durner and Lipsius, 2005; see Chapter 75, Determining Soil Hydraulic Properties, Volume 2). However, it is currently not clear as to what degree dynamic effects will affect water flow on time and spatial scales that are relevant for most applications. Despite its early notion, the problem of dynamic nonequilibrium during water flow in soils has not been adequately treated yet. Reasons for this are the high requirements with respect to the measurement, and even the historic inability to tackle the theoretical problem. Without the availability of fast computers and fast measurement techniques, quantitative comparisons of observed transient-flow process variables with simulation results obtained by models of increasing complexity, such as Richards’ equation, the coupled two-phase flow model, and the pore-network models, were not possible. Since dynamic nonequilibrium in water flow is closely related to hysteresis, a satisfying solution of the hysteresis problem also depends on the progress of this issue. In part, the distinction between hysteresis and dynamic nonequilibrium in water flow can be seen as a matter of the time scale, since a true equilibrium distribution of the water phase (which actually may require enormous time) will cause much of the apparently “static” hysteresis to disappear.

**VARIABILITY**

Soils are heterogeneous from the pore to the geologic scale. We have already noted that homogeneity in soils does not exist unless we refer to the concept of the REV. If the scale of observation is less than the REV, the soil is heterogeneous (Figure 1). Spatial variability of soils critically affects the effective capacity and transmission properties of water, solutes and gases, and controls by the formation of micro-niches and gradients of intensive soil properties during flow processes and the ecological functioning of the vadose zone. Reviews by Warrick and Nielsen (1980), Peck (1983) and Jury (1985) have shown that in soils, water flow and transport properties are the most variable. Jury *et al.* (1991), Kutilek and Nielsen (1994), Warrick (1998), Mulla and McBratney (2002), and van Es (2002) have reviewed techniques to handle spatial variability and soil heterogeneity.

Solute transport in soils is critically dependent on the variability of soil water fluxes. Increased travel time variance results in increased probability of short travel times, that is, preferential flow and transport. Soil water flux variability depends on the soil moisture status (Roth, 1995), and can be enhanced as compared to the variability of the soil moisture or hydraulic conductivity parameters. Owing to the nonlinearity of unsaturated water transport, spatial variability leads to scale dependence of hydraulic properties. Investigation of these questions is amongst the most critical topics in contemporary soil physical and hydrologic research (Sposito, 1998; Roth *et al.*, 1999). Scale issues are treated in this Encyclopedia in Chapter 66, Soil Water Flow at Different Spatial Scales, Volume 2 (Hopmans and Schoups, 2005, this issue).

Basically, there are two ways to tackle properties of heterogeneous soils. One way lies in the investigation of frequency distributions of local properties and in the characterization of their spatial structure. The other way is to search for effective transport processes and parameters, allowing for the application of overall initial and boundary conditions for a quasi-homogeneous system. Unfortunately, there is no general upscaling rule for hydraulic properties of unsaturated soils, and even the question about the existence of appropriate process descriptions with effective hydraulic properties at larger scales is not resolved yet. This is of fundamental importance when comparing the results of laboratory and field experiments.

**Spatial and Temporal Variability**

Since measurement scales are often smaller than the REV scale, soil hydraulic properties are often extremely variable in space and do not conform to conventional statistical assumptions. The structure of variability may be random, correlated, periodic, or in any combination, and may also be scale dependent (van Es, 2002). Spatial variability of hydraulic properties has been documented in many articles (e.g. Nielsen *et al.*, 1973; Simmons *et al.*, 1979, Russo and Bouton, 1992; Istok *et al.*, 1994; Jarvis and Messing, 1995; Mohanty *et al.*, 1994; Shouse *et al.*, 1995; Russo *et al.*, 1997; Shouse and Mohanty, 1998). Water contents in the field are mostly found to be normally distributed with medium variation (coefficient of variation, 0.15 < cv < 0.5; Jury *et al.*, 1991; Warrick, 1998), with increasing variability when the soils dry out (Shouse *et al.*, 1995). Field investigations of hydraulic conductivities yield in most cases lognormal distributions, with variances of log(K) in the order of 1 (Dagan, 1984; Butters *et al.*, 1989; Heuvelman and McInnes, 1997; Holt *et al.*, 2002).

Spatial variability is overlaid by temporal variability. The soil pore system is affected by seasonal changes, by diurnal cycles of temperature and corresponding energy gradients, by plant and root growth, by irrigation cycles, and by freezing and thawing. Water transport in soil is
generally driven by weather-related processes or irrigation, which have high short-term dynamics. Soil properties that affect hydraulic behavior, aeration, erosion and runoff, and aggregation are strongly affected by processes such as soil wetting and drying, freezing and thawing, and weather-related behavior of soil organisms (van Es, 2002).

Cultural influences cause spatial and temporal extrinsic variability. Drainage, tillage, vehicle traffic, overburden pressure, plant cover, and soil amendments may dramatically alter mean behavior and variability patterns. For example, tillage tends to homogenize soils spatially, but may cause greater temporal variability from loosening and subsequent settling and recompaction (Mapa et al., 1986; Starr, 1990, van Es et al., 1999).

**Geostatistical Characterization of Random Variability**

Variations of soil properties are not completely disordered over the field and contain systematic and random components. To describe spatial correlation structures, tools of geostatistics, based on the theory of regionalized variables, have been developed (Matheron, 1963; Journel and Huijbregts, 1978) and successfully applied in soil science (e.g. Warrick and Nielsen, 1980). On the basis of proper statistical information, stochastic random fields can be created and the effective behavior of these “virtual realities” can be investigated.

In stochastic models, it is often assumed that the correlation lengths of different unsaturated parameters are the same (Holt et al., 2002). Horizontal correlation scales for hydraulic properties were reported to be in the order of 6 m to 20–30 m (Ciollaro and Romano, 1995; Shouse et al., 1995), whereas vertical correlation scales are in the order of 0.1 to 0.3 m (Robin et al., 1991; Wierenga et al., 1991). Heuvelman and McInnes (1997) investigated spatial variability of water fluxes on a scale of centimeters and found water flux densities to be normally distributed in a sandy topsoil, becoming lognormally distributed in a loamy subsoil. Jury (1985) and Shouse et al. (1995) noted that the range of spatial influence may be related to the scale of the sampled area, and that the commonly observed high nugget variance of soil hydraulic properties may be more closely related to measurement error than to high frequency, small-scale variations.

A particular problem of vadose zone spatial variability results from the fact that the typical travel distance is small compared with the distance required for a solute plume to reach asymptotic, Fickian behavior (Jury and Flühler, 1992). This means that hydraulic properties on the scale of interest are not randomly distributed. Hence, application of geostatistical concepts to derive asymptotic large-scale behavior is of limited value. This is further aggravated by the fact that proper three-dimensional geostatistical characterization of hydraulic properties needs such a high number of measurements that it cannot be obtained in practice. It appears that a reliable prediction of solute transport in natural fields requires the quasi-deterministic characterization of the largest spatial system structures, which are often on the same scale as the scale of interest (Vogel and Roth, 2003). This cannot be obtained by point measurements. Therefore, the further development of hydrogeophysical methods, possibly coupled with remote sensing technologies, and the linking of these data with functional hydraulic properties, for example, by stochastic data fusion (Yeh and Simunek, 2001), will play a crucial role in future developments.

**Influence of Measurement Errors on Spatial Data Analysis**

The analysis of spatial variability depends on measurements that are assumed to have little or no experimental or calibration error. Unfortunately, this is rarely the case. Many estimated hydraulic properties are likely to contain systematic error, or bias, in particular, if they are not measured directly. Although most studies carefully document instrumental procedures, the magnitude of errors in hydraulic property estimates and their impact on spatial statistics determined from field data are rarely evaluated. Ciollaro and Romano (1995), for example, used inverse modeling of evaporation experiments to derive the hydraulic properties of a 135-m transect, and found a discrepancy between optimized saturated conductivity and directly measured saturated conductivity by a factor of 10. Shouse et al. (1995) conclude that more information is needed on experimental error, its causes, and remedies.

If nonlinear inversion models, whether analytical or numerical, are used to infer property values from observed system responses and boundary conditions, random error in the observations (observation error) can lead to spatially correlated, systematic error, or bias, in the derived property value. Spatial bias may also result when the inversion model (e.g. governing equations, boundary conditions, initial conditions, constitutive models, etc.) is inadequate (inversion model error) (Kemphorne and Allmaras, 1986). Holt et al. (2002) used a Monte Carlo approach to explore the potential impact of observation and inversion model errors on the spatial statistics of field-estimated unsaturated hydraulic properties. For this analysis they simulated tension infiltrometer measurements in a series of idealized realities, each consisting of spatially correlated random property fields. They showed that estimated hydraulic properties are strongly biased even by small, simple observations, and inversion model errors. Error in spatial statistics was more than an order-of-magnitude, and artificial cross-correlation between measured properties occurred. Unfortunately, there are no unique indicators of bias, as property values may appear reasonable and spatial statistics may look realistic.
Errors in the spatial statistics of hydraulic properties cause critical stochastic model assumptions to be violated, limiting the usable parameter space for model predictions. Even where critical assumptions are valid, stochastic model predictions show significant error, and the magnitude and pattern of error changes with the true property means, the flow conditions, and the type of measurement error. Experimental technologies need to be developed that reduce experimental error and increase precision.

Scaling of Hydraulic Functions

A particular challenge for handling spatially variable unsaturated hydraulic properties lies in the fact that not just distributions of single parameters, like \( K_s \), must be characterized, but the variability of the whole constitutive relationships, \( K(\theta) \) and \( \theta(h) \). A way to handle this is using the concept of similar media scaling. The objective of scaling is to coalesce a set of hydraulic relationships into a single reference curve using scaling factors that describe the set as a whole. This method was introduced by Miller and Miller (1956), and its concepts and limitations were reiterated and assessed later by Miller (1980), Tillotson and Nielsen (1984), and Sposito and Jury (1985). Simmons et al. (1979) extended similar media scaling theory to characterize the spatial variability of field water retention measurements. Jury (1985) indicated that an obvious limitation of scaling theories is that the errors involved in measuring the properties used to calculate the scale factors are carried along as a part of the scale factor sample variance. Reviews and application examples of scaling soil hydraulic functions are given by Hillel and Elrick (1990), Kutilek and Nielsen (1994), and Roth (1995).

Mean and Effective Soil Hydraulic Properties

Since the scale of interest is in general larger than a local REV, upscaling of hydraulic properties is required to get a proper description on the scale of interest. When nonlinear properties have to be averaged, the mean property is in the general case not identical to the “effective” property. An effective property represents a homogeneous medium with a macroscopic behavior that is quasi-identical to the heterogeneous system. For the unsaturated case, the differences between mean and effective properties are dependent on the spatial variability of the local properties, but furthermore also on the boundary conditions, as will be illustrated for the case of steady state water flow into a moderately heterogeneous soil.

Figure 4 shows a simulated domain of a weakly heterogeneous soil, composed of irregular lenses of sand and silt forming a 2-m deep soil body (Flühler and Roth, 2004). The structures have a wide extension in the horizontal and a small extension in the vertical scale. Water movement takes place from top to bottom. The water flux \( q \) through the upper boundary is continuous and constant. At the lower boundary there is contact to the groundwater table, that is, \( h = 0 \). The depicted structure is assumed to be periodic, that is, water flowing out through a vertical boundary is assumed to flow in at the opposite boundary. The hydraulic properties of the sand are given by the van Genuchten/Mualem coefficients \( \alpha = 0.02 \text{ cm}^{-1}, \ n = 4.0, \ \theta_s = 0.3, \ \theta_r = 0, \ K_s = 10^{-4} \text{ m s}^{-1} \), and those of the silt by \( \alpha = 0.005 \text{ cm}^{-1}, \ n = 1.33, \ \theta_s = 0.4, \ \theta_r = 0, \ K_s = 10^{-5} \text{ m s}^{-1} \). The situation shown in Figure 4 is the stationary flow field corresponding to a weak rain, with a flux density at the upper boundary of \( 1.8 \times 10^{-7} \text{ m s}^{-1} = 1.56 \text{ cm d}^{-1} \). At this low infiltration rate, the silt in the upper 50-cm region, far from the groundwater table, has a much higher conductivity than the sand, and water flows convergent through the silt lenses. In this stationary situation,
the water saturation reflects the substrate changes very sharply, whereas the lateral distribution of the pressure head is much smoother, in particular, close to the groundwater table. As a fundamental consequence, single point measurements of matric potential are more representative for the average pressure than point measurements of water contents for the average water content. At about 1.2-m depth, the two materials have essentially the same conductivity, leading to a uniform flow of water. Close to the groundwater table, the silt lenses are obstacles to water flow, and water flows preferentially in the sand matrix. Figure 5 shows the water saturation in the same profile, corresponding to a stationary flow with an increased infiltration rate of \( q = 5.4 \times 10^{-5} \text{ m s}^{-1} \). Now, the silt lenses are obstacles to water flow in the whole profile.

Figure 6 shows the corresponding depth profiles of the mean hydraulic properties \( (h) \), \( (\theta) \), and \( \log_{10}(K) \), obtained by horizontally averaging the local values of the heterogeneous soil. The dashed lines indicate the situation for the low water flux density, the drawn lines for the high flux density. It is apparent that the average water content, \( (\theta) \), varies much more than the average water potential, \( (h) \), which reflects a typical gravitational flow situation for the high flux rate (Figure 6, left, drawn line) and a situation close to hydrostatic equilibrium conditions for the low flux rate (Figure 6, left, dashed line). Figure 6 (right) depicts the mean (thick lines) and effective hydraulic conductivity (thick lines) for the corresponding one-dimensional water flux. We observe that even for the case of quasi-gravitational unit-gradient flow, the mean hydraulic conductivity, \( \log_{10}(K) \), in the upper part of the profile is higher than the infiltration rate, since the local water flux is strongly reflected from the straight vertical movement.

The effective hydraulic conductivity for the flow situation is obtained by dividing the macroscopic water flux density by the gradient of the mean water potential. For the high flux density, the resulting effective hydraulic conductivity is in the range of the infiltration rate, which is about 5 times smaller than the mean hydraulic conductivity. For the low infiltration rate, the difference between mean and effective conductivity is in the order of two magnitudes.

**CONCLUDING REMARKS**

Enormous advances have been made during the last few decades regarding our understanding and ability to model flow and transport processes in the vadose zone (Simunek, 2005; see Chapter 78, Models of Water Flow and Solute Transport in the Unsaturated Zone, Volume 1). The current scientific perspective is that the Richards’ equation provides the most appropriate tool on which any description of water transport in the vadose zone can be based. Other simplified descriptions, such as bucket models, are not considered as a viable alternative (Vanclooster et al., 2004). More comprehensive descriptions, such as the use of two-phase flow models (for water and air) appear superfluous for standard uses. The validity of the Richards’ equation for larger scales is a matter of ongoing debate. From a theoretical standpoint and based on stochastic hydrology a Richards’-type equation for water flow appears to retain its validity for larger scales, yet with different constitutive relationships, which are dependent both on the stochastic distribution of local soil hydraulic properties and also on the boundary conditions (Harter and Hopmans, 2004).

For the characterization of soil hydraulic properties the most critical problem lies in the derivation of effective properties for large-scale applications, such as for hill slopes,
catchments, or entire regions. Nevertheless, the description of hydraulic functions on the measurement scale, needs improvement as well, both in the moisture range near saturation (for solute transport and infiltration), and also toward dryness. The latter is of particular importance for simulating evaporation and transpiration under conditions where the conductivity of the soil becomes a limiting factor. Furthermore, a valid description of hysteresis in the intermediate moisture range is required for transient-flow conditions. The development of accurate, fast and reliable measurement methods, a further assessment of “dynamic effects”, and the description of the hydraulic properties for shrinking and swelling soils (Smiles and Raats, 2005; see Chapter 67, Hydrology of Swelling Clay Soils, Volume 2) remain major challenges to soil scientists and hydrologists. Some important issues with respect to the future use of Richards’ equation are (i) a suitable integration of boundary conditions on large scales (integration in landscape models), (ii) the adequate representation of soil hydraulic functions on large scales, (iii) hysteresis, (iv) spatial and temporal variability, (v) structured soils and macropores, and (vi) unstable wetting fronts as well as factors dealing with soil hydrophobicity.

REFERENCES


