Chapter 4
Soil Properties

CHAPTER 4: SOIL PROPERTIES

Infiltration is the movement of water into the soil. This is possible, because soil is not solid matter; instead it is a porous medium comprising a matrix of solid granular particles and voids that may be filled with air or water (Figure 21). Flow in a porous medium may be unsaturated when some of the voids are occupied by air, or saturated when all the voids are occupied by water. Considering the cross section of a porous medium illustrated in Figure 21, the *porosity* is defined as

$$n = \frac{\text{volume of voids}}{\text{total volume}} \tag{6}$$

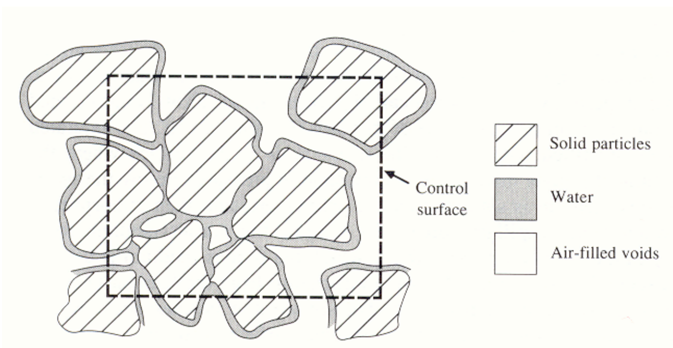


Figure 21. Cross section through an unsaturated porous medium (from Chow et al., 1988).

The range of n for soils is approximately 0.25 to 0.75 depending upon the soil texture. A part of the voids is occupied by water, and the remainder by air. The volume occupied by water being measured by the *volumetric soil moisture content* is defined as

$$\theta = \frac{\text{volume of water}}{\text{total volume}} \tag{7}$$

Hence $0 \leq \theta \leq n$; the soil moisture content is equal to the porosity when the soil is saturated. Soil moisture content is also sometimes characterized by the degree of saturation, defined as

$$S_d = \theta/n \tag{8}$$

The degree of saturation varies between 0 and 1.

Referring to Figure 21, the soil particle density, ρ_m, is the weighted average density of the mineral grains making up a soil

$$\rho_m = M_m / V_m \tag{9}$$

where M_m is the mass and V_m the volume of the mineral grains. The value of ρ_m is rarely measured, but is estimated based on the mineral composition of the soil. A value of 2650 kg/m^3, which is the density of the mineral quartz, is often assumed. The bulk density, ρ_b, is the dry density of the soil

$$\rho_b = \frac{M_m}{V_s} = \frac{M_m}{V_a + V_w + V_m} \tag{10}$$

where V_s is the total volume of the soil sample which is the sum of the volume of the air, V_a, liquid water, V_w, and mineral components, V_m, of the soil respectively. In practice, bulk density is defined as the mass of a volume of soil that has been dried for an extended period (16 hr or longer) at 105 °C, divided by the original volume. The porosity (6) is given by

$$n = \frac{V_a + V_w}{V_s} = \frac{V_s - V_m}{V_s} = 1 - \frac{M_m / V_s}{M_m / V_m} = 1 - \frac{\rho_b}{\rho_m} \tag{11}$$

and n is usually determined by measuring ρ_b and assuming an appropriate value for ρ_m. Laboratory determination of volumetric moisture content θ is by first weighing a soil sample of known volume, oven drying it at 105 °C, reweighing it and calculating

$$\theta = \frac{M_{swet} - M_{sdry}}{\rho_w V_s} \tag{12}$$

Here M_{swet} and M_{sdry} are the masses before and after drying, respectively, and ρ_w is the density of water (1000 kg/m^3). This method for determining soil moisture is referred to as the gravimetric method. In the field moisture content can be measured in a number of other ways. Electrical resistance blocks use the inverse relationship between water content and the electrical resistance of a volume of porous material (e.g. gypsum, nylon or fiberglass) in equilibrium with the soil. Neutron probe moisture meters are combined sources and detectors of neutrons that are inserted into access tubes to measure the scattering of neutrons by hydrogen atoms, which is a function of moisture content. Gamma-ray scanners measure the absorption of gamma rays by water molecules in soil between a source and a detector. Capacitance and time-domain reflectometry (TDR) techniques measure the dielectric property of a volume of soil, which increases strongly with water content. Nuclear magnetic resonance techniques measure the response of hydrogen nuclei to magnetic fields. Remote sensing and specifically, microwave remote sensing can provide information about surface soil water content over large areas. Both active and passive microwave systems exist, with active systems (radar) having higher resolution. Because of the importance of soil moisture in hydrologic response, as well as land surface inputs to the atmosphere, the relationship of soil moisture to remote sensing measurements is an area of active research. The assimilation of remote sensing measurements of soil moisture into hydrologic and atmospheric forecasting models is one exciting aspect of this research that holds the potential for improving hydrologic and atmospheric model forecasts. For details on these methods for soil moisture measurement the reader is referred to soil physics texts, or the research literature (e.g., Hillel, 1980)

The flow of water through soil is controlled by the size and shape of pores, which is in turn controlled by the size and packing of soil particles. Most soils are a mixture of grain sizes, and the grain size distribution is often portrayed as a cumulative-frequency plot of grain diameter (logarithmic scale) versus the weight fraction of grains with smaller diameter (Figure 22). The steeper the slopes of such plots, the more uniform the soil grain-size distribution.

For many purposes the particle size distribution is characterized by the *soil texture,* which is determined by the proportions by weight of clay, silt and sand. Clay is defined as particles with diameter less than 0.002 mm. Silt has a particle diameter range from 0.002 mm to 0.05 mm and sand has particle diameter range from 0.05 to 2 mm. Figure 23 gives the method developed by the U.S. Department of Agriculture for defining textures based on proportions of sand, silt and clay. Larger particles with grain sizes greater than 2 mm are excluded from this proportioning in the determination of texture. Grain size distributions are obtained by sieve analysis for particles larger than 0.05 mm and by sedimentation for smaller grain sizes. Sieve analysis is a procedure where the soil is passed through a stack of successively finer sieves and the mass of soil retained on each sieve is recorded. Because soil grains are irregular shapes, the practical definition of diameter then amounts to whether or not the soil grain passes through a sieve opening of specified size. Sedimentation is a procedure whereby the settling rate in water of soil particles is measured. For details see a soil physics reference (e.g. Hillel, 1980).

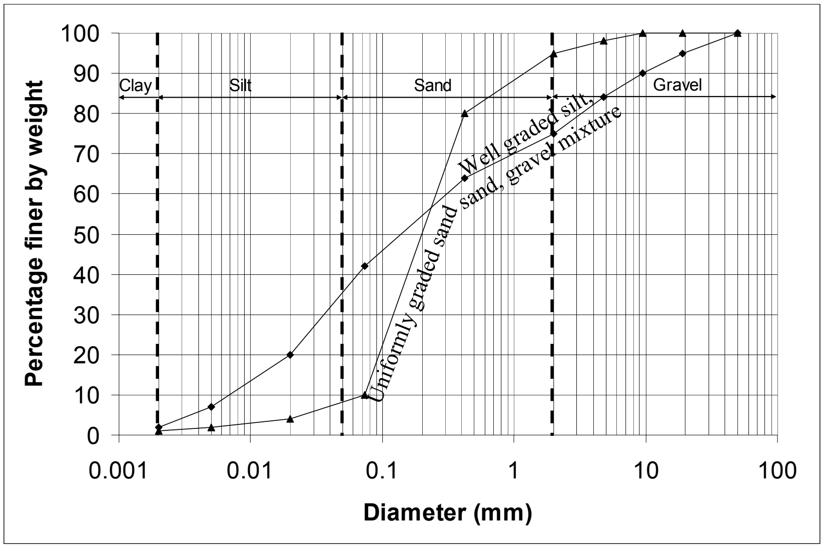


Figure 22. Illustrative grain-size distribution curves. The boundaries between size classes designated as clay, silt, sand and gravel are shown as vertical lines.

Following are the grain sizes used for the determination of texture for the soils illustrated in Figure 22.

Diameter (mm)	A. % Finer	B. % Finer	A. % Finer < 2mm only	B. % Finer < 2mm only
50	100	100		
19	95	100		
9.5	90	100		
4.76	84	98		
2	75	95	100	100
0.42	64	80	85.3	84.2
0.074	42	10	56	10.5
0.02	20	4	26.7	4.2
0.005	7	2	9.3	2.1
0.002	2	1	2.7	1.1

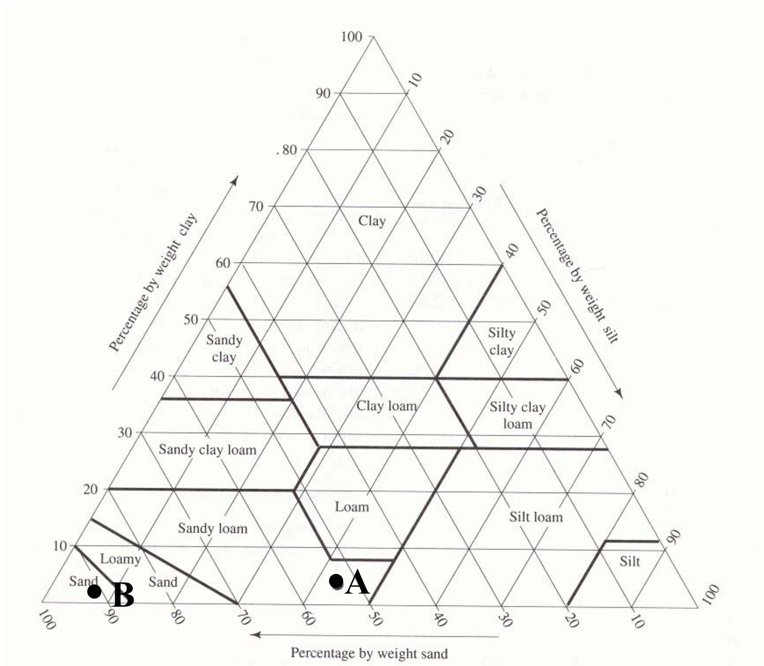


Figure 23. Soil texture triangle, showing the textural terms applied to soils with various fractions of sand, silt and clay (Dingman, Physical Hydrology, 2/E, © 2002. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey)

Soil A is a well-graded mixture comprising gravel, silt and sand in roughly equal proportions. The majority of grains in soil B are all from the same sand size class. It is therefore described as uniformly graded sand. The percentages of Sand, Silt and Clay, for these soils determine the texture as indicated by the dots in Figure 23

	A	B
% Sand	52.8	91.4
% Silt	44.5	7.6
% Clay	2.7	1.1
Texture (Figure 23)	Sandy Loam	Sand

The soil properties, porosity, moisture content, bulk density are defined in terms of averages over a volume referred to as the representative elementary volume (Bear, 1979). It is not meaningful, for example, to talk about these quantities as a very small scale where we are looking at individual soil grains or particles. These properties (and others such as hydraulic conductivity and specific discharge to be defined below) represent averages over the representative elementary volume and are referred to as continuum properties of the porous medium. The macroscopic continuum representation of flow through a porous medium relies on this concept to overlook the complexity of the microscopic flow paths through individual pores in a porous medium (see Figure 24). Typically the representative elementary volume is about 1 to 20 cm^3. Where heterogeneity exists in a porous medium at all scales, the definition of macroscopic continuum properties can be problematic.

At the macroscopic scale, flow through a porous medium is described by *Darcy's equation*, or *Darcy's law*. The experimental setup used to define Darcy's equation is illustrated in Figure 25. A circular cylinder of cross section A is filled with porous media (sand), stoppered at each end, and outfitted with inflow and outflow tubes and a pair of piezometers. (A piezometer is a tube inserted to measure fluid pressure based on the height of rise of fluid in the tube.) Water is introduced into the cylinder and allowed to flow through it until such time as all the pores are filled with water and the inflow rate Q is equal to the outflow rate. Darcy found that the flow rate Q is proportional to cross sectional area A, the piezometer height difference Δh, and inversely proportional to the distance between piezometers, Δl. This allows an equation expressing this proportionality to be written

$$Q = -KA\frac{\Delta h}{\Delta l} \tag{13}$$

where the negative sign is introduced because we define $\Delta h = h_2 - h_1$ to be in the direction of flow. K, the proportionality constant is called the hydraulic conductivity. Hydraulic conductivity is related to the size and tortuousity of the pores, as well as the fluid properties of viscosity and density. Because the porous medium in this experiment is saturated, K here is referred to as the saturated hydraulic conductivity. The *specific discharge*, q, representing the per unit area flow through the cylinder is defined as

$$q = \frac{Q}{A} \tag{14}$$

Q has dimensions $[L^3/T]$ and those of A are $[L^2]$ so q has the dimensions of velocity $[L/T]$. Specific discharge is sometimes known as the *Darcy velocity*, or *Darcy flux*. The specific discharge is a macroscopic concept that is easily measured. It must be clearly differentiated from the microscopic velocities associated with the actual paths of water as they wind their way through the pores (Figure 24).

Figure 24. Macroscopic and microscopic concepts of porous medium flow (Freeze/Cherry, Groundwater, © 1979. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey).

The proportion of the area A that is available to flow is nA. Accordingly the average velocity of the flow through the column is (Bear, 1979)

$$V = Q/nA = q/n \tag{15}$$

Using specific discharge, Darcy's equation may be stated in differential form

$$q = -K\frac{dh}{dl} \tag{16}$$

In equation (16) h is the *hydraulic head* and dh/dl is the *hydraulic gradient*. Since both h and l have units of length [L], a dimensional analysis of equation (16) shows that K has the dimensions of velocity [L/T]. Hydraulic conductivity is an empirical porous medium and fluid property. We discuss later how it can be related to pore sizes and the viscosity of water. In Figure 25, the piezometers measure hydraulic head. The pressure in the water at the bottom of a piezometer (location 1 or 2 in Figure 25) is given by

$$p = (h-z)\gamma = (h-z)\rho_w g \tag{17}$$

where $\gamma = \rho_w g$ is the specific weight of water, the product of the density and gravitational acceleration, g (9.81 m/s^2). The hydraulic head h is comprised of elevation z above any convenient datum and a *pressure head* term $\psi = p/\gamma$.

$$h = z + p/\gamma = z + \psi \tag{18}$$

Pressure head represents the pressure energy per unit weight of water. The elevation z above the datum is also termed *elevation head*, and represents the potential energy, relative to the gravitational field, per unit weight of water. It is important to note that equations (13) and (16) state that flow takes place from a higher hydraulic head to a lower hydraulic head and not necessarily from a higher pressure to a lower pressure. The pressure at location 2 can still be higher than the pressure at location 1, with flow from 1 to 2. The hydraulic head difference Δh in (13) represents a hydraulic energy loss due to friction in the flow through the narrow tortuous paths (Figure 24) from 1 to 2. Actually, in Darcy's equation, the kinetic energy of the water has been neglected, as, in general, changes in hydraulic head due to pressure and elevation along the flow path are much larger than changes in the kinetic energy.

See Online Resource

View the Darcy Experiment Example

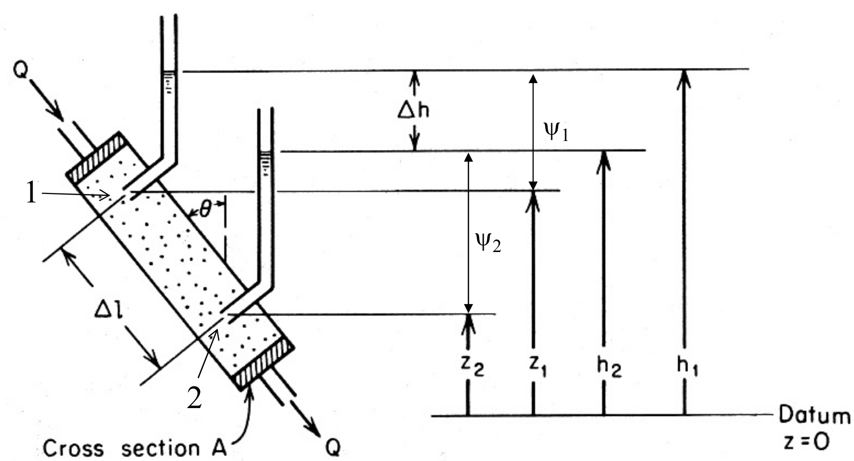


Figure 25. Experimental apparatus for the illustration of Darcy's equation (Freeze/Cherry, Groundwater, © 1979. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey).

Darcy's equation, as presented here, is for one dimensional flow. Flow in porous media can be generalized to three dimensions in which case the hydraulic gradient, becomes a hydraulic gradient vector, and the hydraulic conductivity becomes a hydraulic conductivity tensor matrix in the most general case of an anisotropic medium. Refer to advanced texts (e.g. Bear, 1979) for a discussion of this.

One conceptual model for flow through a porous media is to represent the media as a collection of tiny conduits with laminar flow in each (Figure 26). The average velocity in each conduit is given by the Hagen-Poiseville equation (Bras, 1990, p291)

$$v_i = -\frac{\gamma d_i^2}{32\mu}\frac{dh}{dl} \tag{19}$$

where d_i is the conduit diameter and μ the dynamic viscosity (which for water at 20 °C is 1.05 x 10^{-3} N s m^{-2}).

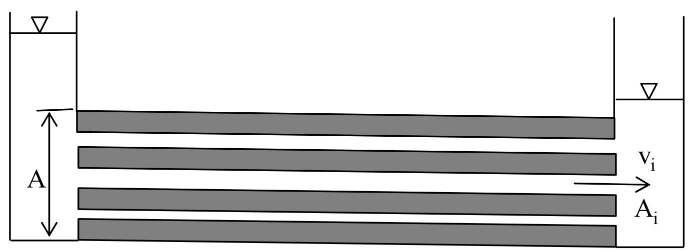


Figure 26. Parallel conduit conceptual model for porous media flow.

The flow in each conduit may be expressed as v_iA_i where A_i is the cross sectional area of each conduit. Summing these and expressing flow per unit area, the specific discharge is

$$q = \frac{\sum v_i A_i}{A} = -\frac{1}{A}\sum A_i \frac{\gamma d_i^2}{32\mu}\frac{dh}{dl} \tag{20}$$

By comparison with (16) the hydraulic conductivity is

$$K = \frac{1}{A}\sum A_i \frac{\gamma d_i^2}{32\mu} = \frac{\gamma}{\mu}\underbrace{\left(\frac{1}{A}\sum A_i d_i^2 / 32\right)}_{\text{medium property k}} = \frac{\gamma}{\mu}k \tag{21}$$

Here hydraulic conductivity K has been expressed in terms of fluid properties (γ/μ) and medium properties grouped together into the quantity k, which is called the medium's *intrinsic permeability*. Intrinsic permeability has units of area [L^2] and (21) suggests this should be related to the average pore area. Equation (21) represents a conceptual model useful to understand the intrinsic permeability of porous media. Real soils are more complex than straight tiny conduits. Nevertheless, experiments with fluids with different viscosity and density, and a porous media comprising glass beads of different diameter have supported the extension of Darcy's equation to

$$q = -\frac{Cd^2\gamma}{\mu}\frac{dh}{dl} \qquad (22)$$

where d is effective grain diameter and $k=Cd^2$. C is a constant of proportionality that accounts for the geometry and packing in the porous media. Effective grain diameter d may be taken as mean grain diameter, or d_{10}, the diameter such that 10% by weight of grains are smaller than that diameter. Differences in these definitions of d are absorbed in the constant C. The intrinsic permeability quantifies the permeability of a porous medium to flow of any fluid (e.g. air, oil, water) and is more general than the concept of hydraulic conductivity. The viscosity of water is temperature and salinity dependent, and this can be accounted for using (21), although this is rarely done in practice in infiltration and runoff generation calculations.

The conceptual model above relied on laminar flow and the linear relationship in Darcy's equation is a consequence of the flow through porous media being laminar. Limits to this linearity have been suggested. For fine grained materials of low permeability some laboratory evidence (see discussion in Bear, 1979; Freeze and Cherry, 1979) has suggested that there may be a threshold hydraulic gradient below which flow does not take place. Of greater (but still limited) practical importance is the limitation of Darcy's equation at very high flow rates where turbulent flow occurs. The upper limit to Darcy's equation is usually identified using Reynolds number, which for flow through porous media is defined as

$$Re = \frac{\rho q d}{\mu} \qquad (23)$$

Various definitions are used for d, the pore size length scale (e.g. mean grain size, d_{10} or $(k/n)^{1/2}$). In spite of these differences Bear

(1979) indicates that "Darcy's law is valid as long as the Reynolds number does not exceed some value between 1 and 10." Departures from linearity are discussed by Bear (1979) but are not used in any modeling of infiltration.

The discussion of flow through porous media thus far has developed Darcy's equation for saturated porous media. Infiltration and the generation of runoff often involve unsaturated flow through porous media. As illustrated in Figure 21 when a porous medium is unsaturated part of the porosity void space is occupied by air. The simplest configuration of saturated and unsaturated conditions is that of an unsaturated zone near the surface and a saturated zone at depth (Figure 27).

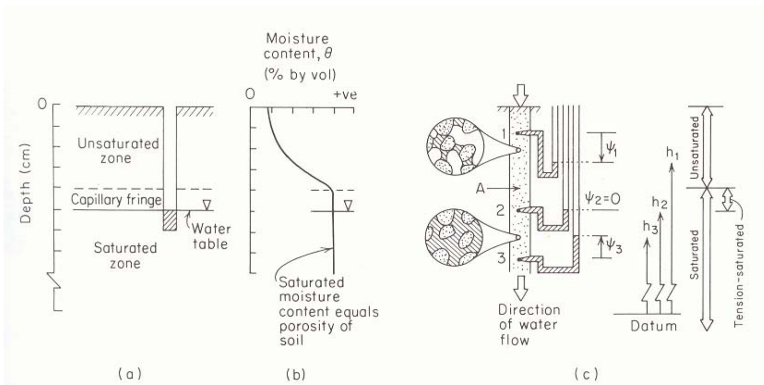


Figure 27. Groundwater conditions near the ground surface. (a) Saturated and unsaturated zones; (b) profile of moisture content versus depth; (c) pressure-head and hydraulic head relationships; insets: water retention under pressure heads less than (top) and greater than (bottom) atmospheric pressure (Freeze/Cherry, Groundwater, © 1979. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey).

We commonly think of the water table as being the boundary between them. The water table is defined as the surface on which the fluid pressure p in the pores of a porous medium is exactly atmospheric. The location of this surface is revealed by the level at which water stands in a shallow well open along its length and penetrating the surficial deposits just deeply enough to encounter standing water at the bottom. If p is measured in terms of gage pressure (i.e. relative to atmospheric pressure), then at the water table p=0. This implies $\psi=0$, and since $h=\psi+z$, the hydraulic head at any point on the water table must be equal to the elevation z of the water table. Positive pressure head occurs in the saturated zone ($\psi > 0$ as indicated by piezometer measurements). Pressure head is zero ($\psi = 0$) at the water table. It follows that pressure head is negative ($\psi < 0$) in the unsaturated zone. This reflects the fact that water in the unsaturated zone is held in the soil pores under tension due to surface-tension forces. A microscopic inspection would reveal a concave meniscus extending from grain to grain across each pore channel (as shown in the upper circular inset in Figure 27c). The radius of curvature on each meniscus reflects the surface tension on that individual, microscopic air-water interface. In reference to this physical mechanism of water retention, negative pressure head is also referred to as tension head or suction head. Above the water table, where $\psi < 0$, piezometers are no longer a suitable instrument for the measurement of h. Instead h must be obtained indirectly from measurements of ψ determined with tensiometers. A tensiometer consists of a porous cup attached to an airtight, water-filled tube. The porous cup is inserted into the soil at the desired depth, where it comes into contact with the soil water and reaches hydraulic equilibrium. The vacuum created at the top of the airtight tube is usually measured by a vacuum gage or pressure transducer attached to the tube above the ground surface, but it can be thought of as acting like an inverted manometer shown for point 1 in the soil in Figure 27c.

Small pores are able to sustain a larger tension head than larger pores, because the surface tension force induced around the pore perimeter is larger relative to the pore cross section area and pressure is force over area. Thus, under hydrostatic conditions (when water is not flowing) water is able to be held higher above the water table in small pores than in larger pores. This effect is illustrated in Figure 27b, and Figure 28 where conceptually (and greatly exaggerated) the height to which water rises in a capillary tube is greater for smaller pores. This leads to the moisture content being a function of the suction head, because as suction increases only the capillary forces in smaller pores can retain water.

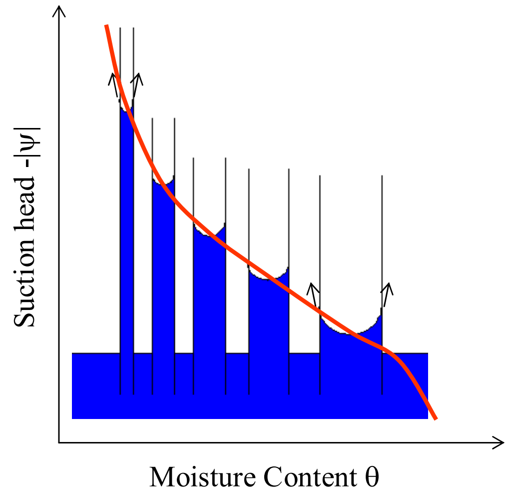
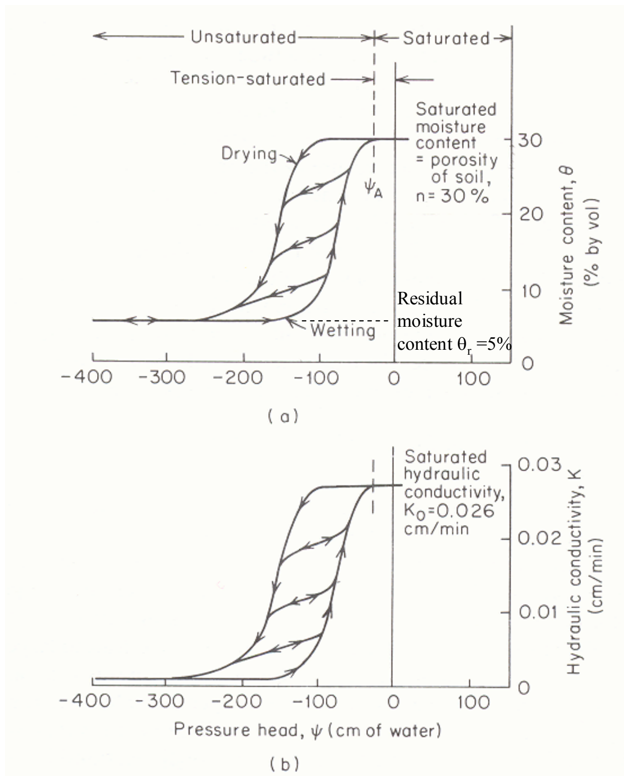


Figure 28. Illustration of capillary rise due to surface tension and relationship between pore size distribution and soil water retention curves.

The flow of water in unsaturated porous media is also governed by Darcy's equation. However, since the moisture content and the size of the pores occupied by water reduces as the magnitude of the suction head is increased (becomes more negative), the paths for water to flow become fewer in number, of smaller cross section and more tortuous. All these effects serve to reduce hydraulic conductivity. Figure 29 illustrates the form of the relationships giving the dependence of suction head and hydraulic conductivity on soil moisture content. This issue is further complicated in that it has been observed experimentally that the $\psi(\theta)$ relationship is hysteretic; it has a different shape when soils are wetting than when they are drying. This also translates into hysteresis in the relationship between hydraulic conductivity and moisture content. The physical processes responsible for hysteresis are discussed by Bear (1979). The curves illustrating the relationship between ψ, θ and K are referred to as *soil water characteristic* curves, or *soil water retention* curves. While hysteresis can have a significant influence on soil-moisture movement, it is difficult to model mathematically and is therefore not commonly incorporated in hydrologic models.



See Online Resource

View the animation of Hysteresis

Figure 29. Characteristic curves relating hydraulic conductivity and moisture content to pressure head for a naturally occurring sand soil (Freeze/Cherry, Groundwater, © 1979. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey).

Note in Figure 29 that the pressure head is 0 when the moisture content equals the porosity, i.e. is saturated, and that the water content changes little as tension increases up to a point of inflection. This more or less distinct point represents the tension at which significant volumes of air begin to appear in the soil pores and is called the air-entry tension, ψ_a. This retention of soil moisture at (or practically close to) saturation for pressures less than atmospheric gives rise to the capillary fringe illustrated in Figures 12 and 28. The capillary fringe plays an important role in the generation of saturation excess runoff where the water table is close to the surface, and also in the generation of return flow and subsurface storm flow as described above.

In Figure 29 pressure head was given as the independent variable on the x-axis. It is sometimes more convenient to think of moisture content as the independent variable. Figure 30 gives an example of the soil water characteristic curves with moisture content as the independent variable. This representation has the advantage of avoiding some of the problem of hysteresis, because $K(\theta)$ is less hysteretic than $K(\psi)$ (Tindall et al., 1999).

Note also in Figure 29 that as tension head is increased a point is reached where moisture content is no longer reduced. A certain amount of water can not be drained from the soil, even at high tension head, due to being retained in disconnected pores and immobile films. This is called the *residual moisture content* or in some cases the *irreducible moisture content* θ_r. For practical purpose flow only occurs in soil for moisture contents between saturation, n, and the residual water content θ_r. This range is referred to as the *effective porosity* $\theta_e=n-\theta_r$. When considering flow in unsaturated soil, moisture content is sometimes quantified using the *effective saturation* defined to scale the range from θ_r to n between 0 and 1.

$$S_e = \frac{\theta - \theta_r}{n - \theta_r} \tag{24}$$

The soil water characteristic curves are a unique property of each soil, related to the size distribution and structure of the pore space. For a specific soil the soil water characteristic functions can be determined experimentally through drainage experiments. For practical purposes it is convenient to mathematically represent the characteristic functions using equations and a number of empirical equations have been proposed. Three functional forms proposed by Brooks and Corey (1966), Van Genuchten (1980) and Clapp and Hornberger (1978) are listed. There are no fundamental differences between these equations, they are simply convenient mathematical expressions that approximately fit the empirical shape of many soil characteristic functions.

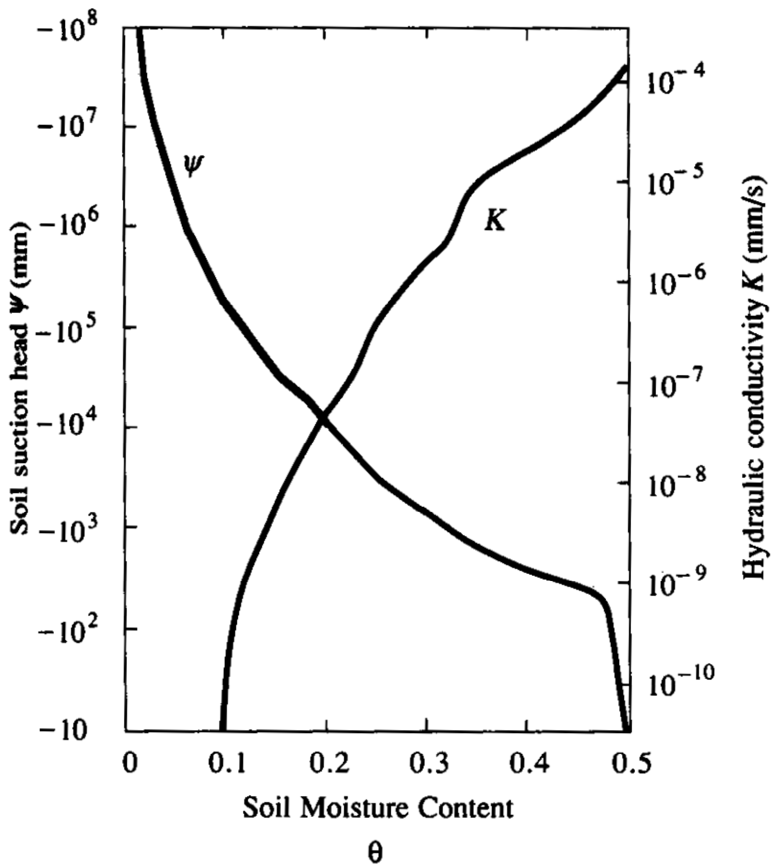


Figure 30. Variation of soil suction head, $|\psi|$, and hydraulic conductivity, K, with moisture content (from Chow et al., 1988).

Brooks and Corey (1966):

$$
\begin{aligned}
|\psi(S_e)| &= |\psi_a| S_e^{-b} \\
K(S_e) &= K_{sat} S_e^{c}
\end{aligned}
\tag{25}
$$

van Genuchten (1980):

$$
\begin{aligned}
|\psi(S_e)| &= \frac{1}{\alpha}\left(S_e^{-1/m} - 1\right)^{1-m} \\
K(S_e) &= K_{sat} S_e^{1/2}(1-(1-S_e^{1/m})^m)^2
\end{aligned}
\tag{26}
$$

Clapp and Hornberger (1978) simplifications of Brooks and Corey functions:

$$\begin{aligned} |\psi(\theta)| &= |\psi_a| \left(\frac{\theta}{n}\right)^{-b} \\ K(\theta) &= K_{sat}\left(\frac{\theta}{n}\right)^{c} \end{aligned} \qquad (27)$$

In these equations K_{sat} is the saturated hydraulic conductivity and b, c, α and m are fitting parameters. The parameter b is referred to as the pore size distribution index because the pore size distribution determines relationship between suction and moisture content (Figure 28). The parameter c is referred to as the pore disconnectedness index because unsaturated hydraulic conductivity is related to how disconnected and tortuous flow paths become as moisture content is reduced. There are theoretical models that relate the soil water retention and hydraulic conductivity characteristic curves. Two common such models are due to Burdine (1953) and Mualem (1976). The Burdine (1953) model suggests $c \approx 2b+3$ in equations (25) and (27). The more recent Mualem (1976) model suggests $c \approx 2b+2.5$. The $K(S_e)$ equation due to van Genuchten uses the Mualem theory. The relative merits of these theories are beyond the scope discussed here. The Brooks and Corey (1966) and Clapp and Hornberger (1978) equations apply only for $\psi < \psi_a$ and assume $\theta = n$ for $\psi > \psi_a$, while the van Genuchten (1980) equations provide for a smoother representation of the inflection point in the characteristic curve near saturation. Clapp and Hornberger (1978) and Cosby et al (1984) statistically analyzed a large number of soils in the United States to relate soil moisture characteristic parameters to soil texture class. Parameter values that Clapp and Hornberger (1978) obtained are given in table 1. The Clapp and Hornberger simplification (equation 27) neglects the additional parameter of residual moisture "which generally gives a better fit to moisture retention data" (Cosby et al., 1984) but was adopted in their analysis because "the large amount of variability in the available data suggests a simpler representation." When using values from table 1, one should be aware of this considerable within-soil-type variability as reflected in the standard deviations listed in table 1. Figures 31 and 32 show the characteristic curves for soils with different textures using the parameter values from table 1. The USDA-ARS Salinity Laboratory has developed a software program Rosetta that estimates the soil moisture retention function $\psi(\theta)$ and hydraulic conductivity function and $K(\theta)$ based upon soil texture class or sand, silt and clay percentages.

See Online Resource

Rosetta soil property program
http://www.ussl.ars.usda.gov/models/rosetta/rosetta.HTM

Table 1. Clapp and Hornberger (1978) parameters for equation (27) based on analysis of 1845 soils. Values in parentheses are standard deviations.

Soil Texture	Porosity n	K_{sat} (cm/hr)	$\lvert\psi_a\rvert$ (cm)	b
Sand	0.395 (0.056)	63.36	12.1 (14.3)	4.05 (1.78)
Loamy sand	0.410 (0.068)	56.16	9 (12.4)	4.38 (1.47)
Sandy loam	0.435 (0.086)	12.49	21.8 (31.0)	4.9 (1.75)
Silt loam	0.485 (0.059)	2.59	78.6 (51.2)	5.3 (1.96)
Loam	0.451 (0.078)	2.50	47.8 (51.2)	5.39 (1.87)
Sandy clay loam	0.420 (0.059)	2.27	29.9 (37.8)	7.12 (2.43)
Silty clay loam	0.477 (0.057)	0.612	35.6 (37.8)	7.75 (2.77)
Clay loam	0.476 (0.053)	0.882	63 (51.0)	8.52 (3.44)
Sandy clay	0.426 (0.057)	0.781	15.3 (17.3)	10.4 (1.64)
Silty clay	0.492 (0.064)	0.371	49 (62.0)	10.4 (4.45)
Clay	0.482 (0.050)	0.461	40.5 (39.7)	11.4 (3.7)

See Online Resource

Excel spreadsheet with table and Figures in electronic form

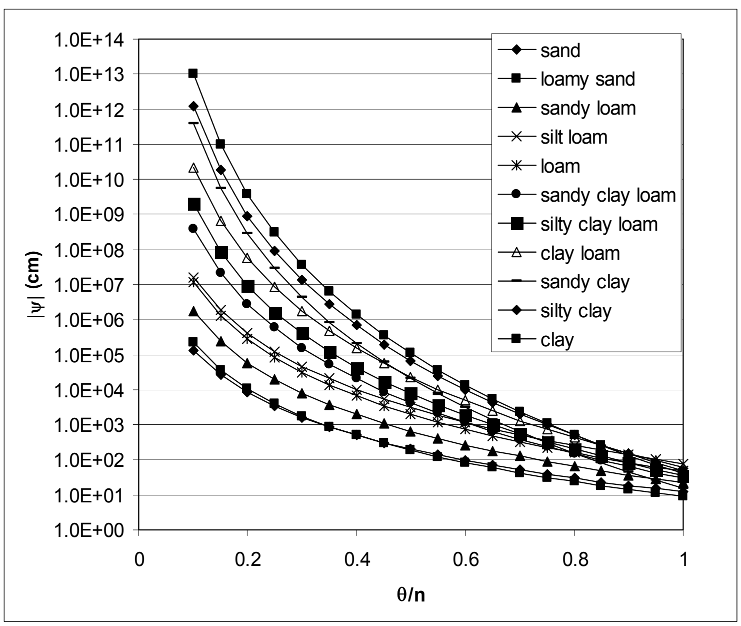


Figure 31. Soil suction head, $|\psi|$, for different soil textures using the Clapp and Hornberger (1978) parameterization (Equation 27).

One can infer from the soil moisture retention curves that as moisture drains from soil under gravitational processes, the hydraulic conductivity is reduced and drainage rate reduced. A point is reached where, for practical purposes, downward drainage has materially ceased. The value of water content remaining in a unit volume of soil after downward gravity drainage has materially ceased is defined as *field capacity*. A difficulty inherent in this definition is that no quantitative specification of what is meant by "materially ceased" is given. Sometimes a definition of drainage for three days following saturation is used. This is adequate for sandy and loamy soils, but is problematic for heavier soils that drain for longer periods.

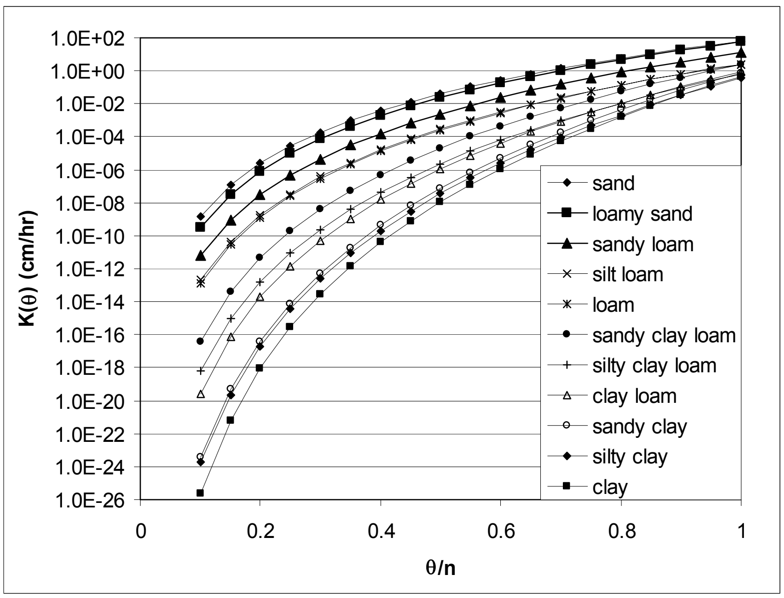


Figure 32. Hydraulic conductivity K(θ) for different soil textures using the Clapp and Hornberger (1978) parameterization (Equation 27).

Because of the difficulty associated with precisely when drainage has materially ceased, a practical approach is to define field capacity as the moisture content corresponding to a specific pressure head. Various studies define field capacity as the moisture content corresponding to a pressure head, ψ, in the range -100 cm to -500 cm with a value of -340 cm being quite common (Dingman, 2002) . The difference between moisture content at saturation and field capacity is referred to as *drainable porosity*, i.e. $n_d = n-\theta(\psi=-100\text{ cm})$.

The notion of field capacity is similar to the notion of residual moisture content defined earlier; however some equations (e.g. equation 27) do not use residual moisture content. A distinction in the definitions can be drawn in that residual moisture content is a theoretical value below which there is no flow of water in the soil, i.e. hydraulic conductivity is 0, while field capacity is a more empirical quantity practically defined as the moisture content corresponding to a specific negative pressure head.

In nature, water can be removed from a soil that has reached field capacity only by direct evaporation or by plant uptake. Plants cannot exert suctions stronger than about -15000 cm and when the water content is reduced to the point corresponding to that value on the moisture characteristic curve, transpiration ceases and plants wilt. This water content is called the *permanent wilting point* θ_{pwp}. The difference between the field capacity and permanent wilting point is the water available for plant use, called *plant available water content*, $\theta_a = \theta_{fc}-\theta_{pwp}$. Although most important for irrigation scheduling in agriculture this is relevant for runoff generation processes because during dry spells vegetation may reduce the surface water content to a value between field capacity and permanent wilting point. The antecedent moisture content plays a role in the generation of runoff. Figure 33 shows a classification of water status in soils based on pressure head. Figure 34 shows ranges of porosity, field capacity and wilting point for soils of various textures.

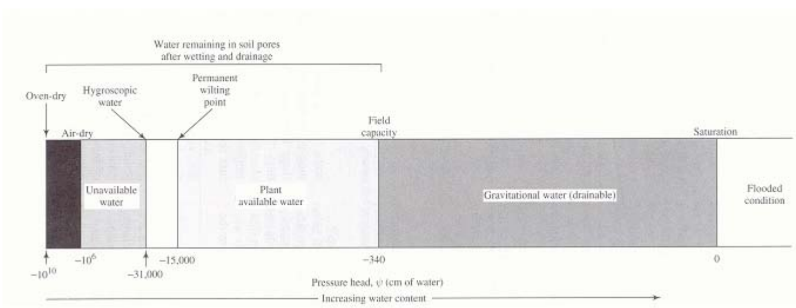


Figure 33. Soil-water status as a function of pressure (tension). Natural soils do not have tensions exceeding about -31000 cm; in this range water is absorbed from the air (Dingman, Physical Hydrology, 2/E, © 2002. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey).

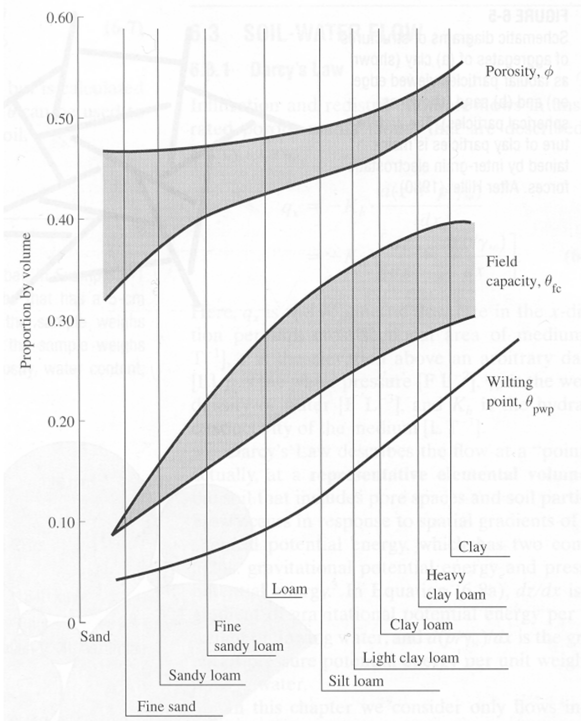


Figure 34. Ranges of porosities, field capacities, and permanent wilting points for soils of various textures (Dingman, Physical Hydrology, 2/E, © 2002. Electronically reproduced by permission of Pearson Education, Inc., Upper Saddle River, New Jersey).

Exercises

See Online Resource

Do the Chapter 4 quiz

1. Infiltration capacity is likely to be larger where:
   A. Hydraulic conductivity is small
   B. Hydraulic conductivity is large

2. Infiltration excess overland flow is likely to be larger where
   A. Hydraulic conductivity is large
   B. Hydraulic conductivity is small

3. Hydraulic conductivity is likely to be large for:
   A. Sandy soils
   B. Clayey soils

4. Porosity is defined as:
   A. Volume of voids/Total volume
   B. Volume of voids/Volume of solids
   C. Volume of water/Volume of solids
   D. Volume of air/Volume of water
   E. Mass of water/Density of soil

5. Volumetric moisture content is defined as:
   A. Volume of air/Volume of water
   B. Mass of water/Density of soil
   C. Volume of water/Total volume
   D. Volume of water/Volume of solids
   E. Volume of voids/Total volume

6. Degree of saturation is defined as:
   A. Volumetric moisture content/Porosity
   B. Volume of water/Volume of voids
   C. Volume of water/Total volume
   D. Volume of water/Volume of solids
   E. Both A and B
   F. Both A and C
   G. Both B and C
   H. A, B, C
   I. B, C, D
   J. A, B, C, D

7. Field and oven-dry weights of a soil sample taken with a 10 cm long by 5 cm diameter cylindrical tube are given in the accompanying table. Assuming ρ_m=2650 kg/m^3 = 2.65 g/cm^3, calculate the volumetric soil moisture content, degree of saturation, bulk density and porosity of those soils.

Field mass	**g**	**302.5**
Oven dry mass	**g**	**264.8**
Bulk Density	**g/cm^3**	
Porosity		
Volumetric soil moisture content		
Degree of saturation		

8. Field and oven-dry weights of a soil sample taken with a 10 cm long by 5 cm diameter cylindrical tube are given in the accompanying table. Assuming ρ_m=2650 kg/m^3 = 2.65 g/cm^3, calculate the volumetric soil moisture content, degree of saturation, bulk density and porosity of those soils.

Field mass	**g**	**390.5**
Oven dry mass	**g**	**274.5**
Bulk Density	**g/cm^3**	
Porosity		
Volumetric soil moisture content		
Degree of saturation		

9. Indicate which (more than one) of the following instruments may be used to measure soil moisture:
   A. Electrical resistance block
   B. Infrared satellite sensor
   C. Capacitance probe
   D. Time domain reflectometry probe
   E. Thermometer
   F. X-Ray sensor
   G. Microwave satellite sensor
   H. Hygrometer
   I. Neutron probe

10. Plot a grain size distribution curve and determine the soil texture for the following soil sieve analysis data.

Diameter (mm)	Percentage passing
50	100
19	100
9.5	100
4.76	98
2	95
0.42	80
0.074	60
0.020	42
0.005	35
0.002	30

11. Plot a grain size distribution curve and determine the soil texture for the following soil sieve analysis data.

Diameter (mm)	Percentage passing
50	100
19	100
9.5	100
4.76	95
2	92
0.42	80
0.074	70
0.020	65
0.005	40
0.002	20

12. Hydraulic conductivity is determined in a Darcy experiment conducted using water at 20 ºC to be 30 cm/hr. The viscosity of water at 20 ºC is 1.05×10^{-3} N s m^{-2}. Using $g=9.81$ m/s^2 and $\rho_w=1000$ kg/m^3 calculate the intrinsic permeability of this material.

13. Following is data from a Darcy experiment using the notation depicted in figure 25. Fill in the blanks and calculate the hydraulic conductivity. The internal diameter of the circular tube used was 10 cm and the length Δl, between piezometers, 40 cm. This experiment is conducted at 20 ºC.

h_1 (cm)	70
h_2 (cm)	58
z_1 (cm)	50
z_2 (cm)	30
n	0.32
Q (l/hr)	0.5
ψ_1 (cm)	
ψ_2 (cm)	
p_1 (Pa)	
p_2 (Pa)	
dh/dl	
q (cm/hr)	
K (cm/hr)	
k (cm^2)	
V (cm/hr)	
Re	

14. Following is data from a Darcy experiment using the notation depicted in figure 25. Fill in the blanks and calculate the hydraulic conductivity. The internal diameter of the circular tube used was 10 cm and the length Δl, between piezometers, 40 cm. This experiment is conducted at 20 °C.

h_1 (cm)	56
h_2 (cm)	35
z_1 (cm)	50
z_2 (cm)	30
n	0.4
Q (l/hr)	2.2
ψ_1	
ψ_2	
p_1 (Pa)	
p_2 (Pa)	
dh/dl	
q (cm/hr)	
K (cm/hr)	
k (cm^2)	
V (cm/hr)	
Re	

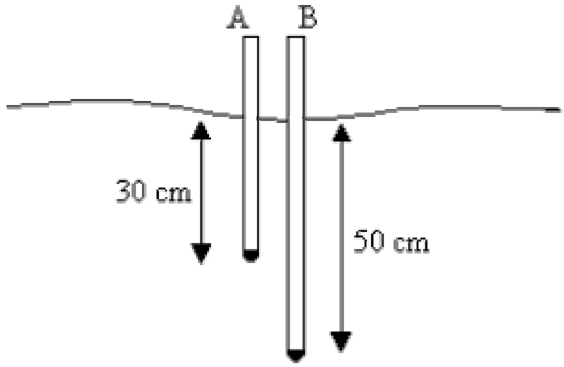
15. Consider the following soil with parameters from Table 1. Evaluate the field capacity moisture content, θ_{fc}, at which pressure head is -340 cm, permanent wilting point moisture content, θ_{fc}, at which the pressure head is -15000 cm and plant available water, θ_a, using the Clapp and Hornberger (1978) soil moisture characteristic parameterization.

Texture	Porosity n	$\vert\psi_a\vert$ (cm)	b	θ_{fc}	θ_{pwp}	θ_a
sand	0.395	12.1	4.05			

16. Consider the following soil with parameters from Table 1. Evaluate the field capacity moisture content, θ_{fc}, at which pressure head is -340 cm, permanent wilting point moisture content, θ_{fc}, at which the pressure head is -15000 cm and plant available water, θ_a, using the Clapp and Hornberger (1978) soil moisture characteristic parameterization.

Texture	Porosity n	$\vert\psi_a\vert$ (cm)	b	θ_{fc}	θ_{pwp}	θ_a
loamy sand	0.41	9	4.38			

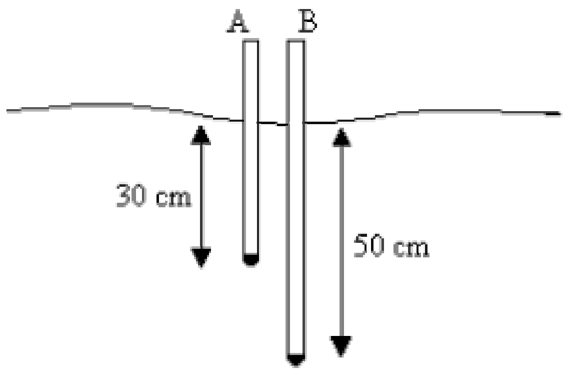
17. Consider the following experimental situation. A and B are vertical tensiometers that measure pore water pressure (tension) relative to atmospheric pressure, at depths 30 and 50 cm below the ground.



Following are pressure measurements recorded at A and B. Negative denotes suction. Evaluate the pressure head at A and B, and total head at A and B using the surface as a datum. Indicate the direction of flow (i.e. downwards into the ground from A to B, or upwards from B to A).

Pressure at A (Pa)	-4000
Pressure at B (Pa)	-3000
ψ at A (cm)	
ψ at B (cm)	
Total head at A (cm)	
Total head at B (cm)	
Direction of flow	

18. Consider the following experimental situation. A and B are vertical tensiometers that measure pore water pressure (tension) relative to atmospheric pressure, at depths 30 and 50 cm below the ground.



Following are pressure measurements recorded at A and B. Negative denotes suction. Evaluate the pressure head at A and B, and total head at A and B using the surface as a datum. Indicate the direction of flow (i.e. downwards into the ground from A to B, or upwards from B to A).

Pressure at A (Pa)	-5500
Pressure at B (Pa)	-3000
ψ at A (cm)	
ψ at B (cm)	
Total head at A (cm)	
Total head at B (cm)	
Direction of flow	

References

Bear, J., (1979), Hydraulics of Groundwater, McGraw-Hill, New York, 569 p.

Bras, R. L., (1990), Hydrology, an Introduction to Hydrologic Science, Addison-Wesley, Reading, MA, 643 p.

Brooks, R. H. and A. T. Corey, (1966), "Properties of Porous Media Affecting Fluid Flow," J. Irrig. and Drain. ASCE, 92(IR2): 61-88.

Burdine, N. T., (1953), "Relative Permeability Calculations from Pore Size Distribution Data," Petroleum Transactions AIME, 198: 71-78.

Chow, V. T., D. R. Maidment and L. W. Mays, (1988), Applied Hydrology, McGraw Hill, 572 p.

Clapp, R. B. and G. M. Hornberger, (1978), "Empirical Equations for Some Soil Hydraulic Properties," Water Resources Research, 14: 601-604.

Cosby, B. J., G. M. Hornberger, R. B. Clamp and T. R. Ginn, (1984), "A Statistical Exploration of the Relationships of Soil Moisture Characteristics to the Physical Properties of Soils," Water Resources Research, 20(6): 682-690.

Dingman, S. L., (2002), Physical Hydrology, 2nd Edition, Prentice Hall, 646 p.

Freeze, R. A. and J. A. Cherry, (1979), Groundwater, Prentice Hall, Englewood Cliffs, 604 p.

Hillel, D., (1980), Fundamentals of Soil Physics, Academic Press, New York, NY.

Mualem, Y., (1976), "A New Model for Predicting the Hydraulic Conductivity of Unsaturated Porous Media," Water Resources Research, 12(3): 513-522.

Tindall, J. A., J. R. Kunkel and D. E. Anderson, (1999), Unsaturated Zone Hydrology for Scientists and Engineers, Prentice Hall, Upper Saddle River, New Jersey, 624 p.

Van Genuchten, M. T., (1980), "A Closed Form Equation for Predicting the Hydraulic Conductivity of Unsaturated Soils," Soil Sci. Soc. Am. J., 44: 892-898.