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GROUNDWATER

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rounded particles, have lower porosities than soils rich in platy clay minerals; and poorly sorted deposits [Figure 2.11(b)] have lower porosities than well-sorted deposits [Figure 2.11(a)].

The porosity n can be an important controlling influence on hydraulic conductivity K. In sampling programs carried out within deposits of well-sorted sand or in fractured rock formations, samples with higher n generally also have higher K. Unfortunately, the relationship does not hold on a regional basis across the spectrum of possible rock and soil types. Clay-rich soils, for example, usually have higher porosities than sandy or gravelly soils but lower hydraulic conductivities. In Section 8.7 techniques will be presented for the estimation of hydraulic conductivity from porosity and from grain-size analyses.

The porosity n is closely related to the *void ratio* e, which is widely used in soil mechanics. The void ratio is defined as $e = V_v/V_s$, and e is related to n by

$$e = \frac{n}{1-n} \quad \text{or} \quad n = \frac{e}{1+e} \tag{2.40}$$

Values of e usually fall in the range 0–3.

The measurement of porosity on soil samples in the laboratory will be treated in Section 8.4.

2.6 Unsaturated Flow and the Water Table

Up until this point, Darcy's law and the concepts of hydraulic head and hydraulic conductivity have been developed with respect to a *saturated* porous medium, that is, one in which all the voids are filled with water. It is clear that some soils, especially those near the ground surface, are seldom saturated. Their voids are usually only partially filled with water, the remainder of the pore space being taken up by air. The flow of water under such conditions is termed *unsaturated* or *partially saturated*. Historically, the study of unsaturated flow has been the domain of soil physicists and agricultural engineers, but recently both soil scientists and groundwater hydrologists have recognized the need to pool their talents in the development of an integrated approach to the study of subsurface flow, both saturated and unsaturated.

Our emphasis in this section will be on the hydraulics of *liquid-phase* transport of water in the unsaturated zone. We will not discuss *vapor-phase* transport, nor will we consider *soil water-plant interactions*. These latter topics are of particular interest in the agricultural sciences and they play an important role in the interpretation of soil geochemistry. More detailed consideration of the physics and chemistry of moisture transfer in unsaturated soils can be found at an introductory level in Baver et al. (1972) and at a more advanced level in Kirkham and Powers (1972) and Childs (1969).

Moisture Content

If the total unit volume V_T of a soil or rock is divided into the volume of the solid portion V_s, the volume of the water V_w, and the volume of the air V_a, the *volumetric moisture content* θ is defined as $\theta = V_w/V_T$. Like the porosity n, it is usually reported as a decimal fraction or a percent. For saturated flow, $\theta = n$; for unsaturated flow, $\theta < n$.

Water Table

The simplest hydrologic configuration of saturated and unsaturated conditions is that of an *unsaturated zone* at the surface and a *saturated zone* at depth [Figure 2.12(a)]. We commonly think of the water table as being the boundary between the two, yet we are aware that a saturated capillary fringe often exists above the water table. With this type of complication lurking in the background, we must take care to set up a consistent set of definitions for the various saturated-unsaturated concepts.

The *water table* is best 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 in the bottom. If p is measured in gage pressure, then on 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 at that point. On figures we will often indicate the position of the water table by means of a small inverted triangle, as in Figure 2.12(a).

Negative Pressure Heads and Tensiometers

We have seen that $\psi > 0$ (as indicated by piezometer measurements) in the saturated zone and that $\psi = 0$ on the water table. It follows that $\psi < 0$ in the unsaturated zone. This reflects the fact that water in the unsaturated zone is held in the soil pores under 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 on Figure 2.12(c)]. 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, soil physicists often call the pressure head ψ, when $\psi < 0$, the *tension head* or *suction head.* In this text, on the grounds that one concept deserves only one name, we will use the term *pressure head* to refer to both positive and negative ψ.

Regardless of the sign of ψ, the hydraulic head h is still equal to the algebraic sum of ψ and z. However, 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*. Kirkham (1964) and S. J. Richards (1965) provide detailed descriptions of the design

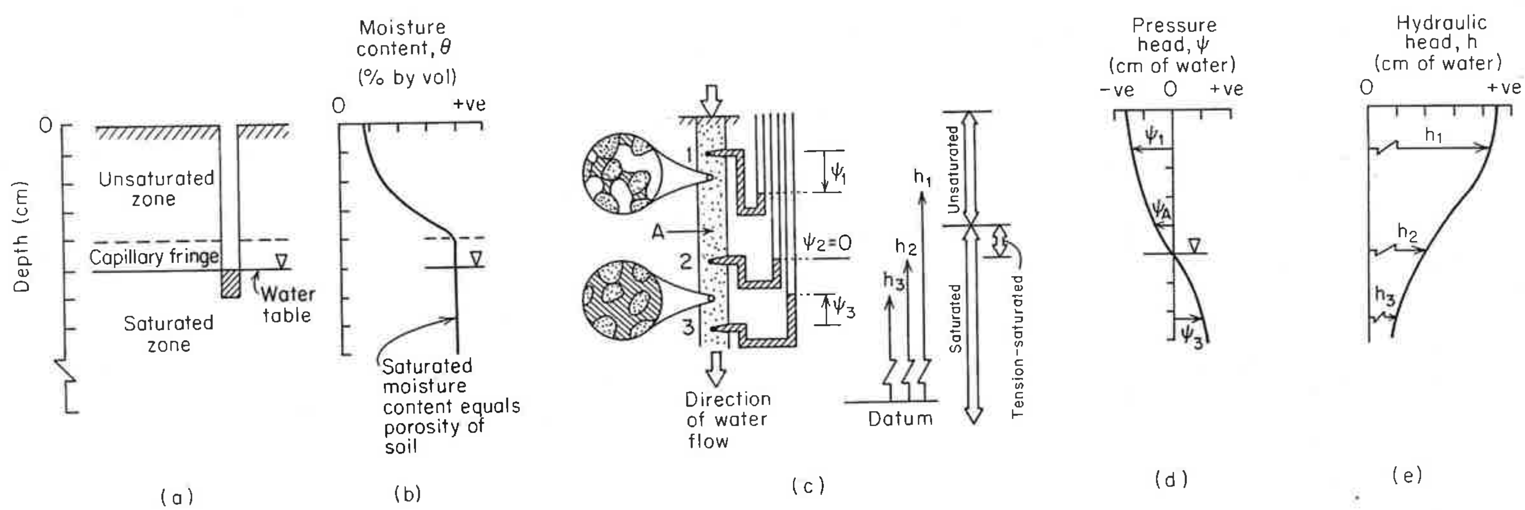

Figure 2.12 Groundwater conditions near the ground surface. (a) Saturated and unsaturated zones; (b) profile of moisture content versus depth; (c) pressure-head and hydraulic-head relationships; insets: water retention under pressure heads less than (top) and greater than (bottom) atmospheric; (d) profile of pressure head versus depth; (e) profile of hydraulic head versus depth.

and use of these instruments. Very briefly, 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 equilibration process involves the passage of water through the porous cup from the tube into the soil. The vacuum created at the top of the airtight tube is a measure of the pressure head in the soil. It is usually measured by a vacuum gage attached to the tube above the ground surface, but it can be thought of as acting like the inverted manometer shown for point 1 in the soil profile of Figure 2.12(c). To obtain the hydraulic head h, the negative ψ value indicated by the vacuum gage on a tensiometer must be added algebraically to the elevation z of the point of measurement. In Figure 2.12(c) the instrument at point 1 is a tensiometer; the one at point 3 is a piezometer. The diagram is, of course, schematic. In practice, the tensiometer would be a tube with a gage and a porous cup at the base; the piezometer would be an open pipe with a well point at the base.

Characteristic Curves of the Unsaturated Hydraulic Parameters

There is a further complication to the analysis of flow in the unsaturated zone. Both the moisture content θ and the hydraulic conductivity K are functions of the pressure head ψ. On reflection, the first of these conditions should come as no great surprise. In that soil moisture is held between the soil grains under surface-tension forces that are reflected in the radius of curvature of each meniscus, we might expect that higher moisture contents would lead to larger radii of curvature, lower surface-tension forces, and lower tension heads (i.e., less-negative pressure heads). Further, it has been observed experimentally that the θ–ψ relationship is hysteretic; it has a different shape when soils are wetting than when they are drying. Figure 2.13(a) shows the hysteretic functional relationship between θ and ψ for a naturally occurring sand soil (after Liakopoulos, 1965a). If a sample of this soil were saturated at a pressure head greater than zero and the pressure was then lowered step by step until it reached levels much less than atmospheric ($\psi \ll 0$), the moisture contents at each step would follow the *drying curve* (or *drainage curve*) on Figure 2.13(a). If water were then added to the dry soil in small steps, the pressure heads would take the return route along the *wetting curve* (or *imbibition curve*). The internal lines are called *scanning curves*. They show the course that θ and ψ would follow if the soil were only partially wetted, then dried, or vice versa.

One would expect, on the basis of what has been presented thus far, that the moisture content θ would equal the porosity n for all $\psi > 0$. For coarse-grained soils this is the case, but for fine-grained soils this relationship holds over a slightly larger range $\psi > \psi_a$, where ψ_a is a small negative pressure head [Figure 2.13(a)] known as the *air entry pressure head*. The corresponding pressure p_a is called the *air entry pressure* or the *bubbling pressure*.

Figure 2.13(b) displays the hysteretic curves relating the hydraulic conductivity K to the pressure head ψ for the same soil. For $\psi > \psi_a$, $K = K_0$, where K_0

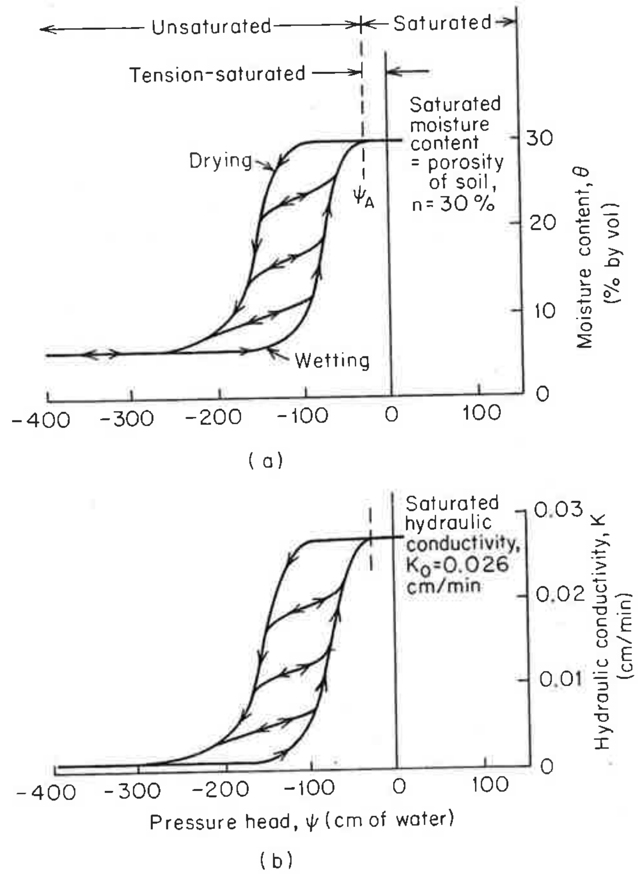

Figure 2.13 Characteristic curves relating hydraulic conductivity and moisture content to pressure head for a naturally occurring sand soil (after Liakopoulos, 1965a).

is now known as the *saturated hydraulic conductivity*. Since $K = K(\psi)$ and $\theta = \theta(\psi)$, it is also true that $K = K(\theta)$. The curves of Figure 2.13(b) reflect the fact that the hydraulic conductivity of an unsaturated soil increases with increasing moisture content. If we write Darcy's law for unsaturated flow in the x direction in an isotropic soil as

$$v_x = -K(\psi)\frac{\partial h}{\partial x} \tag{2.41}$$

we see that the existence of the $K(\psi)$ relationship implies that, given a constant hydraulic gradient, the specific discharge v increases with increasing moisture content.

In actual fact, it would be impossible to hold the hydraulic gradient constant while increasing the moisture content. Since $h = \psi + z$ and $\theta = \theta(\psi)$, the hydraulic head h is also affected by the moisture content. In other words, a hydraulic-head gradient infers a pressure-head gradient (except in pure gravity flow), and this in turn infers a moisture-content gradient. In Figure 2.12, the vertical profiles for these three variables are shown schematically for a hypothetical case of downward infiltration from the surface. Flow must be downward because the hydraulic heads displayed in Figure 2.12(e) decrease in that direction. The large positive values of h infer that $|z| \gg |\psi|$. In other words, the $z = 0$ datum lies at some depth. For a real case, these three profiles would be quantitatively interlinked through the $\theta(\psi)$ and $K(\psi)$ curves for the soil at the site. For example, if the $\theta(\psi)$ curve were known for the soil and the $\theta(z)$ profile measured in the field, then the $\psi(z)$ profile, and hence the $h(z)$ profile, could be calculated.

The pair of curves $\theta(\psi)$ and $K(\psi)$ shown in Figure 2.13 are characteristic for any given soil. Sets of measurements carried out on separate samples from the same homogeneous soil would show only the usual statistical variations associated with spatially separated sampling points. The curves are often called the *characteristic curves*. In the saturated zone we have the two fundamental hydraulic parameters K_0 and n; in the unsaturated zone these become the functional relationships $K(\psi)$ and $\theta(\psi)$. More succinctly,

$$\begin{aligned} \theta &= \theta(\psi) \qquad \psi < \psi_a \\ \theta &= n \qquad \psi \geq \psi_a \end{aligned} \tag{2.42}$$

$$\begin{aligned} K &= K(\psi) \qquad \psi < \psi_a \\ K &= K_0 \qquad \psi \geq \psi_a \end{aligned} \tag{2.43}$$

Figure 2.14 shows some hypothetical single-valued characteristic curves (i.e., without hysteresis) that are designed to show the effect of soil texture on the shape of the curves. For a more complete description of the physics of moisture retention in unsaturated soils, the reader is directed to White et al. (1971).

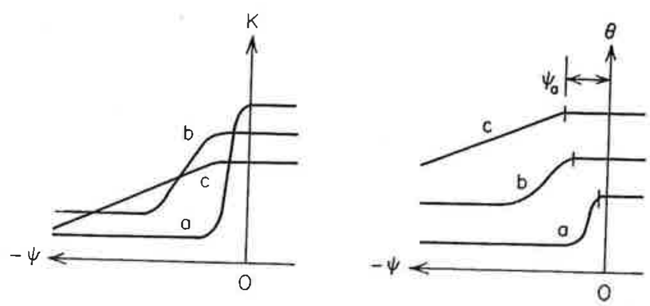

Figure 2.14 Single-valued characteristic curves for three hypothetical soils. (a) Uniform sand; (b) silty sand; (c) silty clay.

Saturated, Unsaturated, and Tension-Saturated Zones

It is worthwhile at this point to summarize the properties of the saturated and unsaturated zones as they have been unveiled thus far. For the *saturated zone*, we can state that:

1. It occurs below the water table.
2. The soil pores are filled with water, and the moisture content θ equals the porosity n.
3. The fluid pressure p is greater than atmospheric, so the pressure head ψ (measured as gage pressure) is greater than zero.
4. The hydraulic head h must be measured with a piezometer.
5. The hydraulic conductivity K is a constant; it is not a function of the pressure head ψ.

For the *unsaturated zone* (or, as it is sometimes called, the *zone of aeration* or the *vadose zone*):

1. It occurs above the water table and above the capillary fringe.
2. The soil pores are only partially filled with water; the moisture content θ is less than the porosity n.
3. The fluid pressure p is less than atmospheric; the pressure head ψ is less than zero.
4. The hydraulic head h must be measured with a tensiometer.
5. The hydraulic conductivity K and the moisture content θ are both functions of the pressure head ψ.

In short, for saturated flow, $\psi > 0$, $\theta = n$, and $K = K_0$; for unsaturated flow, $\psi < 0$, $\theta = \theta(\psi)$, and $K = K(\psi)$.

The capillary fringe fits into neither of the groupings above. The pores there are saturated, but the pressure heads are less than atmospheric. A more descriptive name that is now gaining acceptance is the *tension-saturated zone*. An explanation of its seemingly anomalous properties can be discovered in Figure 2.13. It is the existence of the air entry pressure head $\psi_a < 0$ on the characteristic curves that is responsible for the existence of a capillary fringe. ψ_a is the value of ψ that will exist at the top of the tension-saturated zone, as shown by ψ_A for point A in Figure 2.12(d). Since ψ_a has greater negative values in clay soils than it does in sands, these fine-grained soils develop thicker tension-saturated zones than do coarse-grained soils.

Some authors consider the tension-saturated zone as part of the saturated zone, but in that case the water table is no longer the boundary between the two zones. From a physical standpoint it is probably best to retain all three zones—saturated, tension-saturated, and unsaturated—in one's conception of the complete hydrologic system.

A point that follows directly from the foregoing discussion in this section may warrant a specific statement. In that fluid pressures are less than atmospheric, there can be no natural outflow to the atmosphere from an unsaturated or tension-saturated face. Water can be transferred from the unsaturated zone to the atmosphere by evaporation and transpiration, but natural outflows, such as springs on streambanks or inflows to well bores, must come from the saturated zone. The concept of a saturated seepage face is introduced in Section 5.5 and its importance in relation to hillslope hydrology is emphasized in Section 6.5.

Perched and Inverted Water Tables

The simple hydrologic configuration that we have considered thus far, with a single unsaturated zone overlying the main saturated groundwater body, is a common one. It is the rule where homogeneous geologic deposits extend to some depth. Complex geological environments, on the other hand, can lead to more complex saturated-unsaturated conditions. The existence of a low-permeability clay layer in a high-permeability sand formation, for example, can lead to the formation of a discontinuous saturated lense, with unsaturated conditions existing both above and below. If we consider the line *ABCDA* in Figure 2.15 to be the $\psi = 0$ isobar, we can refer to the *ABC* portion as a *perched water table* and *ADC* as an *inverted water table*. *EF* is the true water table.

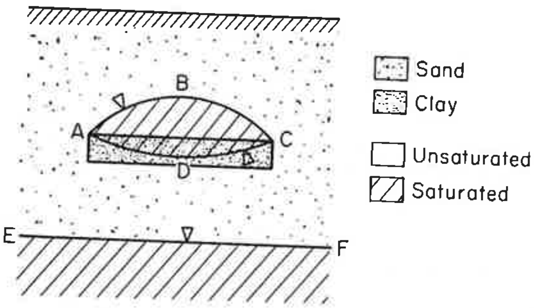

Figure 2.15 Perched water table *ABC*, inverted water table *ADC*, and true water table *EF*.

Saturated conditions can be discontinuous in time as well as space. Heavy rainfall can lead to the formation of a temporary saturated zone at the ground surface, its lower boundary being an inverted water table underlain by unsaturated conditions. Saturated zones of this type dissipate with time under the influence of downward percolation and evaporation from the surface. In Chapter 6 we will examine the interactions of rainfall and infiltration in saturated-unsaturated systems in greater detail.

Multiphase Flow

The approach to unsaturated flow outlined in this section is the one used almost universally by soil physicists, but it is, at root, an approximate method. Unsaturated flow is actually a special case of *multiphase flow* through porous media, with two phases, air and water, coexisting in the pore channels. Let θ_w be the volumetric moisture content (previously denoted by θ) and θ_a be the volumetric air content, defined analogously to θ_w. There are now two fluid pressures to consider: p_w for the water phase and p_a for the air phase; and two pressure heads, ψ_w and ψ_a. Each soil now possesses two characteristic curves of fluid content versus pressure head, one for the water, $\theta_w(\psi_w)$, and one for the air, $\theta_a(\psi_a)$.

When it comes to conductivity relationships, it makes sense to work with the permeability k [Eq. (2.28)] rather than the hydraulic conductivity K, since k is independent of the fluid and K is not. The flow parameters k_w and k_a are called the *effective permeabilities* of the medium to water and air. Each soil has two characteristic curves of effective permeability versus pressure head, one for water, $k_w(\psi_w)$, and one for air, $k_a(\psi_a)$.

The single-phase approach to unsaturated flow leads to techniques of analysis that are accurate enough for almost all practical purposes, but there are some unsaturated flow problems where the multiphase flow of air and water must be considered. These commonly involve cases where a buildup in air pressure in the entrapped air ahead of a wetting front influences the rate of propagation of the front through a soil. Wilson and Luthin (1963) encountered the effects experimentally, Youngs and Peck (1964) provide a theoretical discussion, and McWhorter (1971) presents a complete analysis. As will be shown in Section 6.8, air entrapment also influences water-table fluctuations. Bianchi and Haskell (1966) discuss air entrapment problems in a field context, and Green et al. (1970) describe a field application of the multiphase approach to the analysis of a subsurface flow system.

Much of the original research on multiphase flow through porous media was carried out in the petroleum industry. Petroleum reservoir engineering involves the analysis of three-phase flow of oil, gas, and water. Pirson (1958) and Amyx et al. (1960) are standard references in the field. Stallman (1964) provides an interpretive review of the petroleum multiphase contributions as they pertain to groundwater hydrology.

The two-phase analysis of unsaturated flow is an example of *immiscible displacement*; that is, the fluids displace each other without mixing, and there is a distinct fluid-fluid interface within each pore. The simultaneous flow of two fluids that are soluble in each other is termed *miscible displacement*, and in such cases a distinct fluid-fluid interface does not exist. Bear (1972) provides an advanced theoretical treatment of both miscible and immiscible displacement in porous media. In this text, the only examples of immiscible displacement are those that have been discussed in this subsection. In the rest of the text, unsaturated flow will be treated as a single-phase problem using the concepts and approach of the first part of this section. The most common occurrences of miscible displacement in groundwater hydrology involve the mixing of two waters with different chemistry (such as seawater and fresh-water, or pure water and contaminated water). The transport processes associated with miscible displacement and the techniques of analysis of groundwater contamination will be discussed in Chapter 9.