CHAPTER

13

Infiltration

In previous chapters we have introduced fundamental concepts about water in soil, including static water in soil and water movement in saturated soil. In this chapter we consider infiltration, the tension infiltrometer, and four soil characters (unsaturated hydraulic conductivity, sorptivity, repellency, and mobility) that we can measure with the tension infiltrometer.

13.1 DEFINITION OF INFILTRATION

Infiltration rate may be defined as the meters per unit time of water entering into the soil regardless of the types or values of forces or gradients. The term *hydraulic conductivity*, which has been defined as the meters per day of water seeping into the soil under the pull of gravity or under a unit hydraulic gradient, should not be confused with infiltration rate. Infiltration rate need not refer to saturated conditions. If two raindrops of total volume $2\ \text{mm}^3 = 0.000002\ \text{m}^3$ fall per day on a square meter of soil and are absorbed into the soil, the infiltration rate is $0.000002\ \text{m/day}$.

Water entry into soil is caused by matric and gravitational forces. Therefore, this entry may occur in the lateral and upward directions as well as the downward one (Baver et al., 1972, p. 365). Infiltration normally refers to the downward movement. The matric force usually predominates over the gravitational force during the early stages of water entry into soil, so that observations made during the early stages of infiltration are valid when considering the absence of gravity.

If water infiltrates into a dry soil, a definite *wetting front*, also called a *wet front*, can be observed. This is the boundary between the wetted upper part of the soil and the dry lower part of the soil. If water is infiltrating into soil contained in a clear plastic column, one can observe the progress of the wet front and mark wet fronts as they change with time (Figure 13.1). At present, it is impossible to measure the matric potential exactly at the

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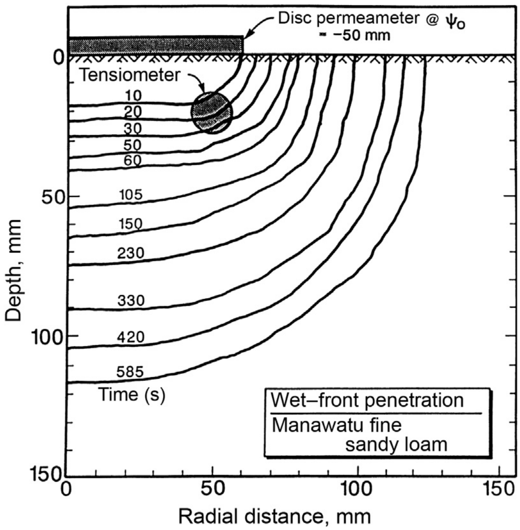

FIGURE 13.1 Wet fronts for a sandy loam soil. *From Kirkham, M.B., Clothier, B.E., ©1994a. Ellipsoidal description of water flow into soil from a surface disc. Trans. Int. Congr. Soil Sci. 2b, 38–39. Reprinted by permission of The International Society of Soil Science.*

wet front, because it progresses too rapidly into the soil. However, one can measure the amount of water infiltrated and the depth and shape of the wet front, and come to important conclusions about the entry of water into the soil. Infiltration is extremely important, because it determines not only the amount of water that will enter a soil, but also the entrainment of the "passenger" chemicals (nutrients and pollutants) dissolved in it.

13.2 FOUR MODELS OF ONE-DIMENSIONAL INFILTRATION

Four models for infiltration into the soil have been developed. They all deal with one-dimensional, downward infiltration into the soil (Baver et al., 1972; pp. 366–371).

13.2.1 Lewis Equation

From work initiated in 1926, Mortimer Reed Lewis, an irrigation engineer at Oregon State College, used the following equation for infiltration:

$$I = \gamma t^{a} \tag{13.1}$$

where I is the cumulative infiltration between time zero and t, and γ and a are constants. The *cumulative infiltration* is the total volume of water infiltrated per unit area of soil surface during a specified time period (Soil

Science Society of America, 2008). It has units such as length (e.g., centimeters). Note that infiltration rate has units of length per unit time (e.g., centimeters per second). Equation (13.1) has been erroneously attributed to A.N. Kostiakov, and often appears in the literature as the "Kostiakov" equation (Swartzendruber, 1993). The parameters in Eqn (13.1) are evaluated by fitting the model to experimental data. By definition, the infiltration rate $i = \mathrm{d}I/\mathrm{d}t$. Thus, the infiltration rate for the Lewis equation is given by

$$i = a\gamma t^{a-1}. \tag{13.2}$$

The Lewis equation has no physical basis. Exponential relations are simple and useful and can cover a wide range of values. They are often the first ones thought of by scientists when trying to describe a process (Don Kirkham, personal communication, undated).

13.2.2 Horton Equation

In the 1930s, Robert E. Horton, a pioneer in the study of infiltration in the field, developed the following equation:

$$i = i_\mathrm{f} + (i_\mathrm{o} - i_\mathrm{f})\exp(-\beta t) \tag{13.3}$$

where i_o is the initial infiltration rate at $t = 0$, i_f is the final constant infiltration rate that is achieved at large times, and β is a soil parameter that describes the rate of decrease of infiltration.

Horton felt that the reduction in infiltration rate with time was largely controlled by factors operating at the soil surface. These included swelling of soil colloids and the closing of small cracks, which progressively sealed the soil surface. He also recognized that a bare soil surface was compacted by raindrops, but crop cover mitigated their effect. Horton's field data showed that the infiltration rate eventually approached a constant value, which was often somewhat smaller than the saturated permeability of the soil. The latter observation was thought to be due to air entrapment.

13.2.3 Green and Ampt Equation

The preceding models are empirical. W. Heber Green and G.A. Ampt in Australia published in 1911 an infiltration equation that was based on a simple physical model of the soil. It has the advantage that the parameters in the equation can be related to physical properties of the soil. Physically, Green and Ampt assumed that the soil was saturated behind the wetting front and that one could define some "effective" matric potential at the wetting front. During infiltration, if the soil surface is held at a constant matric potential or head h_o with associated water content θ_o (e.g., by ponding water over it), water enters the soil behind a sharply defined wet front that moves downward with time (Figure 13.2(A)) (Jury et al., 1991, pp. 131–134). Green and Ampt replaced this process with one that has a discontinuous change in water content at the wetting front (Figure 13.2(B)). In addition, they made the following assumptions: (1) The soil in the wetted region has constant properties (K_o, θ_o, D_o, h_o, where K_o and D_o are the hydraulic conductivity and soil water diffusivity in the Green–Ampt model, respectively) and (2) the matric potential (head) at the moving front is constant and equal to h_F. *Soil water diffusivity* is the unsaturated hydraulic conductivity times the rate of change of soil matric potential with water content. It has units of square millimeters per second. In the 1930s, Ernest Childs in England pointed out that, under unsaturated conditions, water moves according to diffusion equations. It is beyond the scope of this book to discuss the theory of water flow in unsaturated soils, in which partial differential equations are used. However, for a brief, general description, see Kirkham (1994), and for a description that requires calculus, see Jury and Horton (2004, p. 105).

The Green–Ampt model can be used to calculate the infiltration rate into a horizontal soil column initially at a uniform water content θ_i such that $\theta_o > \theta_i$ and an associated matric potential or head h_o maintained at the entry surface for all times >0. Using the assumptions of the model and Darcy's law, the following equation can be derived:

$$i = (\mathrm{d}I/\mathrm{d}t) = \Delta\theta(D_o/2t)^{1/2} \tag{13.4}$$

where i = infiltration rate; I = infiltration; t = time; $\Delta\theta = \theta_o - \theta_i > 0$, $D_o = K_o\Delta h/\Delta\theta$ is the soil water diffusivity of the wet soil region $0 < x < L$, the depth of the wetting front, and $\Delta h = h_o - h_F > 0$, and K_o is the constant hydraulic conductivity of the wet region, and h_F is the matric potential or head of the moving front. Note in this model that the infiltration rate into the soil is proportional to $t^{-1/2}$. A similar expression is obtained for infiltration into a vertical soil column at short times after infiltration begins.

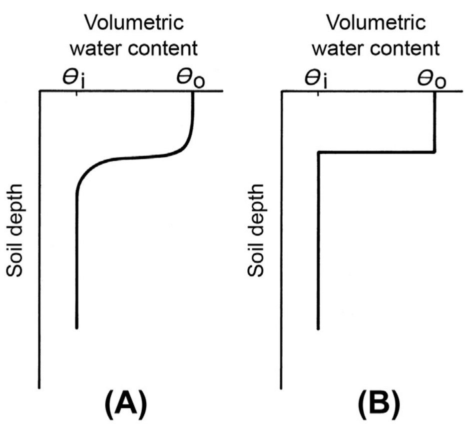

FIGURE 13.2 Water content profiles during infiltration. (A) A profile that actually occurs during infiltration. (B) A profile corresponding to the Green–Ampt infiltration model. *From Jury, W.A., Gardner, W.R., Gardner, W.H., ©1991. Soil Physics, fifth ed. John Wiley & Sons, New York, p. 132. This material is used by permission of John Wiley & Sons, Inc.*

The model has been used as a conceptual aid in visualizing a complex process. Indirect evaluation of h_F has permitted the model to be used in practical applications.

13.2.4 Philip Infiltration Model

J.R. Philip in 1957 suggested an approximate algebraic equation (based on sound physical reasoning) for vertical infiltration under ponded conditions. (See the Appendix, Section 13.11, for a biography of Philip.) The equation, which is simple yet physically well founded, is as follows:

$$I = St^{1/2} + At \tag{13.5}$$

where I is the cumulative infiltration (millimeters), S is the sorptivity (millimeters hour$^{-1/2}$), and A is an empirical constant (millimeters per hour). The first term on the right-hand side of Eqn (13.5) gives the gravity-free absorption into a ponded soil due to capillarity and adsorption. The second term represents the infiltration due to the downward force of gravity. S and A may be found empirically by fitting Eqn (13.5) to infiltration data. Alternatively, these parameters may be derived from the hydraulic properties of the soil. This is not possible for other empirical infiltration equations. For horizontal (gravity-free) infiltration, cumulative infiltration I is given by

$$I = St^{1/2} \tag{13.6}$$

13.3 TWO- AND THREE-DIMENSIONAL INFILTRATION

The previous discussion dealt with one-dimensional infiltration in which water is assumed to flow vertically (or more rarely horizontally) into the soil. Multidimensional infiltration theory is an area of soil physics research dominated by the works of J.R. Philip, who published on the topic in a major paper in 1969. Sequels to his work have been carried out by Raats (1971) in the Netherlands and Wooding (1968) in New Zealand.

According to Jury et al. (1991, p. 143), Wooding (1968) derived an approximate expression for the steady rate of infiltration from a circular pond of radius r_o, overlying a soil in which the hydraulic conductivity-matric potential function was assumed to be

$$K(h) = K_o \exp(\alpha h) \tag{13.7}$$

where K_o and α are constants representing the soil properties and h is the matric potential (head). Wooding (1968) says that the parameter α is defined as the logarithmic derivative of the hydraulic conductivity with respect to capillary potential (what we now call matric potential). Since

$K = K_o$ when $h = 0$, K_o represents the saturated hydraulic conductivity of a soil. Using this expression and a simplified form of the three-dimensional water flow equation, Wooding derived the following equation for the steady infiltration flux rate:

$$i_f = K_o[1 + (4/\pi\alpha r_o)]. \tag{13.8}$$

(Note that in the equation on p. 143 of Jury et al. (1991), there is a mistake. The last symbol, r_o is squared in Jury et al. This is incorrect. Wooding's equation is written correctly in Eqn (13.8) above. Wooding's equation is written correctly, but using different symbols, in Jury and Horton (2004, p. 136).) It is notable that, contrary to one-dimensional flow, the final infiltration rate in Wooding's equation (Eqn (13.8)) exceeds K_o. This occurs because water may enter and move laterally as well as vertically.

Multidimensional infiltration models have utilized difficult mathematics. However, practical advances in infiltration can be made with simple models. For example, a simple, ellipsoidal description of the pattern of wetting to approximate the depth to the wetting front underneath a disc permeameter, set at Ψ_o and supplying water to soil initially at water content θ_n, will be described in Section 13.10 of this chapter.

13.4 REDISTRIBUTION

The term *redistribution* refers to the continued movement of water through a soil profile after irrigation or rainfall has stopped at the soil surface. Redistribution occurs after infiltration and is complex, because the lower part of the profile ahead of the wet front will increase its water content and the upper part of the profile near the surface will decrease its water content, after infiltration ceases. Thus, hysteresis can have an effect on the overall shape of the water content profile. See Jury et al. (1991, p. 144−151) or Jury and Horton (2004, p. 147−155) for a discussion of redistribution and figures illustrating it.

13.5 TENSION INFILTROMETER OR DISC PERMEAMETER

Recognition of the importance of macropores and preferential flow has led to the development of instruments that can be used in the field to control preferential water flow through macropores and soil cracks. Let us first define macropores and see their size in relation to other soil pores. Pores in the soil can be classified into five categories (Clothier, 2008): macropores, with diameters ranging from 75 to >5000 μm; mesopores with diameters ranging from 30 to 75 μm; micropores with diameters ranging from 5 to 30 μm; ultramicropores with diameters ranging from 0.5 to 5 μm; and cryptopores with diameters <0.1 μm. Xylem vessels, by comparison, range in diameter from 8 to 500 μm, with 40 μm a reasonable value to use in calculations (Nobel, 1983, p. 493).

The first practical instrument to control macropore flow was developed in 1981 by Brent E. Clothier of New Zealand and Ian White of Australia (Clothier and White, 1981). This simple instrument was known as the sorptivity tube (Figure 13.3), then as the tension infiltrometer, and later still it evolved into the disc permeameter (Figure 13.4), as described by Perroux and White (1988). Originally, the term "disc permeameter" was used when three-dimensional infiltration was being considered, and the term "tension infiltrometer" was used when one-dimensional infiltration was being considered, but today the terms are used interchangeably. One must state if one is considering one-dimensional or three-dimensional flow when using the instruments. With these instruments, the amount of macropore flow measured is controlled by applying water to soil at water potentials Ψ_o, less than 0. The maximum diameter of vertical pores, connected to the soil surface, through which

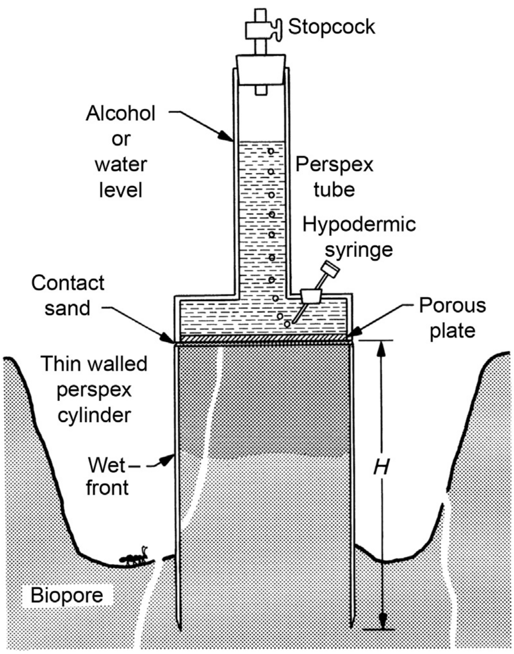

FIGURE 13.3 The sorptivity tube. An ant hole of greater than 0.75 mm in diameter is shown, which has no effect on the infiltration process. $H = 9.5$ cm. The original sorptivity tube was made out of glass. Alcohol can be put in the glass supply tower to determine if a soil is repellent. *From Clothier, B.E., White, I., ©1988, Soil Science Society of America. Reprinted by permission of the Soil Science Society of America.*

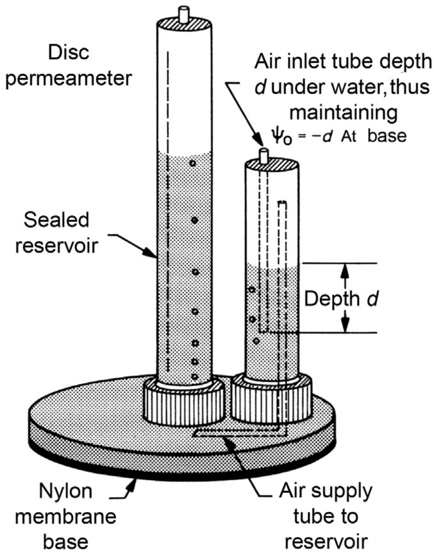

FIGURE 13.4 The disc permeameter. *From a talk given by Brent E. Clothier at the International Soil Science Congress, Acapulco, Mexico, July, 1994. The paper from the Congress is published as Clothier et al., 1994; but the figure is not published in the paper. Reprinted by permission of Brent E. Clothier.*

water can enter is given by the capillary rise equation (also given in Chapter 6 as Eqn 6.9):

$$h_c = (2\sigma \cos \alpha)/(r\rho g), \tag{13.9}$$

where σ = surface tension (surface tension coefficient) of the liquid (units of grams per square second or dynes per centimeter); α = contact angle between the liquid and tube; r = radius of tube (centimeter); ρ = density of liquid (grams per cubic centimeter); and g = acceleration due to gravity (centimeters per square second).

This maximum diameter is proportional to the matric potential, $(-\Psi_o)^{-1}$. The more negative the Ψ_o, the smaller the maximum diameter of a pore that can participate in flow from the soil surface. These two instruments (tension infiltrometer and disc permeameter) are being used to supply water to soil in situ at readily selectable zero or negative pressures. A "ready reckoner" of the relationship between the negative pressure Ψ, where Ψ is in terms of energy per unit weight (in centimeters of H_2O head), and the capillary diameter d in millimeters is $-3/d$ (Clothier, 2008). For example, a 4-cm head will fill pores up to 0.75 mm.

In Figure 13.4 we see that a disc permeameter consists of two towers: one that is open to the air and one that is sealed off from the air. The one that is sealed off from the air supplies water to the soil under tension. The amount of tension is determined by the depth, *d*, that the air-inlet tube is below the water surface. The two towers are connected by an air-supply tube to the reservoir at the bottom. But note that the bubble tower (with air-inlet tube) is not connected to the reservoir. Hence, one needs only to replenish the reservoir with a tracer solution such as KBr, when determining mobility (see Section 13.9). The nylon membrane base determines the tension that can be held, and it is similar to the porous cup on a tensiometer. The amount of tension that a tensiometer can hold is determined by the size of the pores in the ceramic cup. The smaller the pores, the more tension that can be held. The nylon membrane used in the type of disc permeameter shown in Figure 13.4 is Nybolt Nylon Monofil Mesh, Reference PA40/23, 40 μm opening and 23% free surface, purchased from Ure Pacific, Auckland, New Zealand (Kirkham and Clothier, 2000).

The maximum tension that a tension infiltrometer can hold is about −150 mm, but on a good day it can hold −200 mm (B.E. Clothier, personal communication, October 18, 1996). Although this might not seem like much compared to 1 bar (−1020 cm water), when we consider the height of water being held (15−20 cm), this is a significant amount of water.

Remember that we are infiltrating water into soil under tension, but the pores are filled with water. The water in the pores is under tension. Whether the water under the tension infiltrometer is part of the vadose zone depends on the definition of the term. The *Glossary of Soil Science Terms* (Soil Science Society of America, 2008) defines *vadose zone* as "The aerated region of soil above the permanent water table". Sometimes it is defined as the "unsaturated" zone. The *Glossary*'s definition would not include the water under a tension infiltrometer. The water under a tension infiltrometer is similar to water in the falling-water-table situation, when the water table has fallen below the soil surface, and the capillary tubes (soil pores) hold water under tension (see Figure 6.9). When the water table is at the soil surface, the water is not under tension. Under these conditions, or saturated conditions with no tension, Darcy's law applies.

When using the tension infiltrometer, vegetation is scraped away from the soil and a thin layer of contact sand is put between the tension infiltrometer and the soil surface, as shown in Figure 13.3. When set up this way, tension-disc infiltrometry is a practical means for hydraulic characterization of soil close to saturation (Angulo-Jaramillo et al., 1997). The tension infiltrometer is used to determine the unsaturated hydraulic conductivity and sorptivity; the repellency of soils; and the mobility of chemicals through the soil, which we shall discuss in Sections 13.7−13.9, respectively.

The tension infiltrometer is sold in the United States by Soil Measurements Systems (SMS), Tucson, Arizona, with a system patented by Iowa State University (Ankeny et al., 1989). SMS pays a royalty to Iowa State University for sale of the equipment. SMS makes two models, one with an 8-cm-diameter base plate and one with a 20-cm base plate. The reason the equipment could be patented was because Ankeny et al. (1989) added an adjustment to the bubble tower to provide variable tension. The Clothier and White (1981) and Perroux and White (1988) models (Figures 13.3 and 13.4; not patented) have only one air-inlet tube and provide only one tension, but a tension that can be adjusted by moving the air-inlet tube up and down in the water reservoir. The Ankeny et al. (1989) model has three air-inlet tubes. Two are clamped off while the third tension is used.

13.6 MINIDISK INFILTROMETER

The minidisk infiltrometer is made by Decagon Devices (Pullman, WA, USA). The original minidisk infiltrometer, first sold in 1997, consists of a plastic tube, 22.5 cm long and 3.1 cm in outside diameter, marked with milliliter gradation (0–100 ml), a rubber stopper placed in the top, and a styrofoam-looking base that holds the tension. One-half centimeter above the base is an air-inlet tube. It infiltrates water at a set suction (tension) of 2.0 cm and has a radius of 1.59 cm. Decagon Devices developed two more minidisk infiltrometers at set suctions of 0.5 and 6.0 cm (each with a radius of 1.59 cm). The minidisk infiltrometer is especially suited for greenhouse work with pots, even though it can also be used in the field. The hydraulic conductivity of soil can be measured with it using the method of Zhang (1997).

What the base is made out of is proprietary information, but it corresponds to the nylon membrane at the bottom of the tension infiltrometer. The base of the minidisk infiltrometer can get clogged when working on organic soils, such as occurred in the experiments of Kirkham and Clothier (2000), but the disk can be modified by puncturing the bottom with small needle holes and replacing the tension it held with the nylon used in the tension infiltrometer designed by Clothier and White (1981). Kirkham and Clothier (2000) did this modification to maintain functionality of the minidisk infiltrometer. The original base was not removed because it was necessary to have a base to place on the soil during infiltration.

The tension of 2 cm is established by the length of the plastic tube inserted near the bottom of the infiltrometer. There are two forces that balance out to establish the 2 cm. First, there is a gravitational head that must be offset. This is the height of the air entry point from the base of the minidisk. This head is offset by the capillarity of the thin plastic tube, plus the hydraulic resistance of the length of the tube. Personnel at Decagon Devices have worked out this length, so that the tension is 2 cm (B.E. Clothier, personal communication, March 28, 2001).

Some people have misunderstood the minidisk infiltrometer and think that water in it is under pressure. Water is not under pressure, but has to be under tension; if it were under pressure, the water would jump out of the tube. To get water into the infiltrometer before a run, one places the minidisk infiltrometer in a large bucket of water and plugs the top with the rubber stopper. The water stays in the minidisk infiltrometer until it is placed on soil, when it is then sucked out of the tube.

The minidisk infiltrometer has the advantage of being portable. It can be carried to any soil on the globe in a purse or carry-on suitcase, where it can be used to determine the hydraulic conductivity of the soil. No checking of baggage is needed, as for the larger versions of the tension infiltrometer.

13.7 MEASUREMENT OF UNSATURATED HYDRAULIC CONDUCTIVITY AND SORPTIVITY WITH THE TENSION INFILTROMETER

Even though the minidisk infiltrometer can be used to get hydraulic conductivity, the tension infiltrometer (Figure 13.4) is more widely used. Two methods can be used to get unsaturated hydraulic conductivity with the tension infiltrometer: the method of Smettem and Clothier (1989) and the method of Ankeny et al. (1991). In the Smettem and Clothier method, two tension infiltrometers with different radii are used to get both unsaturated hydraulic conductivity and sorptivity. In their method, two equations with two unknowns (unsaturated hydraulic conductivity and sorptivity) are solved simultaneously. The Ankeny et al. (1991) method requires the use of the equation developed by Gardner (1958) to get the relationship between the hydraulic conductivity and water content. In both methods, three-dimensional infiltration is assumed and the Wooding equation is used. For the Wooding equation to be used, infiltration must be at steady state. Thus, before either method is used, one must make sure that the water is infiltrating into the soil at steady state (i.e., the infiltration rate, millimeters per second, is a constant value). Let us first look at sorptivity, because we will be determining it along with hydraulic conductivity when we consider the method of Smettem and Clothier (1989).

The concept of sorptivity comes from the theoretical work that Philip (1969) did on infiltration. The sorptivity of a soil is a measure of the ability of the soil to attract water by capillary action. When considering sorptivity, think of the soil as a sponge. Each soil (sponge) has its ability to absorb water. The units of sorptivity are length divided by square root of time (e.g., millimeters second$^{-1/2}$). Infiltration into a soil is affected by gravity and capillarity. Equation (13.5) gives Philip's equation for vertical infiltration and Eqn (13.6) gives Philip's equation for gravity-free infiltration. In both equations, infiltration goes as the square root of time and from Eqn (13.6) we can see that the units of sorptivity are length divided by the square root of time.

Here is the theoretical development of Smettem and Clothier (1989) based on theory developed by Philip (1969):

$$S_o^2 = [(\theta_o - \theta_n)/b] \int_{\theta_n}^{\theta_o} D\mathrm{d}\theta, \tag{13.10}$$

where

S_o = sorptivity
θ_n = initial soil water content
θ_o = water content to which the soil surface is wetted
D = diffusivity
$D = k \cdot (\partial\varphi/\partial\theta)$
k = hydraulic conductivity under unsaturated conditions

b is a parameter and is a function of the slope of the diffusivity function; $\frac{1}{2} < b < \pi/4$; b can be approximated as 0.55.

Combining Wooding's (1968) equation (Eqn (13.8)) and Philip's theory (1969), Smettem and Clothier (1989) obtained the following equation:

$$q_\infty/\left(\pi r^2\right) = K_o + \left(2.2 S_o^2\right)/(\pi r \Delta\theta), \tag{13.11}$$

where q_∞ = flow at long times (steady conditions); this is the same as i_f in Eqn 11.8.

$$\Delta\theta = \theta(\Psi_o) - \theta(\Psi_n). \tag{13.12}$$

Ψ_o is the supply potential (i.e., the tension with which the water is applied to the soil). Smettem and Clothier (1989) used a supply potential of −35 mm when they did their experiment. Ψ_n is the potential of the soil before the tension infiltrometer is put on top of it. For practical purposes, $\Delta\theta = \theta_o - \theta_n$.

The theoretical development is difficult, but the resulting equation is simple to use for the experimentor. Two tension infiltrometers with different radii are taken to the field where one wants to determine unsaturated hydraulic conductivity and sorptivity. The antecedent volumetric water content is determined (θ_n) and the tension infiltrometers are set close by. After steady infiltration has been reached for each, one records the value. One then takes the tension infiltrometers off the soil and gets the volumetric water content of soil right under the tension infiltrometers. This is θ_o. One then determines $\Delta\theta$ and solves Eqn (13.11) for K_o and S_o by simultaneous solution using data from the discs with two different radii. We have the known parameters, $\Delta\theta$, r, and q_∞, and we have the two unknowns, S_o and K_o.

The method of Smettem and Clothier (1989) is simple and results in two values (K_o and S_o). However, many people cannot obtain or afford two tension infiltrometers with different radii; hence the method of Ankeny et al. (1991) provides a means of obtaining unsaturated hydraulic conductivity with just one tension infiltrometer.

The determination of unsaturated hydraulic conductivity using the method of Ankeny et al. (1991) is not straightforward or obvious from reading the paper. On November 7, 1995, Dr Brent E. Clothier visited Kansas State University and demonstrated to students how to determine unsaturated hydraulic conductivity with this method. We now go through the procedure, step by step, as he explained it.

We need to know the hydraulic conductivity as a function of head, h. We use the following equation (rewritten from Eqn 11.7 where K_s is used instead of K_o used by Jury et al. (1991)):

$$K(h) = K_s \exp(\alpha h), \tag{13.13}$$

where K_s and α are constants. In Eqn (13.13), K_s represents gravity and α represents capillarity; α also is a slope (the slope of the hydraulic conductivity versus head). We can relate K in Eqn (13.13) to the K in a form of Darcy's law, as follows:

$$J_w = K(\partial\Psi)/(\Psi z), \tag{13.14}$$

where J_w is the soil water flux.

The soil that Clothier used in his demonstration was a Haynie sandy loam. It has 65% sand, 24% silt, and 11% clay.

In the Ankeny et al. (1991) method, we use two heads. In this experiment, the two heads will be as follows:

$$h_1 = -10 \text{ cm}$$

$$h_2 = -2 \text{ cm}$$

We will have two unknowns and two equations and will solve the equations to get the unsaturated hydraulic conductivity. We need to run the tension infiltrometer two times; therefore, we shall put it two times on the soil. Water flows into the soil by gravity and capillarity.

In his analysis, Clothier used the following form of Wooding's (1968) equation:

$$Q = \pi r^2 K_o + (4rK_o)/\alpha, \tag{13.15}$$

where Q (cubic meters per second) is the flow through a disc, which is proportional to the surface area of the disc times the hydraulic conductivity plus capillarity (movement of water that goes off the perimeter).

The first term on the right-hand side of Eqn (13.15) represents the gravity component, and the second term on the right-hand side of Eqn (13.15) represents the capillary component. We can measure Q, how fast the water in the sealed reservoir drops in the tension infiltrometer.

From Eqn (13.15), we write

$$q = K_o[1 + 4/\pi\alpha r]. \tag{13.16}$$

We shall apply this equation at one head and then at another head.

$$K = K_1 \exp(\alpha h_1) = K_2 \exp(\alpha h_2) \tag{13.17}$$

$$K_1/K_2 = \exp[\alpha(h_1 - h_2)] \tag{13.18}$$

$$\alpha = [\ln(K_1/K_2)]/(h_1 - h_2) = [\ln(q_1/q_2)]/(h_1 - h_2). \tag{13.19}$$

We are taking the functional form, $K = K_s \exp(\alpha h)$. We have a fixed radius and α is constant.

$$(K_1/K_2) = q_1/q_2. \tag{13.20}$$

So we shall measure two q's (q_1 and q_2).

If we have a clay soil, capillarity dominates and α (slope of the hydraulic conductivity versus head) is small. If we have a sandy soil, α is large. For our first head, $h_1 = -10$ cm (the air tube is 10 cm under water). We apply the tension infiltrometer to the soil. We wait until steady state flow.

We need to know the ratio, R, of the areas of the reservoir to the disc (the area that touches the soil) for the tension infiltrometer. The nylon base of the tension infiltrometer had a diameter of 6.57 cm. The reservoir had a diameter of 3.37 cm.

$$(3.37/6.57)^2 = R = 0.263. \tag{13.21}$$

We time the drop in the sealed reservoir and get the steady state rate. We multiply this rate by 0.263 to get q_1. We find the following:

$$q_1 = 0.115 \text{ cm/s}. \tag{13.22}$$

We have done the -10 cm head.

Now we set the tension infiltrometer at the -2 cm head. We find the following:

$$q_2 = 0.175 \text{ cm/s}. \tag{13.23}$$

From Eqn (13.19), we have

$$\alpha = [(\ln(q_1/q_2)]/(h_1 - h_2)$$

$$\alpha = [\ln(0.115/0.175)]/[-10 \text{ cm} - (-2 \text{ cm})] = 0.05/\text{cm} \tag{13.24}$$

$$K = \exp(0.05h). \tag{13.25}$$

We use Eqn (13.16), Wooding's equation, to calculate K_o:

$$q = 0.175 \text{ cm/s}$$

$$\alpha = 0.05/\text{cm}$$

$$r = 6.57 \text{ cm}/2 = 3.285 \text{ cm}.$$

We solve for K_o:

$$K_o = 0.02 \text{ cm/s} \tag{13.26}$$

at 0.175 cm/s (q_2 for -2 cm head).

$$K = 0.02 \exp(0.05h) \tag{13.27}$$

Equation 13.27 is for $K(h)$ for $h > 2$ cm.

13.8 MEASUREMENT OF REPELLENCY WITH THE TENSION INFILTROMETER

Although some sands and peats are observed to become water repellent when dry, most soils show no obvious reluctance to wet, so it is assumed that they absorb water and ethanol freely (Scotter et al., 1989). Given the same effective contact angle, sorptivity of a liquid into a porous material should be proportional to $(\sigma/\mu)^{½}$ where σ and μ are the surface tension and viscosity of the liquid, respectively (Philip, 1969; see his Eqn 45 on p. 238). Thus, dry soil should imbibe water about twice as fast as alcohol. Organic coatings on peds or particles in some soils can induce some water repellence, but do not affect their ability to absorb alcohol. So in water-repellent dry soil the sorptivity of alcohol will be greater than water. Scotter et al. (1989) checked this observation out in the field, where they took a sorptivity tube (Figure 13.3) to measure the sorptivity of ethanol and water by a fine sandy loam, a common agricultural soil in the region of Palmerston North, New Zealand. They did as much as they could to reduce the likelihood of water repellence. They removed the top 50 mm of soil because organic coatings are most likely to be at the surface. They did the experiments in November in the first drying cycle following an unusually wet winter, when the initial water content of the soil was $0.24 \text{ m}^3/\text{m}^3$. The soil appeared to absorb water normally, proportional to the square root of time and with a uniform wet front. But to their surprise the ethanol was absorbed an order of magnitude faster than the water, indicating significant water repellence.

On the same day of the first experiment, Scotter et al. (1989) took some of the soil back to the laboratory, sieved it, packed it into a tube and again measured the sorptivity of the two liquids. This time the water went in faster than the ethanol, presumably due to the abrasion during sieving.

Other experiments showed that after a few days sieved soil again became water repellent, apparently due to a reorientation of the organic coatings on the soil particles or aggregates.

Further work on repellent soils of New Zealand (Clothier et al., 2000) confirmed the earlier work (Scotter et al., 1989). Clothier et al. (2000) showed that a Ramiha silt loam, another agricultural soil in the region of Palmerston North, was ephemerally hydrophobic. Temporal changes in the measured infiltration rate changed as the repellency broke down. Clothier et al. (2000) cautioned that repellency might not be evident in a soil if infiltration is not observed over a long period. It may take time for the repellency to break down before the infiltration rate climbs to a rate characteristic of the nonrepellent soil.

Peat soils are notoriously repellent. They are common in The Netherlands and much research has been done in that country to study repellency (Ritsema, 1998; Dekker, 1998). When the polders in The Netherlands were drained, the peat soils were allowed to get too dry. They never could be wetted up because of their extreme repellency. So the engineers learned that the peat soils had to be kept wet, if farmers were going to be able to plant seeds and make the soils arable (Don Kirkham, personal communication, undated). In Turkey highly organic soils have been seen to catch on fire (Don Kirkham, personal communication, undated). Forest soils are also repellent because they are organic (Kirkham and Clothier, 2000). John Letey at the University of California at Riverside, in the heart of orange-grove country before population expansion, is well known for his studies of the repellency of soils (Letey et al., 1975). Organic drippings from citrus groves make orchard soils repellent. Soils with organic materials, such as no-till soils with residues, often do not wet readily (Woche et al., 2005). For a further discussion of water repellency and methods to determine it, see Bachmann et al. (2003).

One cannot use alcohol in the commercially available tension infiltrometers from Soil Measurement Systems (Tucson, AZ, USA) and Decagon Devices, because the towers are made of plastic and the alcohol would corrode the plastic. One needs a tension infiltrometer made out of glass, as was the original sorptivity tube (Figure 13.3) to carry out experiments in which one infiltrates alcohol into the soil to determine repellency.

13.9 MEASUREMENT OF MOBILITY WITH THE TENSION INFILTROMETER

Many water flow processes of interest such as groundwater recharge are concerned only with area-averaged water input. Therefore, preferential flow of water through structural voids does not necessarily invalidate equations that assume homogeneous flow, like Darcy's law. However, preferential flow is of critical importance in solute transport, because it enhances chemical mobility and can increase pollution hazards. Many times we need to monitor chemical mobility along with hydraulic properties. We now shall see how we can determine mobility of chemicals.

Water and nutrients not taken up by roots move to depth and eventually to groundwater with deleterious consequences. With the tension infiltrometer (disc permeameter), we can measure hydraulic properties of soil that control infiltration and retention. In particular, we can distinguish mobile and immobile water with it.

The soil water content, θ, is made up of the mobile-water content, θ_m, and the immobile-water content, θ_{im}. Figure 13.5 shows a mobile-water soil, and Figure 13.6 shows a mobile–immobile water soil. Water is immobilized due to several factors: it can be in an occluded pore, it can be bound water, it can be in a dead-end pore, or it can be in the soil's microporosity and unable to move. θ_m is active in chemical transport,

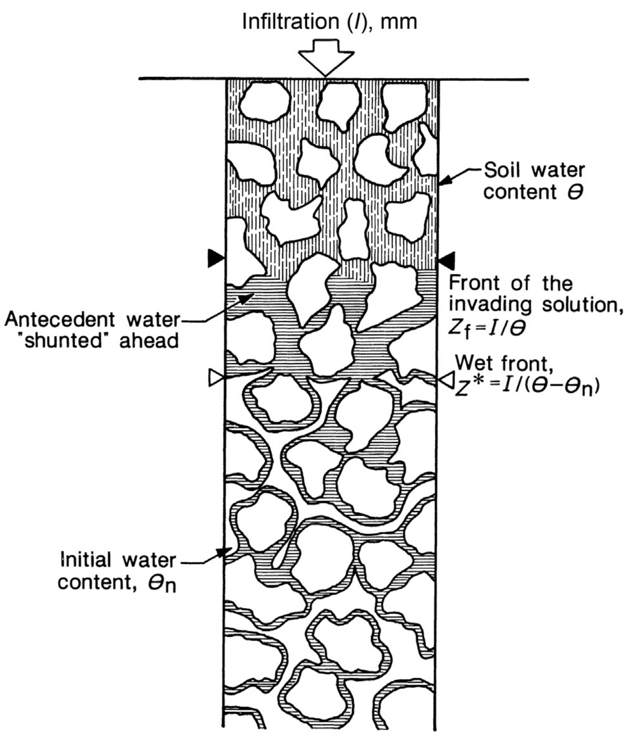

FIGURE 13.5 A "mobile-water" soil. *Redrawn from a slide by Brent E. Clothier. Reprinted by permission of Brent E. Clothier.*

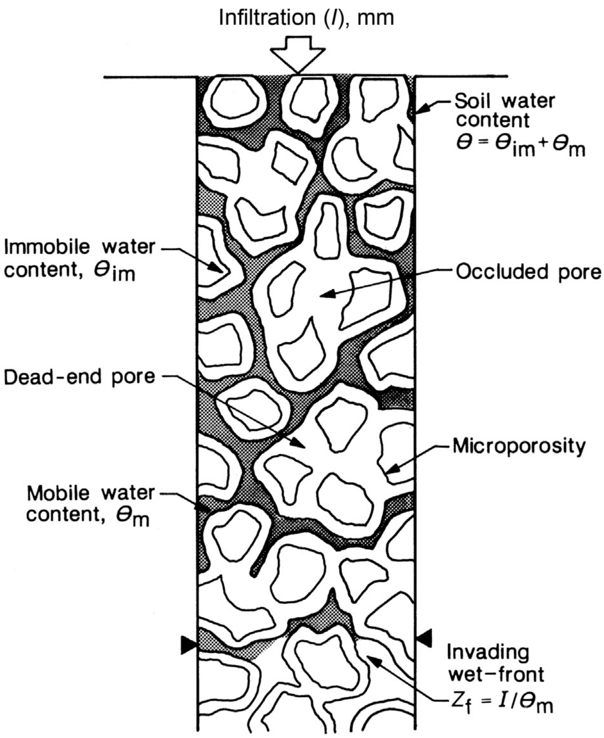

FIGURE 13.6 A "mobile–immobile" water soil. *Redrawn from a slide by Brent E. Clothier. Reprinted by permission of Brent E. Clothier.*

while θ_{im} is not. If we know θ_m, we can develop management strategies to minimize leaching losses of chemicals.

The method of using the disc permeameter to get θ_m and θ_{im} relies on using it to supply a tracer solution to soil (Clothier et al., 1992). The tracer is not originally in the soil. If a tracer is added to a disc permeameter at a concentration c_m, then from the observed solution concentration c^* in soil samples extracted from underneath the disc, θ_m can be calculated from the dilution by the water of the immobile phase that must have remained in place during the passage past it of the invading solution of tracer. So we have

$$c^*\theta = c_m\theta_m + c_{im}\theta_{im}. \tag{13.28}$$

But if $c_{im} = 0$ (there was no tracer in the soil to begin with), then the last term on the right-hand side of Eqn (13.28) drops out and we have:

$$\theta_m = \theta(c^*/c_m). \tag{13.29}$$

In other words, the fraction, θ_m/θ, is directly proportional to the relative concentration of solute found in the soil, c^*, to that applied, c_m:

$$c^*/c_m = \theta_m/\theta. \tag{13.30}$$

In the experiments of Clothier et al. (1992), the tracer was bromide. (Chloride could not be used, because of the proximity of the Pacific Ocean and salt in the air that lands on the soil.) They applied water to the soil with a tension infiltrometer with $h_o = \Psi_o = -20$ mm. (See Figure 13.4.) The air-inlet tube was 20 mm under the surface of the water in the tower on the right-hand side of the figure. The soil was near saturation.

In the first experiment, pure water was used to first wet the soil. Then a tracer at concentration 0.1 mol/L KBr or 0.1 M KBr was drawn into this already wet soil predominantly by gravity. The original water content of the soil, a Manawatu fine sandy loam, was 0.414 m^3/m^3. Three runs were done. Bromide in the soil was determined after each run. The method is shown in Figure 13.7. The ratio c^*/c_m was determined for the three runs, and they were 0.46, 0.48, and 0.50, with an average of 0.49. Therefore,

$$\theta_m/0.414 = 0.49\ m^3/m^3 \tag{13.31}$$

$$\theta_m = 0.203\ m^3/m^3 \tag{13.32}$$

so only half of the soil water was mobile (0.414 m^3/m^3 vs 0.203 m^3/m^3).

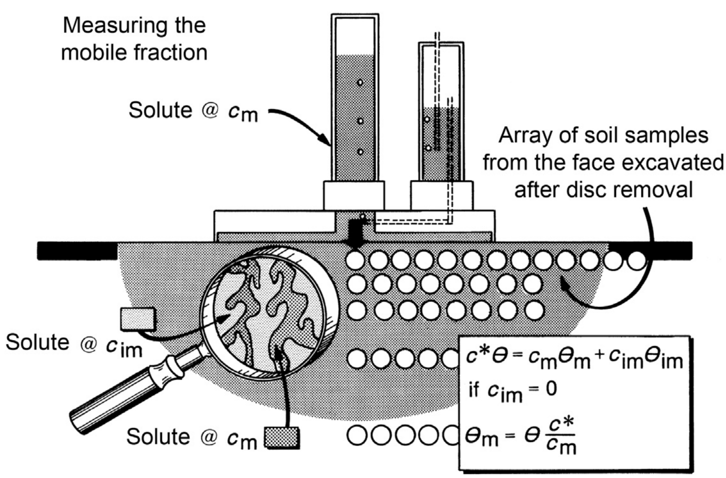

FIGURE 13.7 Measuring the mobile fraction. *From Clothier, B.E., Green, S.R., Magesan, G.N., ©1994. Soil and plant factors that determine efficient use of irrigation water and act to minimise leaching losses. Trans. Int. Congr. Soil Sci. 2a; 41–47. Reprinted by permission of The International Society of Soil Science.*

We can determine the depth of penetration of the solute using the following equation:

$$Z_s^* = I/\theta_m \tag{13.33}$$

where

Z_s^* = depth of the solute front (the asterisk, *, after the s for solute indicates that we are dealing with a tracer)
I = cumulative infiltration
θ_m = mobile-water content.

For the three runs, average I was 15.7 mm. So $Z_s^* = 15.7\ \text{mm}/(0.203\ \text{m}^3/\text{m}^3) = 77.3\ \text{mm}$.

If all the water were mobile, then $\theta_m = \theta$.

$$15.7\ \text{mm}/\left(0.414\ \text{m}^3/\text{m}^3\right) = 37.9\ \text{mm}.$$

The observed front of the bromide, determined experimentally at the end of the experiment, showed that it was at about 77 mm, not at about 38 mm. So the experimental data backed up the calculation. Because not all the water was mobile, the tracer penetrated to deeper depths than it would have, had $\theta_m = \theta$.

The longitudinal mobility (the depth of penetration) is inversely related to θ_m. This is sometimes hard for people to understand when they are using the method initially, because they think that the more mobile the water the deeper will be the penetration of a solute. This is not so. The larger the mobile volume fraction, (i.e., θ_m) the less longitudinally mobile is the solute (the smaller Z_s^*). In other words, the solution carrying the dissolved solute travels through a larger volume fraction of the soil's wetted pore space, such that for a given amount of water infiltrated, I, the less the penetration into the soil. As θ_m becomes a smaller fraction of the wetted θ (total water content), the volume fraction of the soil's wetted pore space that transports the invading solution becomes smaller. Hence, for an equivalent amount of I, the greater is the depth, due to the fact that Z_s^* is inversely related to θ_m.

In another experiment (Clothier et al., 1992), the tracer was put on a dry soil. The tension infiltrometer was still set at −20 mm, but there was no prewetting. The tracer solution was now drawn into the soil more by capillarity than by gravity. The mobile fraction measured right under the disc rose to $\theta_m = 0.291\ \text{m}^3/\text{m}^3$. (Remember that the previous value, when a wet soil was used, $\theta_m = 0.203\ \text{m}^3/\text{m}^3$.) This rise is attributed to the direct, capillary-induced movement of the invading solute into more of the soil's microporosity. The solute invasion depth was less (shallower) than half of the former case (prewetted, $\Psi_o = -20$ mm). The prewetting had destroyed the capillary attractiveness of the micropores. The "dry-soil" effect keeps fertilizers near the soil surface. One can think of bromide as a tracer for nitrate. Thus effluent and other nutrient-laden fertilizer waters should be applied to dry soil. This applied chemical can be drawn by capillarity into more of the soil's microporosity, close to the soil surface (where roots are) and not escape to depth and groundwater. Once in the soil's microporosity, the applied nutrients are rendered less likely to be leached by subsequent rainfall.

13.10 ELLIPSOIDAL DESCRIPTION OF WATER FLOW INTO SOIL FROM A SURFACE DISC

The disc permeameter is being widely used to characterize the hydraulic properties of the surface of the soil. It is important to describe the multidimensional flow of water away from the circular source. Little work has been done to analyze precisely the flow of water away from a circular source of water applied at a constant negative potential, Ψ_o. The Wooding equation (1968; his Eqn 64) does describe the steady rate of three-dimensional infiltration from a circular pond. His equation applies to profiles at infinite time and does not give information about the shape of wet fronts of transient wetting. Kirkham and Clothier (1994a,b) used a unique approach to analyze the flow pattern, because they assumed that the three-dimensional wetting fronts under the circular source were ellipsoidal.

They described mathematically the three-dimensional flow of water away from a disc source placed upon the soil's surface at constant negative potential, Ψ_o less than 0, when the water is being applied to the soil at a steady rate, q in cubic millimeters per second. The wet fronts analyzed are shown in Figure 13.1. The wet fronts resulted from water being infiltrated from a quarter-disc permeameter set in the corner of a plastic box with soil. The setup is shown as Figure 1 in Kirkham and Clothier (1994b). It allowed markings of the wet fronts as they penetrated the soil under the quarter-disc permeameter set at -50 mm supply potential (Ψ_o). The reason for using this suction was to allow the soil to be unsaturated but not too far away from saturation, such that the flow would be unrealistically slow.

The mathematical development is given by Kirkham and Clothier (1994a,b). Assuming the volume wetted in each wetted area (delineated in Figure 13.1) is an ellipsoid, the following equation can be used for the volume of water, $V(t)$, that infiltrated from the disc:

$$V(t) = (2/3\pi)(\theta_m - \theta_n)R^2(t)Z(t), \tag{13.34}$$

where the extent of radial wetting at the surface is R at any time t, and vertically under the disc is Z. The initial volumetric water content of the soil is θ_n and the disc wets the soil surface to water content θ_o, a function of

Ψ_o. A weighted-average water content can be ascribed to the wetted field, designated θ_m.

It can be seen from the wet fronts (Figure 13.1) that the ellipsoid describing the spatial pattern of wetting goes from being an oblate (egg on its side; see Figure 7.3) at early times, through to being spheroidal by about the end of the experiment. It would eventually become prolate (egg on its end; see Figure 7.4), with further extension being limited to the vertical.

As noted above, profiles predicted from Wooding's (1968) equation hold only at infinite time, and thus, they cannot give any information about transient wet fronts. Where the ellipsoidal idea has merit over Wooding's equation is in the practical operation of the disc permeameter. The ellipsoidal equation can give answers to the following questions: How deep is the wet front at any time? By what means can it be reckoned simply? Also, we can answer other questions relating to the modus operandi of the disc permeameter. How much of the soil's volume has been wetted during a disc experiment? What volume of soil has been sampled, and, hence, to what depth of soil do the Wooding K and S values apply?

The disc permeameter might be practically applied to different tillage situations. If the soil will not suck in water, as determined by the disc permeameter, then we need to till the soil. If there is a crust on the soil surface or the soil surface is repellent, water cannot infiltrate. Water runs off and moves preferentially through the macropores instead of through the soil matrix. Tillage of the soil is critical so that rainwater will not sit on the surface and evaporate (Figure 13.8). We want water to go into the soil,

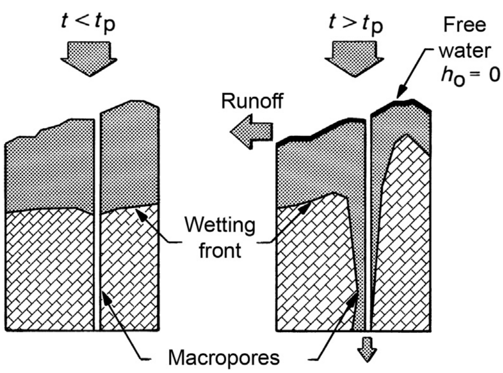

FIGURE 13.8 Infiltration of an applied flux of water into soil. Left: Nonponding infiltration when the flux through the surface is less than the saturated hydraulic conductivity, K_s, or ponding infiltration when the flux through the surface is greater than K_s prior to the time of ponding t_p. Right: Pattern of infiltration after incipient ponding, $t > t_p$, when the possibility of runoff exists, as does the entry of free water into macropores. *From Clothier, B.E., Infiltration. In: Soil and Environmental Analysis. Physical Methods, second ed., Marcel Dekker, Inc., New York, ©2001. This material is reprinted with permission of Marcel Dekker, Inc.*

where it will be used by plants. The disc permeameter also can be practically applied to different irrigation systems, including drip and furrow irrigation. Perhaps one could put the disc permeameter, set at a constant negative potential, on a dry soil and see how far water is sucked out to the side and calculate the volume wetted, based on the ellipsoidal equation. The distance could be measured with a ruler. Then one could place irrigation sources at distances from each other based on the measurements.

The ellipsoidal equation has another advantage. It is simple to use and people with essentially no mathematical training can apply it. A small hand-held calculator could be programmed easily to make the calculations.

13.11 APPENDIX: BIOGRAPHY OF JOHN PHILIP

John Robert Philip, soil physicist, was born on January 13, 1927, in Ballarat in rural Victoria, Australia (Burges et al., 1999). He acquired a love of learning from his schoolteacher mother, and won a scholarship to the prestigious Scotch College for boys in Melbourne, where his mathematical ability was recognized and developed. He also was encouraged to write poetry, and this remained a lifelong passion, his poems appearing in many literary publications. He graduated from Scotch College at age 13 years, and he spent another 2 years at school before being deemed old enough (at 16 years) to study civil engineering at the University of Melbourne. Bored by the undemanding engineering courses, which he described as merely “learning which handbook to look up”, he spent much of his time reading and writing poetry. He earned his Bachelor’s of Civil Engineering degree at age 19 years, the youngest ever engineering graduate.

He was offered a research assistantship by the University of Melbourne and was sent to the Council for Industrial and Scientific Research (renamed CSIRO in 1949) station at Griffith to work on problems of furrow irrigation. His fellowship ended after a year and he left to work as an engineer in Queensland. In 1951, he was asked to join the CSIRO Division of Plant Industry at Deniliquin, in New South Wales. Otto Frankel became Philip’s boss. Frankel, a distinguished plant geneticist, was charged with revitalizing the division. He consulted with John Jaeger (see the Appendix of Chapter 9 for a biography of Jaeger) and Pat Moran, two famous mathematicians at the Australian National University, who reported positively on Philip’s proposed research plans in agricultural physics. Frankel gave Philip freedom to proceed (Burges et al., 1999), and Philip praised Frankel for allowing the environment in which he could thrive (personal communication, December 9, 1997). Philip’s papers at Deniliquin dealt with the analysis of environmental water and heat flow, which earned him a D.Sc. in physics from the University of Melbourne. Modern theory is based on his analysis (American Society of Agronomy, 1999).

In 1964, he moved from Deniliquin to Canberra to head the Agricultural Physics section of the Division of Plant Industry (headed by Frankel) (Burges et al., 1999). A bequest to CSIRO provided the funds to build a laboratory to house the team of researchers Philip assembled to work on fluid mechanics of porous media, micrometeorology, plant physical ecology, and soil physics. Philip helped to design the building, called the F.C. Pye Laboratory, built in 1966. The productivity of the team was aided by the architecture of the Pye Lab, which encouraged collaboration; it was open and such that one easily encountered colleagues in the sunny walkways. In 1970, the Division of Environmental Mechanics was created, and Philip was the chief until his retirement in 1992 (American Society of Agronomy, 1999).

Philip was a strong defender of scientific autonomy. In 1975, he chaired the Science Task Force of the Royal Commission on Australian Government Administration. The report he drafted for this task force argued for governmental science characterized by freedom of action and outlined the environment necessary for effective and creative scientific research (American Society of Agronomy, 1999; Burges et al., 1999). In the article he prepared for the seventy-fifth anniversary issue of *Soil Science* (Philip, 1991), he criticized the lack of freedom that scientists now have.

Philip was the preeminent mathematician who solved difficult unsaturated flow problems. The solutions led to practical benefits, including how to handle infiltrating irrigation water. His work was the basis for numerous theoretical advances in infiltration and soil—plant—water relationships and his vertical infiltration model is used worldwide. He also dominated multidimensional infiltration theory. He published more than 300 papers. While he is best known for his work in soil and porous media physics, fluid mechanics, and hydrology, his papers in the plant physiology literature are classic. He published on osmotic and turgor properties of plant cells. He pioneered the concept of the *s*oil, plant, atmosphere as a thermodynamic *c*ontinuum (SPAC) for water transfer. Everyone now talks about the SPAC, most of the time without acknowledging that Philip was the person who first used the term. (See Chapter 22 for more information about the SPAC.)

Philip received numerous honors, including honorary doctor's degrees from the University of Melbourne, the Agricultural University of Athens, and the University of Guelph in Canada. He was a Fellow of the Australian Academy of Sciences and, in 1974, became a Fellow of the Royal Society of London, the highest scientific honor in the Commonwealth. In 1991, he was elected corresponding member of the All Union (now Russian) Academy of Sciences. In 1995, he received the International

Hydrology prize awarded jointly by UNESCO, the World Meteorological Association, and the International Association for Hydrological Science. He also received much recognition in the United States. He was named Fellow of the Soil Science Society and honorary member of the American Water Resources Association and was the first non-American to receive the Robert E. Horton Medal, the highest award for hydrology from the American Geophysical Union (American Society of Agronomy, 1999). In 1995, he was elected as a foreign associate by the National Academy of Engineering in Washington, D.C., for his pioneering research contributions to soil-water hydrology.

Philip and his wife, Francis, had two sons and a daughter. Philip was struck by a car as he stepped off a bus and was killed on June 26, 1999, in Amsterdam, The Netherlands, where he was visiting the Centre for Mathematics and Information Science (American Society of Agronomy, 1999). The driver was sentenced to 1 month in jail and suspension of his driver's license for a year. A volume honoring Philip was published by the American Geophysical Union (Raats et al., 2002), and a tribute plus complete bibliography appeared in the *Australian Journal of Soil Research* (Smiles, 2001).

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