INFILTRATION MODELS

Infiltration is the process of vertical movement of water into a soil from rainfall, snowmelt, or irrigation. Infiltration of water plays a key role in surface runoff, ground water recharge, evapotranspiration, and transport of chemicals into the subsurface. Models to characterize infiltration for field applications usually employ simplified concepts that predict the infiltration rate assuming surface ponding begins when the surface application rate exceeds the soil infiltration rate. Empirical, physically based, and physical models have all been developed for the infiltration process. A more detailed review of infiltration can be found in Rawls et al. (1993).

Richards Equation, Eq. (9.8), is the physically based infiltration equation used for describing water flows in soils. Philip (1957) solved the equation analytically for the condition of excess water at the surface and given characteristic curves. Their coefficients can be predicted in advance from soil properties and do not have to be fitted to field data. However, the more difficult case where the rainfall rate is less than the infiltration capacity cannot be handled by Philip's equation. Another limitation is that it does not hold valid for extended time periods. Swartzendruber (1987) presented a solution to Richards equation that holds for both small, intermediate and large times.

One of the most interesting and useful approaches to solving the governing equation for infiltration was originally advanced by Green and Ampt (1911). In this method, water is assumed to move into dry soil as a sharp wetting front that separates the wetted and unwetted zones. At the location of the front, the average capillary suction head $\psi = \psi_f$, is used to represent the characteristic curve. The moisture content profile at the moment of surface saturation is shown in Figure 9.6a. The area above the moisture profile is the amount of infiltration up to surface saturation F and is represented by the shaded area of depth L in Figure 9.6a. Thus, $F = (\theta_s - \theta_i)L = M_d L$, where θ_i is the initial moisture content, θ_s is the saturated moisture content, and $M_d = \theta_s - \theta_i$ the initial moisture deficit.

Darcy's law is then applied as an approximation to the saturated conditions between the soil surface and the wetting front, as indicated in Figure 9.6b (Bedient and Huber, 1992).

The volume of infiltration down to the depth L is given by

$$F = L(\theta_s - \theta_i) = LM_d \tag{9.10}$$

Neglecting the depth of ponding at the surface, the original form of the Green-Ampt equation

$$\begin{aligned} f &= K_s[1 - (\theta_s - \theta_i)\psi_f/F] \\ &= K_s[1 - M_d\psi_f/F] \end{aligned} \tag{9.11}$$

Because ψ_f is negative, Eq. (9.11) indicates that the infiltration rate is a value greater than the saturated hydraulic conductivity, as long as there is sufficient water at the surface for infiltration, as sketched in curves C and D of Figure 9.6c. Functionally, the infiltration rate decreases as the cumulative infiltration increases.

There are two extreme cases which should be considered. For large values of moisture content, $\partial\psi/\partial\theta$ is approximately zero and the continuity equation becomes

$$\frac{\partial\theta}{\partial t} = -\frac{\partial K(\theta)}{\partial z} \tag{9.9}$$

Equation (9.9) leads to the kinematic theory of modeling the unsaturated flow, where capillary pressure gradients are neglected. The theory is also applicable if ψ = CONSTANT within the profile. Thus, Darcy's law predicts that flow is downward under a unit gradient. The second extreme case occurs when capillary forces completely dominate gravitational forces, resulting in a nonlinear diffusion equation. This latter form is useful for modeling evaporation processes.

To summarize the properties of the unsaturated zone as compared to the saturated zone, Freeze and Cherry (1979) state that:

For the unsaturated zone (vadose zone):

1. It occurs above the water table and above the capillary fringe.
0. 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.

For the saturated zone:

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 gauge pressure) is greater than zero.
4. The hydraulic head h must be measured with a piezometer.
5. The hydraulic conductivity K is not a function of the pressure head ψ.

Finally, more details on the unsaturated zone can be found in Fetter (1999), Rawls et al. (1993), Charbeneau and Daniel (1993) and Guyman (1994).

The rainfall intensity, i, is often less than the potential infiltration rate given by Eq. (9.11), in which case $f = i$. Let the corresponding volume of infiltration be F_s. With $f = i$, Eq. (9.11) can then be solved for F_s, the volume of infiltration at the time of surface saturation (t_s, the time at which Eq. (9.11) becomes valid),

$$F_s = [(\theta_s - \theta_i)\psi_f]/[1 - i/K_s] = M_d\psi_f/(1 - i/K_s) \tag{9.12}$$

We require $i > K_s$ in Eq. (9.12) and remember that ψ_f is negative. The Green-Ampt infiltration prediction is thus the following:

If i K_s, then $f = i$ (curve A in Figure 9.6c)

If $i > K_s$, then $f = i$ until $F = it_s = F_s$ (Eq. (9.12)

After the surface is saturated, the following is used, $f = K_s[1 - M_d\psi_f/F]$ (Eq. (9.11)) for $i > K_s$ and $f = i$ for i K_s.

The combined process is sketched in curves B and C of Figure 9.6c. As long as the rainfall intensity is greater than the saturated hydraulic conductivity, the infiltration rate asymptotically approaches K_s, as a limiting lower value. Mein and Larson (1973) found excellent agreement when using the Green-Ampt method, numerical solutions of Richard's equation, and experimental soils data. If the rainfall rate starts above K_s, drops below it, and then rises back above it during the infiltration computation, the use of Green-Ampt becomes more complicated, making it necessary to redistribute the moisture in the soil column rather than maintaining the assumption of saturation from the surface down to the wetting front shown in Figure 9.6b. The use of the Green-Ampt procedures for unsteady rainfall sequences is illustrated by Skaggs and Khaleel (1982).

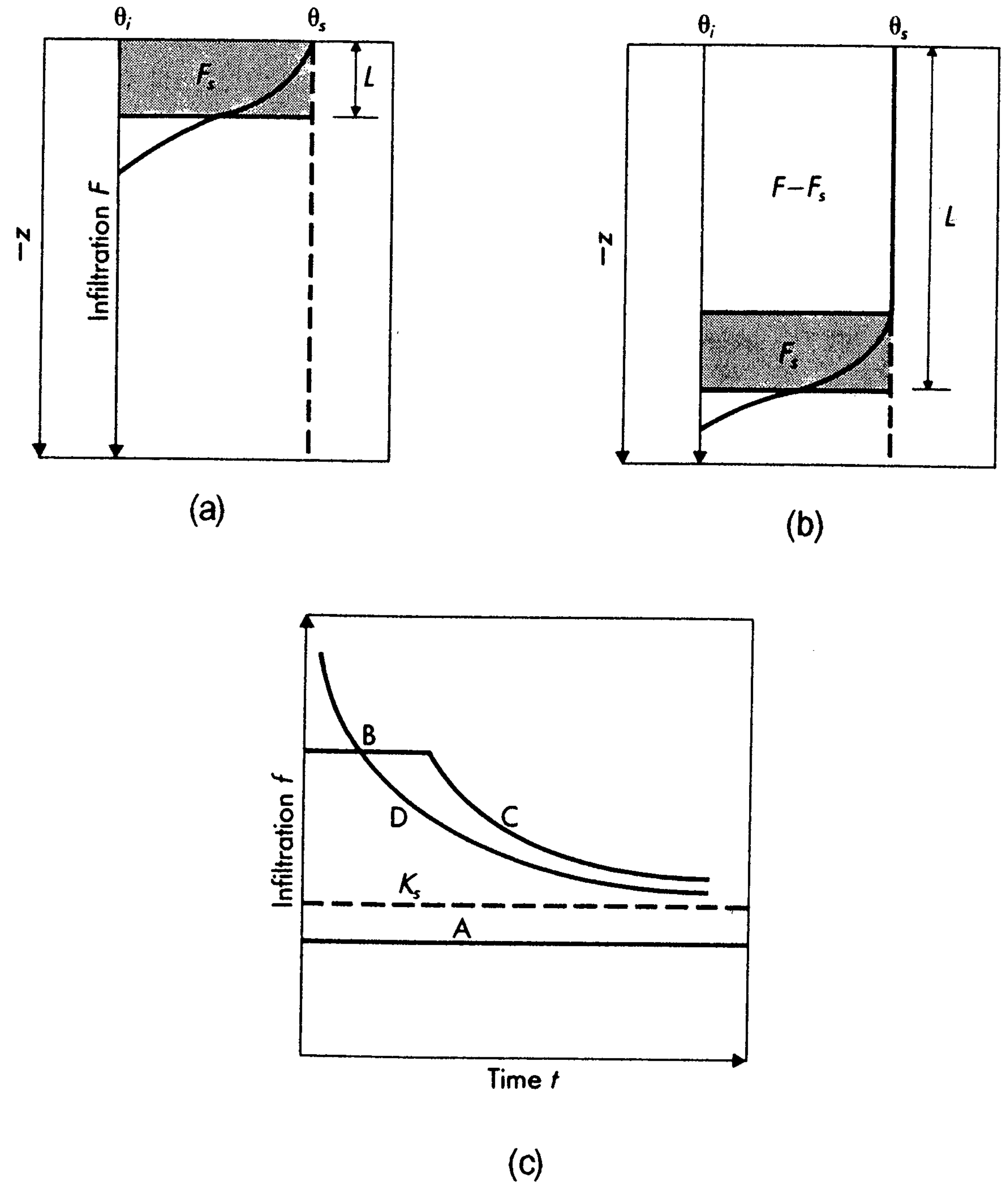

Figure 9.6 Moisture and infiltration relations. (a) Moisture profile at moment of surface saturation. (b) Moisture profile at later time. (c) Infiltration behavior under rainfall. Source: Mein and Larson, 1973.

TABLE 9.1 Green-Ampt infiltration parameters for various soil texture classes

Soil Class	Porosity η	Effective Porosity θ_E	Wetting Front Soil Suction Head ψ (cm)	Hydraulic Conductivity K(cm/hr)	Sample Size
Sand	0.437 (0.374-0.500)	0.417 (0.354-0.480)	4.95 (0.97-25.36)	11.78	762
Loamy sand	0.437 (0.363-0.506)	0.401 (0.329-0.473)	6.13 (1.35-27.94)	2.99	338
Sandy loam	0.453 (0.351-0.555)	0.412 (0.283-0.541)	11.01 (2.67-45.47)	1.09	666
Loam	0.463 (0.375-0.551)	0.434 (0.334-0.534)	8.89 (1.33-59.38)	0.34	383
Silt loam	0.501 (0.420-0.582)	0.486 (0.394-0.578)	16.68 (2.92-95.39)	0.65	1206
Sandy clay loam	0.398 (0.332-0.464)	0.330 (0.235-0.425)	21.85 (4.42-108.0)	0.15	498
Clay loam	0.464 (0.409-0.519)	0.309 (0.279-0.501)	20.88 (4.79-91.10)	0.10	366
Silty clay loam	0.471 (0.418-0.524)	0.432 (0.347-0.517)	27.30 (5.67-131.50)	0.10	689
Sandy clay	0.430 (0.370-0.490)	0.321 (0.207-0.435)	23.90 (4.08-140.2)	0.06	45
Silty clay	0.479 (0.425-0.533)	0.423 (0.334-0.512)	29.22 (6.13-139.4)	0.05	127
Clay	0.475 (0.427-0.523)	0.385 (0.269-0.501)	31.63 (6.39-156.5)	0.03	291

The numbers in parentheses below each parameter are one standard deviation around the parameter value given.
Source: Rawls, Brakensiek, and Miller, 1983.

Example 9.1 GREEN-AMPT TIME TO SURFACE SATURATION
Guelph Loam has the following soil properties (Mein and Larson, 1973) for use in the Green-Ampt equation:

K_s = 3.67×10^{-4} cm/sec
θ_s = 0.523
ψ = –31.4 cm water

For an initial moisture content of $\theta_i = 0.3$, compute the time to surface saturation for the following storm rainfall:

i = $6K_s$ for 10 min
i = $3K_s$ thereafter

Solution. The initial moisture deficit, $M_d = 0.523 - 0.300 = 0.223$. For the first rainfall segment, we compute the volume of infiltration required to produce saturation from Eq. (9.12):

$$Fs = \psi_f M_d / (1 - i/K_s) = (-31.4 \text{ cm})(0.223) / (1 - 6K_s/K_s) = 1.40 \text{ cm}$$

The rainfall volume during the first 10 minutes is

$$10i = (10 \text{ min})\ (6 \cdot 3.67 \times 10^{-4} \text{ cm/sec})(60 \text{ sec/min}) = 1.31 \text{ cm}$$

since 1.31 < 1.40, all rainfall infiltrates and surface saturation is not reached, and $F(10\text{min}) = 1.31$ cm.

The volume required for surface saturation during the lower rainfall rate of i = 3Ks is

$$F_s = (-31.4 \text{ cm})(0.223) / (1 - 3K_s/K_s) = 3.50 \text{ cm}$$

Thus, an incremental volume of $F = F_s - F(10 \text{ min}) = 3.50 - 1.31 = 2.19$ cm must be supplied before surface saturation occurs. This requires an incremental time of

$$t = F/i = (2.19 \text{ cm}) / (3 \cdot 3.67 \times 10^{-4} \text{ cm/sec}) = 1989 \text{ sec}$$
$$= 33.15 \text{ min}$$

Thus, the total time to surface saturation is 10 + 33.15 = 43.15 min.