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Soil and Water

Physical Principles and Processes

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This is a second-order partial differential equation of the elliptical type, and it can be solved in certain cases to obtain a quantitative description of water flow in various systems.

In general, a differential equation can have an infinite number of solutions. To determine the specific solution in any given case, it is necessary to specify the *boundary conditions*, and, in the case of unsteady flow, of the *initial conditions* as well. Various types of boundary conditions can exist (e.g., impervious boundaries, free water surfaces, boundaries of known pressure, or known inflow or outflow rates, etc.), but in each case the flux and pressure head must be continuous throughout the system. In layered soils, the hydraulic conductivity and water content may be discontinuous across interlayer boundaries (that is, they may exhibit abrupt changes). Flow equations for inhomogeneous, anisotropic, and compressible systems were given by Bear *et al.* (1968).

Philip (1969) recently analyzed flow in swelling (compressible) media. In unsteady flow, the solid matrix of a swelling soil undergoes motion, so that Darcy's law applies to water movement relative to the particles, rather than relative to physical space. Experimental work with such soils was carried out by Smiles and Rosenthal (1968).

M. Summary

A proper physical description of water flow in the soil requires that three parameters be specified: flux, hydraulic gradient, and conductivity. Knowledge of any two of these allows the calculation of the third, according to Darcy's law. This law states that the flux equals the product of conductivity by the hydraulic gradient. The hydraulic gradient itself includes both the pressure and the gravitational potential gradients, the first of which is the exclusive cause of flow in a horizontal system, while the second occurs in vertical systems. The hydraulic conductivity at saturation is a characteristic property of a soil toward water flow, and it is related to porosity and pore-size distribution.

5 *Flow of Water in Unsaturated Soil*

A. General

Most of the processes involving soil-water flow in the field, and in the rooting zone of most plant habitats, occur while the soil is in an unsaturated condition. Unsaturated flow processes are in general complicated and difficult to describe quantitatively, since they often entail changes in the state and content of soil water during flow. Such changes involve complex relations among the variable water content (wetness), suction, and conductivity, which may be affected by hysteresis. The formulation and solution of unsaturated flow problems very often require the use of indirect methods of analysis, based on approximations or numerical techniques. For this reason, the development of rigorous theory and methods for treating these problems was rather late in coming. In recent years, however, unsaturated flow has become one of the most important and active topics of research in soil physics, and this research has resulted in significant theoretical and practical advances.

B. Comparison of Unsaturated vs. Saturated Flow

In the previous chapter, we stated that soil-water flow is caused by a driving force resulting from an effective potential gradient, that flow takes place in the direction of decreasing potential, and that the rate of flow (flux) is proportional to the potential gradient and is affected by the geometric properties of the pore channels through which flow takes place. These principles apply in unsaturated, as well as in saturated soils.

The moving force in a saturated soil is the gradient of a positive pressure potential.[1] On the other hand, water in an unsaturated soil is subject to a subatmospheric pressure, or suction, and the gradient of this suction likewise constitutes a moving force. The matric suction is due, as we have pointed out, to the physical affinity of the water to the soil-particle surfaces and capillary pores. Water tends to be drawn from a zone where the hydration envelopes surrounding the particles are thicker, to where they are thinner, and from a zone where the capillary menisci are less curved to where they are more highly curved.[2] In other words, water tends to flow from where suction is low to where it is high. When suction is uniform all along a horizontal column, that column is at equilibrium and there is no moving force. Not so when a suction gradient exists. In that case, water will flow in the pores which remain water-filled at the existing suction, and will creep along the hydration films over the particle surfaces, in a tendency to equilibrate the potential.

The moving force is greatest at the "wetting front" zone of water entry into an originally dry soil (see Fig. 5.2). In this zone, the suction gradient can be many bars per centimeter of soil. Such a gradient constitutes a moving force thousands of times greater than the gravitational force. As we shall see later on, such strong forces are sometimes required (for a given flux) in view of the extremely low hydraulic conductivity which a relatively dry soil may exhibit.

The most important difference between unsaturated and saturated flow is in the hydraulic conductivity. When the soil is saturated, all of the pores are filled and conducting, so that conductivity is maximal. When the soil becomes unsaturated, some of the pores become airfilled and the conductive portion of the soil's cross-sectional area decreases correspondingly. Furthermore, as suction develops, the first pores to empty are the largest ones, which

[1] We shall disregard, for the moment, the gravitational force, which is completely unaffected by the saturation or unsaturation of the soil.

[2] The question of how water-to-air interfaces behave in a conducting porous medium that is unsaturated is imperfectly understood. It is generally assumed, at least implicitly, that these interfaces, or menisci, are anchored rigidly to the solid matrix so that, as far as the flowing water is concerned, air-filled pores are like solid particles. The presence of organic surfactants which adsorb to these surfaces is considered to increase their rigidity or viscosity. Even if the air–water interfaces are not entirely stationary, however, the drag, or momentum transfer, between flowing water and air appears to be very small. The influence of the surface viscosity of air–water interfaces on the rheological behavior of soil water has not been evaluated (Philip, 1970). Preliminary experimental findings by E. E. Miller and D. Hillel suggest that a drag effect does occur, but that its magnitude is negligible for most practical purposes.

are the most conductive,[3] thus leaving water to flow only in the smaller pores. The empty pores must be circumvented, so that, with desaturation, the tortuosity increases. In coarse-textured soils, water sometimes remains almost entirely in capillary wedges at the contact points of the particles, thus forming separate and discontinuous pockets of water. In aggregated soils, too, the large interaggregate spaces which confer high conductivity at saturation become (when emptied) barriers to liquid flow from one aggregate to its neighbors.

For these reasons, the transition from saturation to unsaturation generally entails a steep drop in hydraulic conductivity, which may decrease by several orders of magnitude (sometimes down to 1/100,000 of its value at saturation) as suction increases from zero to one bar. At still higher suctions, or lower water contents, the conductivity may be so low[4] that very steep suction gradients, or very long times, are required for any appreciable flow to occur.

At saturation, the most conductive soils are those in which large and continuous pores constitute most of the overall pore volume, while the least conductive are the soils in which the pore volume consists of numerous micropores. Thus, as is well known, a sandy soil conducts water more rapidly than a clayey soil. However, the very opposite may be true when the soils are unsaturated. In a soil with large pores, these pores quickly empty and become nonconductive as suction develops, thus steeply decreasing the initially high conductivity. In a soil with small pores, on the other hand, many of the pores remain full and conductive even at appreciable suction, so that the hydraulic conductivity does not decrease as steeply and may actually be greater than that of a soil with large pores subjected to the same suction.

Since in the field the soil is unsaturated most of the time, it often happens that flow is more appreciable and persists longer in clayey than in sandy soils. For this reason, the occurrence of a layer of sand in a fine-textured profile, far from enhancing flow, may actually impede unsaturated water movement until water accumulates above the sand and suction decreases sufficiently for water to enter the large pores of the sand. This simple principle is all too often misunderstood.

[3] By Poiseuille's law, the total flow rate of water through a capillary tube is proportional to the fourth power of the radius, while the flow rate per unit cross-sectional area of the tube is proportional to the square of the radius. A 1-mm-radius pore will thus conduct as 10,000 pores of radius 0.1 mm.

[4] As very high suctions develop, there may (in addition to the increase in tortuosity and the decrease in number and sizes of the conducting pores) also be a change in the viscosity of the (mainly adsorbed) water, tending to further reduce the conductivity. (Miller and Low, 1963).

$$I = L_f \, \Delta\theta \tag{6.18}$$

(In the special case where θ_t is saturation and θ_i is zero, $I = fL_f$, where f is the porosity.) Therefore,

$$\frac{dI}{dt} = \Delta\theta \frac{dL_f}{dt} = K\frac{\Delta H_p}{L_f} = K\frac{\Delta\theta \cdot \Delta H_p}{I} \tag{6.19}$$

where dL_f/dt is the rate of advance of the wetting front. The infiltration rate is thus seen to be inversely related to the cumulative infiltration. Rearranging Eq. (6.19), we obtain:

$$L_f \, dL_f = K\frac{\Delta H_p}{\Delta\theta}\, dt = \tilde{D}\, dt \tag{6.20}$$

where the composite term $(K\,\Delta H_p/\Delta\theta)$ can be regarded as an effective diffusivity $\tilde{D}$ for the infiltrating profile. Integration gives

$$\frac{L_f^2}{2} = K\frac{\Delta H_p}{\Delta\theta}\, t = \tilde{D}t \tag{6.21}$$

$$L_f = \sqrt{2Kt\,\Delta H_p/\Delta\theta} = \sqrt{2\tilde{D}t} \tag{6.22}$$

or

$$I = \Delta\theta\sqrt{2\tilde{D}t}, \qquad i = \Delta\theta\sqrt{\tilde{D}/2t} \tag{6.23}$$

which compares with Eqs. (6.4) and (6.5) (the difference being in the $\sqrt{2\pi}$ ratio for the weighting of $\bar{D}$ vs. $\tilde{D}$, both being approximate[7]). Thus the depth of the wetting front is proportional to $\sqrt{t}$, and the infiltration rate is proportional to $1/\sqrt{t}$.

With gravity taken into account, the Green and Ampt approach gives

$$\frac{dI}{dt} = \Delta\theta\frac{dL_f}{dt} = K\frac{H_0 - H_f + L_f}{L_f} \tag{6.24}$$

which integrates to

$$\frac{Kt}{\Delta\theta} = L_f - (H_0 - H_f)\ln\left(1 + \frac{L_f}{H_0 - H_f}\right) \tag{6.25}$$

As t increases, the second term on the right-hand side of Eq. (6.25) increases more and more slowly in relation to the increase in L_f, so that, at very large times, we can approximate the relationship by

[7] $\tilde{D}$ can be regarded as an indication of what wetting-front value must be assumed for the Green and Ampt approach to work.

$$L_f \cong \frac{Kt}{\Delta\theta} + \delta \tag{6.26}$$

or

$$I \cong Kt + \delta$$

where δ can eventually be regarded as a constant.

The Green and Ampt relationships are essentially empirical, since the value of the effective wetting-front suction must be found by experiment. For infiltration into initially dry soil, it may be of the order of -50 to -100 cm H_2O, or ~ -0.1 bar (Green and Ampt, 1911; Hillel and Gardner, 1970). However, in actual field conditions, particularly where the initial moisture is not uniform, H_f may be undefinable. In many real situations, the wetting front is too diffuse to indicate its exact location at any particular time.

G. Infiltration into Layered Soils

The effect of profile stratification on infiltration was studied by Hanks and Bowers (1962),[8] who used a numerical technique for analyzing the flow equation, and by Miller and Gardner (1962), who conducted experiments on the effect of thin layers sandwiched into otherwise uniform profiles. A conducting soil must have continuous matric suction and hydraulic-head values throughout its length, regardless of layering sequence. However, the wetness and conductivity values may exhibit abrupt discontinuities at the interlayer boundaries.

One typical situation is that of a coarse layer of higher saturated hydraulic conductivity, overlying a finer-textured layer. In such a case, the infiltration rate is at first controlled by the coarse layer, but when the wetting front reaches and penetrates into the finer-textured layer, the infiltration rate can be expected to drop and tend to that of the finer soil alone. Thus, in the long run, it is the layer of lesser conductivity which controls the process. If infiltration continues for long, then positive pressure heads (a "perched water table") can develop in the coarse soil, just above its boundary with the impeding finer layer.

In the opposite case of infiltration into a profile with a fine-textured layer over a coarse-textured one, the initial infiltration rate is again determined by the upper layer. As water reaches the interface with the coarse lower layer, however, the infiltration rate may decrease. Water at the wetting front is normally under suction, and this suction may be too high to permit entry into the relatively large pores of the coarse layer. This explains the observation

[8] This technique was used by Green *et al.* (1962) to estimate infiltration in the field.

(Miller and Gardner, 1962) that the wetting-front advance stops for a time (though infiltration at the surface does not stop) until the pressure head at the interface builds up sufficiently to penetrate into the coarse material. Thus, a layer of sand or gravel in a medium or fine-textured soil, far from enhancing water movement in the profile, may actually impede it. The lower layer, in any case, cannot become saturated, since the restricted rate of flow through the less permeable upper layer cannot sustain flow at the saturated hydraulic conductivity of the coarse lower layer (except when the externally applied pressure, i.e., the ponding depth, is large).

The steady-state downflow of water through a two-layer profile into a free-water table beneath was analyzed by Takagi (1960). Where the upper layer is less pervious than the lower, negative pressures (suctions) were shown to develop in the lower layer, and these can remain constant throughout a considerable depth range.

H. Infiltration into Crust-Topped Soils

A very important special case of a layered soil is that of an otherwise uniform profile which develops a crust, or seal, at the surface. Such a seal can develop under the beating action of raindrops (Ekern, 1950; McIntyre, 1958; Tackett and Pearson, 1965), or as a result of the spontaneous slaking and breakdown of soil aggregates during wetting (Hillel, 1960). Surface crusts are characterized by greater density, finer pores, and lower saturated conductivity than the underlying soil. Once formed, a surface crust can greatly impede water intake by the soil (Fig. 6.6), even if the crust is quite

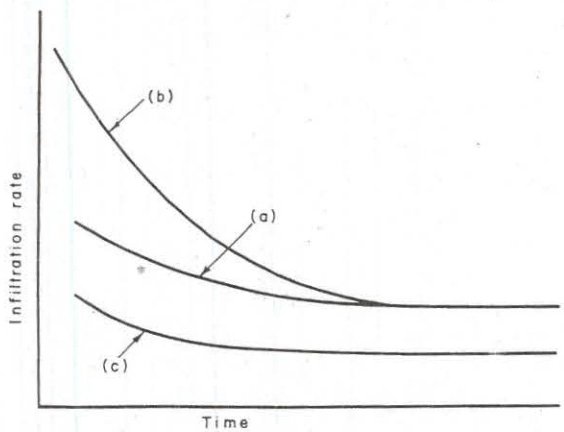

Fig. 6.6. Infiltrability as a function of time: (a) in a uniform soil; (b) in a soil with a more porous upper layer; and (c) in a soil covered by a surface crust.

thin (say, not more than several millimeters in thickness) and the soil is otherwise highly permeable. Failure to account for the formation of a crust can result in gross overestimation of infiltration.

An analysis of the effect of a developing surface crust upon infiltration was carried out by Edwards and Larson (1969), who adapted the Hanks and Bowers (1962) numerical solution to this problem. Hillel (1964), and Hillel and Gardner (1969, 1970) used a quasianalytical approach to calculate fluxes during steady and transient infiltration into crust-capped profiles from knowledge of the basic hydraulic properties of the crust and of the underlying soil.

The problem is relatively simple in the case of steady infiltration. Steady-state conditions require that the flux through the crust q_c be equal to the flux through the subcrust "transmission zone" q_u:

$$q_c = q_u$$

or

$$K_c\left(\frac{dH}{dz}\right)_c = K_u\left(\frac{dH}{dz}\right)_u \tag{6.27}$$

where K_c, $(dH/dz)_c$, K_u, and $(dH/dz)_u$ refer to the hydraulic conductivity and hydraulic-head gradient of the crust and underlying transmission zone, respectively. The gradient through the transmission zone tends to unity when steady infiltration is approached, as the suction gradient decreases with the increase in wetting depth, eventually leaving the gravitational gradient as the only effective driving force. In the absence of a suction-head gradient in the zone below the crust, we obtain (with the soil surface as our reference level)

$$q = K_u(\psi_u) = K_c \frac{H_0 + \psi_u + z_c}{z_c} \tag{6.28}$$

where $K_u(\psi_u)$ is the unsaturated hydraulic conductivity of the subcrust zone, a function of the suction head ψ_u which develops in this zone, beginning just under the hydraulically impeding crust; H_0 is the positive hydraulic head imposed on the surface by the ponded water; and z_c is the vertical thickness of the crust.

Where the ponding depth H_0 is negligible and the crust itself is very thin and of low conductivity (e.g., where z_c is very small in relation to the suction ψ_u which forms at the subcrust interface), we can assume the approximation

$$q_u = q_c = K_c \frac{\psi_u}{z_c} \tag{6.29}$$

The condition that the crust remain saturated even while its lower part will be under suction is that its critical air-entry ψ_a not be exceeded (i.e., $\psi_u < \psi_a$).

This, together with the condition that the subcrust hydraulic-head gradient approximate unity, leads to the approximation

$$\frac{K_u}{\psi_u} = \frac{K_c}{z_c} = \frac{1}{R_c} \tag{6.30}$$

i.e., the ratio of the hydraulic conductivity of the underlying soil transmission zone to its suction is approximately equal to the ratio of the crust's (saturated) hydraulic conductivity to its thickness. The latter ratio is the reciprocal of the hydraulic resistance per unit area of the crust R_c.[9] Also, by Eq. (6.28),

$$q = K_u(\psi_u) = \psi_u/R_c \tag{6.31}$$

Where the unsaturated conductivity of the underlying soil bears a known single-valued relation to the suction, it should be possible to calculate the steady infiltration rate and the suction in the subcrust zone on the basis of the measurable hydraulic resistance of the crust. Where the relation of matric suction to water content is also known, it should be possible to infer the subcrust water content during steady infiltration.

Employing a K vs. ψ relationship of the type $K = a\psi^{-n}$ (where a, and n are characteristic constants of the soil), Hillel and Gardner (1969) obtained the following[10]:

$$q = \frac{a^{1/(n+1)}}{R_c^{\,n/(n+1)}} = \frac{B}{R_c^{\,n/(n+1)}} \tag{6.32}$$

$$\psi_u = (aR_c)^{1/(n+1)} = BR_c^{\,1/(n+1)} \tag{6.33}$$

where $B = a^{1/(n+1)}$ is a property of the subcrust soil. The theoretical consequences of Eqs. (6.32) and (6.33) are illustrated in Fig. 6.7. These equations indicate how the infiltration rate decreases, and the subcrust suction increases, with increasing hydraulic resistance of the crust. Gardner (1956) has shown that the values of a and of n generally increase with increasing coarseness, textural as well as structural, of the soil. Sands may have n values of four or more, whereas clayey soils may have n values of about two. Tillage may pulverize and loosen the soil, thus increasing n, whereas compaction may have the opposite effect.

Both the crust and the underlying soil are seen to affect the infiltration rate and suction profile, and the crust-capped soil is thus viewed as a self-adjusting system in which the physical properties of the crust and underlying

[9] A distinction is made between the hydraulic resistance per unit area, defined as above, and the hydraulic resistivity, the latter being equal to the reciprocal of the conductivity.

[10] The relation of conductivity to suction does not always obey so simple an equation as $K = a\psi^{-n}$. An alternative expression, proposed by Hillel and Gardner (1969), may have more general validity: $K = K_s(\psi_a/\psi)^n$, for $\psi > \psi_a$.

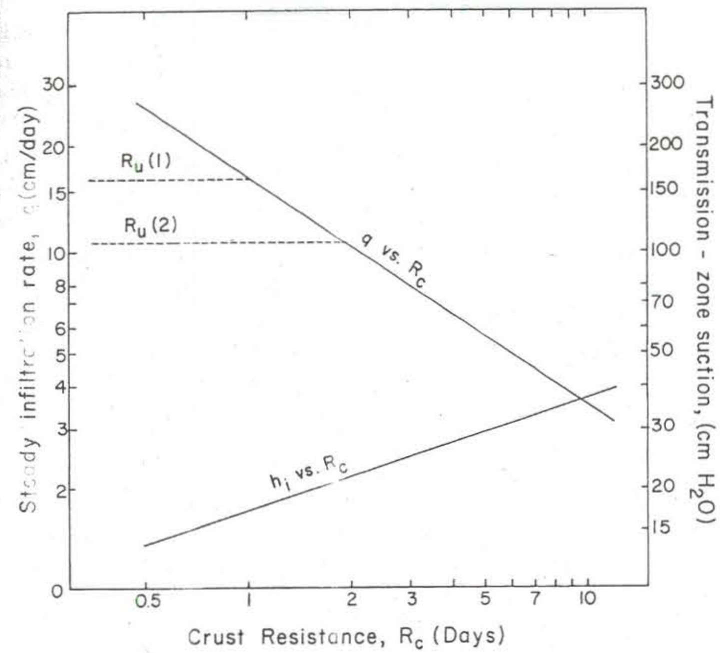

Fig. 6.7. Theoretical effect of crust resistance upon flux and subcrust suction during steady infiltration into crust-capped columns of a uniform soil with $n = 2$, $a = 4.9 \times 10^3$. The broken lines (1) and (2) indicate the hypothetical effect of subcrust hydraulic resistance R_u: $R_u(1) < R_u(2)$. The decreasing q vs. R_c curve applies only where the hydraulic conductance of the subcrust layers is not limiting. (After Hillel and Gardner, 1969.)

soil interact in time to form a steady infiltration rate and moisture profile. In this steadily infiltrating profile the subcrust suction which develops is such as to create a gradient through the crust and a conductivity in the subcrust zone which will result in an equal flux through both layers.

The problem is rather more complicated in the prevalent case of transient infiltration into an initially unsaturated profile, during which the flux, the wetting depth, the subcrust suction, and the conductivity might all be changing with time.

Assuming the Green and Ampt conditions (Section 6F), and with H_0 negligible, Hillel and Gardner (1970) recognized three stages during transient infiltration into crusted profiles: an initial stage, in which the rate is finite and dependent on crust resistance R_c and on an effective subsoil suction; an intermediate stage, in which cumulative infiltration I increases approximately as the square root of time; and a later stage, in which I can be expressed as the sum of a steady and a transient term, the latter becoming negligible at long times. I was shown to decrease with increasing R_c, particularly in coarse-textured and coarse-structured soils. Experimental data indicate that the cumulative infiltration curves of crusted profiles scale as the square root of their transmission-zone diffusivities. Thus, infiltration into a crusted profile can be described by the approximation that water enters into the subcrust soil at a nearly constant suction, the magnitude of which is determined by crust resistance and hydraulic characteristics of the soil.

Where the gravity effect is negligible (e.g., in horizontal flow or during the initial stages of vertical infiltration into an initially dry medium of high matric suction), the infiltration vs. time relationship was given by:

$$I = \sqrt{K_u^{2} R_c^{2} (\Delta\theta)^2 + 2K_u H_f \,\Delta\theta\, t} - K_u R_c \tag{6.34}$$

Where the gravity effect is significant, the expression given is

$$L_f = \frac{K_u t}{\Delta\theta} + (H_f - K_u R_c) \ln\left[\frac{H_f + (K_u t/\Delta\theta) + \delta(t)}{H_f}\right] \tag{6.35}$$

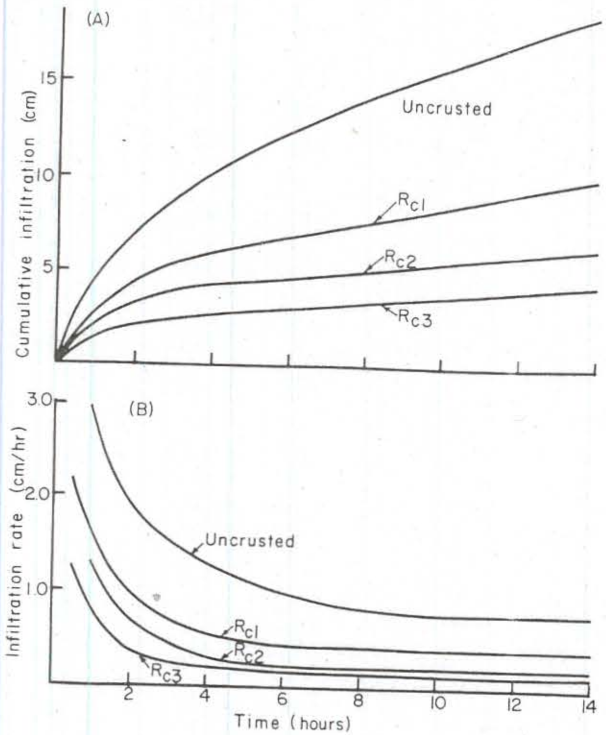

Fig. 6.8. Time dependence of cumulative infiltration (A) and of infiltration rate (B) for uncrusted and crusted columns of Negev loess. Crust resistance values R_{c1}, R_{c2}, R_{c3} are 3.2, 9.1, and 17 days, respectively (after Hillel and Gardner, 1969).

where the correction term $\delta(t)$ becomes negligibly small as t increases. Thus, L_f can be expressed as the sum of a steady and a transient term. Some experimental results are shown in Fig. 6.8.

I. Rain Infiltration

When rain or sprinkling intensity exceeds soil infiltrability, the infiltration process is the same as in the case of shallow ponding. If rain intensity is less than the initial infiltrability value of the soil, but greater than the final value, then at first the soil will absorb water at less than its potential rate and flow in the soil will occur under unsaturated conditions; however, if the rain is continued at the same intensity, and as soil infiltrability decreases, the soil surface will eventually become saturated and henceforth the process will continue as in the case of ponding infiltration. On the other hand, if rain intensity is at all times lower than soil infiltrability (i.e., lower than the saturated hydraulic conductivity), the soil will continue to absorb the water as fast as it is applied without ever reaching saturation. After a long time, as the suction gradients become negligible, the wetted profile will attain a wetness for which the conductivity is equal to the water application rate, and the lower this rate, the lower the degree of saturation of the infiltrating profile. This effect is illustrated in Fig. 6.9.

The process of infiltration under rain or sprinkler irrigation was studied by Youngs (1960) and by Rubin and Steinhardt (1963, 1963), Rubin *et al.* (1964), and Rubin (1966). The latter author, who used a numerical solution of the flow equation for conditions pertinent to this problem, recognized three modes of infiltration due to rainfall: (1) *nonponding infiltration*, involving rain not intense enough to produce ponding; (2) *preponding infiltration*, due to rain that can produce ponding but that has not yet done so; and (3) *rainpond infiltration*, characterized by the presence of ponded water. Rainpond infiltration is usually preceded by preponding infiltration, the transition between the two being called *incipient ponding*. Thus, nonponding and preponding infiltration are *rain-intensity-controlled* (or *flux-controlled*), whereas rainpond infiltration is controlled by the pressure (or depth) of water above the soil surface, as well as by the suction conditions and conductivity relations of the soil. Where the pressure at the surface is small, rainpond infiltration, like ponding infiltration in general, is *profile-controlled.*

In the analysis of rainpond or ponding infiltration, the surface boundary condition generally assumed is that of a constant pressure at the surface, whereas in the analysis of nonponding and preponding infiltration, the water flux through the surface is considered to be constant, or increasing. In actual field conditions, rain intensity might increase and decrease alternately, at times exceeding the soil's saturated conductivity (and its infiltrability) and

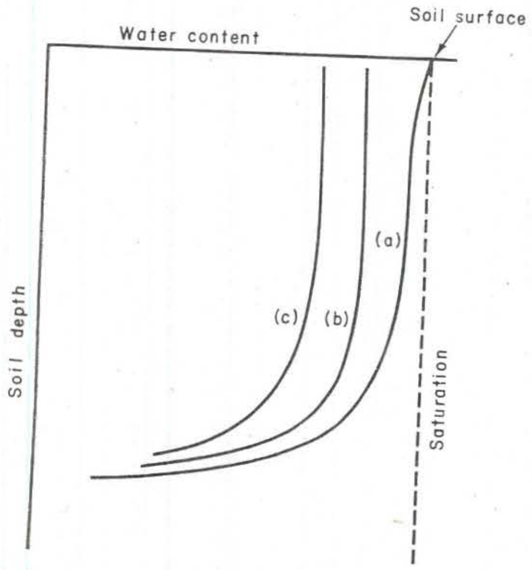

Fig. 6.9. The water-content distribution profile during infiltration: (a) under ponding; (b) under sprinkling at relatively high intensity; and (c) under sprinkling at a very low intensity.

at other times dropping below it. However, since periods of decreasing rain intensity involve complicated hysteresis phenomena, the analysis of composite rainstorms is very difficult and has not yet been carried out satisfactorily.

Rubin's analysis is based on the assumption of no hysteresis. The falling raindrops are taken to be so small and numerous that rain may be treated as a continuous body of "thin" water reaching the soil surface at a given rate. Soil air is regarded as a continuous phase, at atmospheric pressure. The soil is assumed to be uniform and stable (i.e., no fabric changes such as surface crusting).

We shall briefly review the consequences of Rubin's analysis in qualitative terms. If a constant pressure head is maintained at the soil surface (as in rainpond infiltration), then the flux of water into this surface must be constantly decreasing with time. If a constant flux is maintained at the soil surface, then the pressure head at this surface must be constantly increasing with time. Infiltration of constant-intensity rain can result in ponding only if the *relative rain intensity* (i.e., the ratio of rain intensity to the saturated hydraulic conductivity of the soil) exceeds unity. During nonponding infiltration under a constant rain intensity q_r, the surface suction will tend to a limiting value ψ_{lim} such that $K(\psi_{\text{lim}}) = q_r$.

Under rainpond infiltration, the wetted profile consists of two parts: an upper, water-saturated part; and a lower, unsaturated part. The depth of

the saturated zone continuously increases with time. Simultaneously, the steepness of the moisture gradient at the lower boundary of the saturated zone (i.e., at the wetting zone and the wetting front) is continuously decreasing (these phenomena accord with those of infiltration processes under ponding, as described in the previous sections of this chapter). The higher the rain intensity is, the shallower is the saturated layer at incipient ponding and the steeper is the moisture gradient in the wetting zone.

Figure 6.10 describes infiltration rates into a sandy soil during preponding

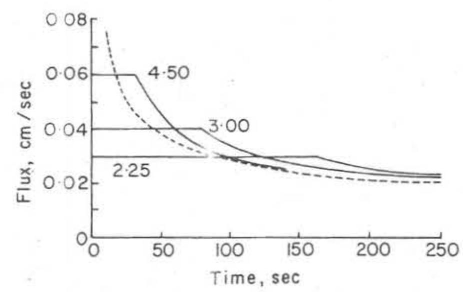

Fig. 6.10. Relation between surface flux and time during infiltration into Rehovot sand due to rainfall (solid lines) and flooding (dashed line). The numbers labeling the curves indicate the magnitude of the relative rain intensity (after Rubin, 1966).

and rainpond infiltration under three rain intensities. The horizontal parts of the curves correspond to preponding infiltration, and the descending parts to rainpond infiltration periods. As pointed out by Rubin (1966), the rainpond infiltration curves are of the same general shape and approach the same limiting infiltration rate, but they do not constitute horizontally displaced parts of a single curve, and do not coincide with the infiltration rate under flooding, which is shown as a broken line in the same graph.

The process of rain infiltration has not yet been studied in sufficient detail in the field to establish the applicability of existing theories. Complications due to the discreteness of raindrops (which causes alternate saturation and redistribution at the surface), as well as to the highly variable nature of rainstorm intensities and raindrop energies, and the unstable nature of many (perhaps even most) soils, can cause anomalies disregarded by idealized theories. Additional complications can arise in cases of air occlusion and when the soil exhibits profile or areal heterogeneity.

J. Surface Runoff

Surface runoff, or overland flow, is the portion of the rain which is not absorbed by the soil and does not accumulate on the surface, but runs down-slope and collects in gullies and streams. Runoff can occur only when rain intensity exceeds the infiltration rate. Even then, however, runoff does not begin immediately, as the excess rain first collects in surface depressions and forms puddles, whose total volume is termed the *surface storage capacity*. Only when the surface storage is filled and the puddles begin to overflow does runoff begin. The rate of the runoff flow depends upon the excess of rain intensity over the infiltration rate. Obviously, the surface storage also depends on the slope, as well as on the roughness of the soil surface.

In agricultural fields, runoff is generally undesirable, since it results in loss of water and often causes erosion, the amount of which increases with increasing rate and velocity of runoff. The way to prevent erosion is to protect the soil surface against raindrop splash (e.g., by mulching), to increase soil infiltrability and surface storage, and to obstruct overland flow so as to prevent it from gathering velocity. Maintenance and stabilization of soil aggregation will minimize slaking and detachment of soil particles by raindrops and running water. A crusted or compacted soil generally has a low infiltration rate and therefore will produce a high rate of runoff. Proper tillage, especially on the contour, can increase infiltration and surface storage capacity, thus reducing runoff (Burwell and Larson, 1969).

In arid regions, it is sometimes desirable to induce runoff artificially in order to supply water for human, industrial, or agricultural use (Hillel *et al.*, 1967).

K. Summary

An important physical property of a soil is the rate at which it can absorb water supplied to its surface. This rate, termed *soil infiltrability*, depends on the following factors:

(1) Time from the onset of the rain or irrigation: The infiltration rate is apt to be relatively high at first, then to decrease, and eventually to approach a constant rate that is characteristic for the soil.

(2) Initial water content: The wetter the soil is initially, the lower will be the initial infiltrability (owing to smaller suction gradients) and the quicker will be the attainment of the final (constant) rate, which itself is generally independent of the initial water content.

(3) Hydraulic conductivity: The higher the saturated hydraulic conductivity of the soil is, the higher its infiltrability tends to be.

(4) Soil surface conditions: When the soil surface is highly porous and of "open" structure, the initial infiltrability is greater than that of a uniform

soil, but the final infiltrability remains unchanged, as it is limited by the lower conductivity of the transmission zone beneath. On the other hand, when the soil surface is compacted and the profile covered by a surface crust of lower conductivity, the infiltration rate is lower than that of the uncrusted (uniform) soil. The surface crust acts as a hydraulic barrier, or bottleneck, impeding infiltration. This effect, which becomes more pronounced the thicker and the denser the crust, reduces both the initial and the final infiltration rate. A soil of unstable structure tends to form such a crust during infiltration, especially as the result of the slaking action of beating raindrops. In such a soil, a plant cover or a surface mulch of plant residues can serve to intercept and break the impact of the raindrops and thus help to prevent crusting.

(5) The presence of impeding layers inside the profile: Layers which differ in texture or structure from the overlying soil may retard water movement during infiltration. Perhaps surprisingly, clay layers and sand layers can have a similar effect, although for opposite reasons. The clay layer impedes flow owing to its lower *saturated* conductivity, while a sand layer retards the wetting front (where unsaturated conditions prevail) owing to the lower *unsaturated* conductivity of the sand. Flow into a dry sand layer can take place only after the pressure head has built up sufficiently for water to move into and fill the large pores of the sand.