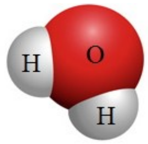
Technical Note 7. Ag Water Literacy IV



7.12 Water Entry and Redistribution In The Soil

Water enters the soil body through the process of ***infiltration*** at the soil surface, generally by downward motion. Without infiltration, water deficit in the soil could not be replenished nor could ground water reserves be built up. Flooding from runoff and overland flow would bring serious harm to communities and disrupt critical infrastructure including that of agriculture. Hence, the detailed study of infiltration is vital not only to those engaged in farming, but also in civil engineering and hydrologic modeling.

Once water has infiltrated the soil, it undergoes the process of ***redistribution***, which enables the movement of water in the soil. Water movement may be omnidirectional: downward via gravitational, pressure and matric potential; upward into the ***unsaturated zone*** from the ***saturated zone*** via matric potential or toward the soil surface via vapor potential (evaporation) or positive pressure head (artesian); horizontal or downslope in the unsaturated zone via matric, osmotic, pressure, gravitational potential, or some combination of flow. Recall that gravitational, pressure, matric, and osmotic potential are the universal forces driving water movement through the Soil-Plant-Atmosphere Continuum (SPAC), described in TN 7 Part II Section 7.4. In fact, these are the same forces that activate infiltration and redistribution processes in the soil so there's nothing new to learn.

Infiltration is a complex subject about which much has been written. Even so, sources of information accessible to the average dirt hog enthusiast are relatively few or at best, fragmented. Here we will sketch the basic outline of infiltration as a physical process; discuss the factors affecting infiltration; examine Darcy's law governing water movement in porous media; and in Section 7.13, probe deeper into quantifying infiltration and runoff, as well as consider some ***empirical*** infiltration models used in irrigation design. Throughout, we will employ real data and images collected from the field to illuminate cardinal points and convey, hopefully, an intuitive understanding of water entry and redistribution in the soil. References are given at the end for those seeking more advanced treatment of individual topics. As always, the starting point for any quest in unfamiliar intellectual territory begins by mobilizing terminology, the lingua franca upon which a mutual understanding can be forged. Infiltration is no different. Let's dig in.

Infiltration rate, denoted by the lower case letter i, is a measure of how fast a particular soil can absorb water, be it natural rainfall or irrigation. Units of measure for infiltration rate (i) are depth of water (millimeters, inches) per unit time (minutes, hours). As the soil becomes saturated the infiltration rate decreases. The maximum rate at which water can enter the soil body when in continuous contact with water at the surface at atmospheric pressure is the ***infiltration capacity***, or ***infiltrability***. When precipitation exceeds the infiltration capacity, ***runoff*** will occur unless there is some physical barrier preventing it (**Figure 7.12.1**). The opposite of runoff is ***run-on***, water that relocates from a change in elevation or soil moisture gradient, sometimes accumulating as free water in surface ***ponding*** where it's not wanted, but not leaving the field like runoff. ***Cumulative infiltration*** (symbol: capital letter I) is the accumulated depth (millimeters, inches) of infiltrated water over a given period of time (t). The *average infiltration rate* (i_{avg}) is the cumulative infiltration divided by the total time of infiltration (i/t) measured from the beginning (t=0).

Figure 7.12.1 Surface runoff from a farm field in Tennessee, USA. Runoff occurs when the precipitation rate exceeds infiltration. Soil and dissolved nutrients may be swept away or worse, become pollutants. *Image source: USDA-NRCS*

Infiltration rate is governed by a combination of gravity, pressure, and matric potential effects at the soil surface, the fluid properties of water, and the ***hydraulic conductivity*** of the underlying mineral soil. These are, in turn, determined by the following factors:

- **Soil wetness** Initial or "antecedent" water content has an effect on infiltration rate and total (cumulative) depth of infiltration (**Figure 7.12.2**). Wet soils have slower infiltration than dry soils because the matric suction gradient between the top wet layers and lower dry layers disappears. Infiltration rate decreases as the soil approaches saturation. At saturation, an equilibrium is established where the infiltration rate is constant. This steady-state or *basic infiltration rate* is the soil's saturated *hydraulic conductivity* (symbol: K_s), with dimension of length (distance) moved per unit time, e.g. millimeters per hour.
- **Soil mineral composition** Sands are granular, permitting rapid infiltration and water movement. Clays have a slower infiltration rate because there are smaller pores between particles and greater matric suction

(surface tension and adhesive forces) due to greater particle surface area. Some clay minerals shrink and swell in response to cycles of wetting and drying, such that infiltration rate depends on the state of mineral hydration. Silty soils are intermediate (**Figure 7.12.3**). Typical values for steady infiltration rate for different soil types are given in **Table 7.12.1**. These values may be adjusted higher in well-aggregated soils, or lower in the presence of a soil crust or shallow impervious layer.

- **Soil structure** Soils with good, water stable aggregate structure have greater infiltration capacity compared to unstructured (massive) soils, or soils with weak, unstable structure (**Figure 7.12.4**).

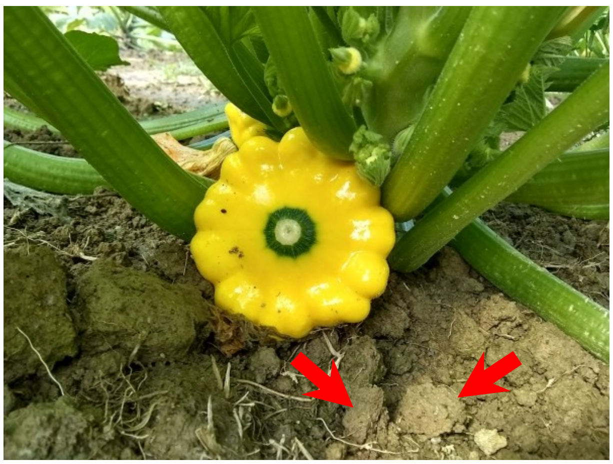


Figure 7.12.4 Well-aggregated soil structure is perhaps the most important factor affecting infiltration and runoff. Crops like this summer squash luxuriate when every drop infiltrates. *Image: Cypress Prong Farms*

Table 7.12.1 Infiltration rates of various soils.		
Soil type	Steady infiltration rate Average mm/hr	Range mm/hr
Sands	50	25-250
Sandy loam	25	13-76
Loam	13	8-20
Clay loam	8	2.5-15
Silty clay	2.5	0.5-5
Clay and Sodic clay	0.5	0.1-1
Source: Waller and Yitayew, 2016; Hillel, 1998.		

- **Organic matter** Generally, soils that are high organic matter infiltrate quicker than soils low in organic matter. Organic matter imbibes water much like a sponge. Water enters the dead cells of plant parts and microorganisms as well as being held on the particle surfaces. The presence of surficial crop or cover crop residues is perhaps the most important factor affecting infiltration because they shield soil aggregates from breakdown by reducing the impact of water droplets, and provide the raw material needed by microorganisms for aggregate formation and stabilization.

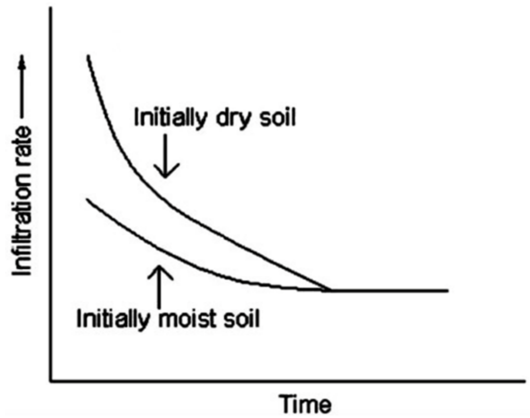


Figure 7.12.2 When conducting an infiltration test the initial water content should always be measured and reported. An initially moist soil near field capacity will exhibit a much lower infiltration rate and total infiltration than a dry soil. The basic or steady infiltration rate will be the same (for the same soil) regardless of initial water content.

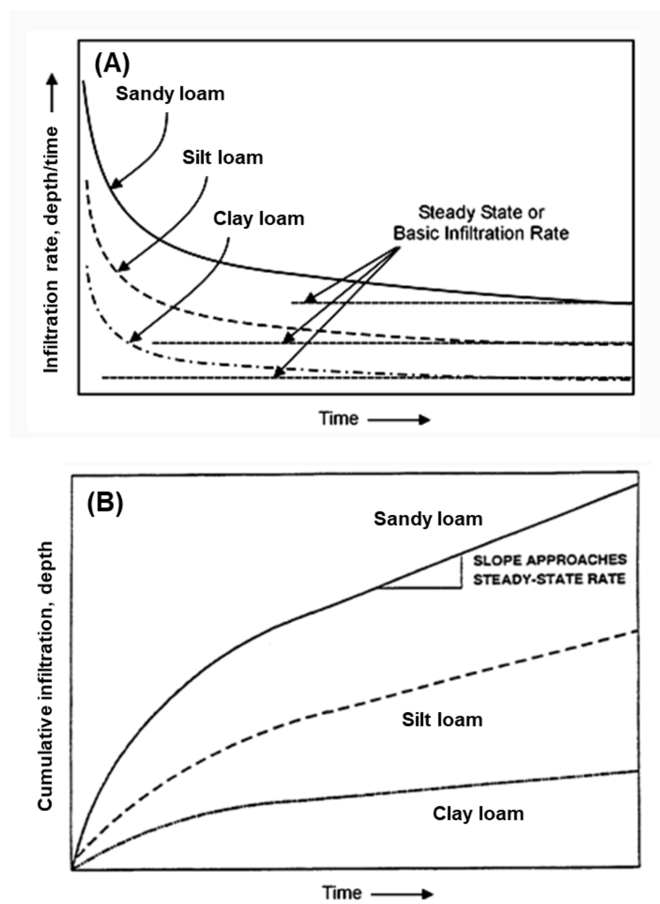


Figure 7.12.3 Infiltration rate (**A**) and cumulative infiltration depth (**B**) proceeding irrigation or precipitation for three contrasting soil types. *Source: adapted from Oklahoma State Univ.*

- **Soil depth** Shallow impervious soil layers reduce the total amount of water that can enter compared to deep soils.
- **Soil porosity** Compaction due to human and/or vehicle traffic, intensive livestock grazing, and farm equipment presses soil particles closer together, reducing pore space and infiltration. Raindrops falling on open ground with no protective layer of mulch can detach fine soil particles that are washed into surface pores, effectively reducing ***permeability***, and forming surface seals that decelerate infiltration rates.
- **Soil temperature** Freezing affects infiltration rate owing to the formation of ice crystals and structural changes that influence hydraulic conductivity. Effects on infiltration rate may be positive or negative.
- **Rainfall intensity** The kinetic energy of raindrops impacting on bare soils causes mechanical breakdown of aggregates, surface sealing, and mineral compaction that greatly reduces infiltration rates (**Figure 7.12.5**).
- **Soluble Salts** Calcium and magnesium assist mineral flocculation while sodium does the opposite. Sodic-affected clay soils have the lowest infiltration rates and highest potential for runoff (**Table 7.12.1**).

When water is added to the soil, from natural agents like rainfall or purposely via irrigation, initially it fills the space between soil particles at the surface. Thereafter, the force of gravity pulls the water downward in pore channels and voids until matric suction forces equal gravitational pull. At this point, the soil is at ***field capacity***. Field capacity is the amount of water the soil can hold against the pull of gravity. When the soil is wetter than field capacity, ***saturated flow*** water movement through the soil begins under the pull of gravity. When the soil is drier than field capacity, ***unsaturated flow*** water movement occurs, where matric suction forces dominate, and the gravitational force is too weak to cause flow. Water descending through the unsaturated soil profile is called ***percolation.*** Percolating water carries away dissolved mineral salts and other substances. The transfer of soluble substances in percolating water is called ***leaching***. Movement of gravitational water into open tile, channels, and ditches, is called ***drainage***. Another term related to drainage is ***seepage***, which describes water movement through a porous media from a source of supply such as reservoir or irrigation canal. Raising the water table in the ***effective root zone*** via subirrigation takes advantage of matric seepage (**Figure 7.12.6**). On the other hand, water may escape from reservoirs, irrigation canals, and the root zone as gravitational seepage (**Figure 7.12.7**).

Other terms often encountered in the literature on water motion are: ***sorptivity***, ***diffusivity***, ***hydraulic conductivity***, and ***transmissivity***. We'll define each of these specialized terms in turn.

Sorptivity (symbol: S) describes the ability of a medium (soil in this case) to absorb or desorb liquid water, generally by capillarity (Philip, 1957). As such, sorptivity as a physical process dominates in early-stage infiltration, helping to suck water into the soil matrix at atmospheric pressure. The term

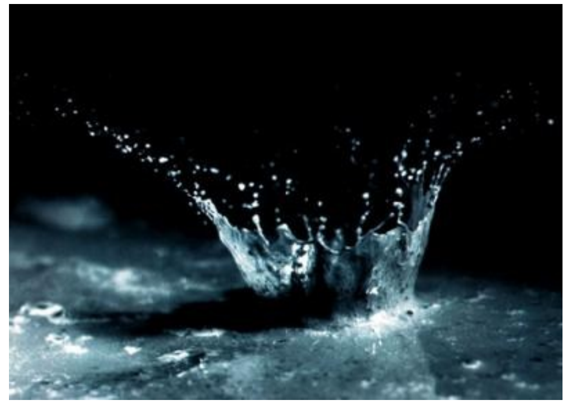


Figure 7.12.5 The impact of raindrops hitting the exposed soil surface can be devastating, effectively shattering aggregates, dislodging soil particles to be washed into pores or lost in runoff. *Source: USDA -NRCS*

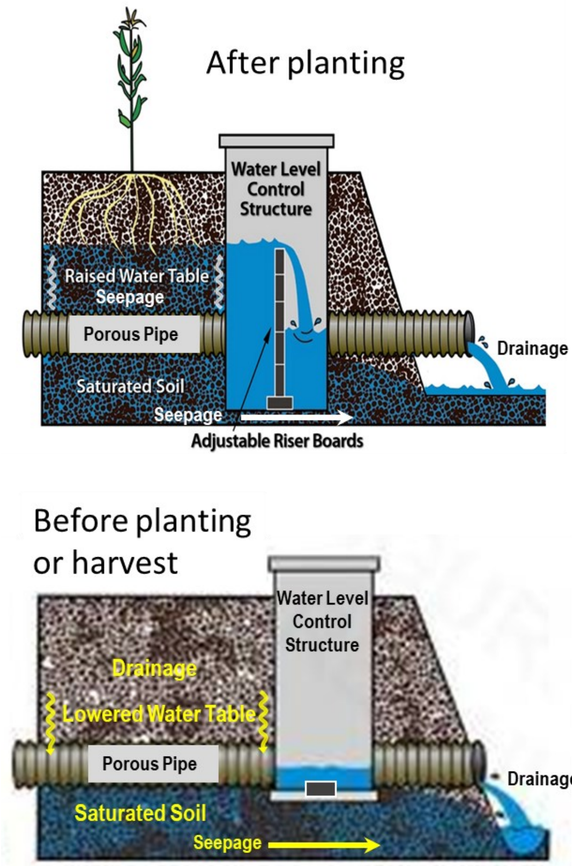


Figure 7.12.6 Controlled drainage infrastructure showing a raised water table for subirrigation after planting. Adjustable riser boards block flow through the drain tile to raise the water table. As the water table rises, seepage flow occurs around the crop root zone and some escapes under the control structure and at field edges. With riser boards removed there is free drainage and the water table is lowered to facilitate field operations. *Source: adapted from Purdue Univ.*

Figure 7.12.7 Surface irrigation infrastructure showing lines of seepage flow into adjacent crop lands. In arid regions seepage through irrigation canals like this can raise the water table in surrounding areas, impeding drainage, and causing salt build up via surface evaporation.

sorptivity is also used in other contexts denoting the accumulation of a substance within another substance when the mechanisms of accumulation (*absorption* into the body or *adsorption* on the surface of particles), or removal (*precipitation*) are unknown. Sorptivity depends on the initial water content (sometimes called *antecedent*) as well as the saturated soil water content; it is, in turn, defined analytically as a function of water capacity and *diffusivity*. Sorptivity is determined by relating cumulative infiltration (I) to the square root of time: $S = I\sqrt{t}$. That is to say, sorptivity is directly proportional to cumulative infiltration at a given time t. Sorptivity is mainly used in Phillip's model of infiltration, mentioned further down in Section 7.13.

The ***diffusivity*** (symbol: D) of soil water is a hydraulic property relating the motion of water to relative differences in water content ('wetness" or "degree of saturation") rather than matric potential. In technical terms, soil water diffusivity is the product of the *hydraulic conductivity* and the *specific water capacity*, i.e. the change in water content in a unit volume of soil per unit change in water potential (Hillel, 1994). ***Hydraulic conductivity*** (symbol: K) measures the soil's ability to conduct water through pores or voids, whether saturated (K_s) or unsaturated (K_ψ). In practice, diffusivity and conductivity are related hydraulic properties that can be used to analyze the behavior of soil water (**Figure 7.12.8**). ***Transmissivity*** describes the movement of water through a layer and is mainly used in hydrogeology to quantify water table fluctuations and aquifer recharge.

Figure 7.12.3A shows typical water infiltration curves for three soil types. In all cases water intake is greatest at the beginning, leveling off gradually in sandy soil, and more abruptly in heavier clay loam soil. The reason for this is that sand is chemically inert so its permeability to water depends solely on porosity and the blockage of pores by particulate matter washing in during infiltration. Porosity and washing also affect clay soils, but their electrostatically charged particle surfaces also interact with water molecules, causing some to swell and reducing water movement. Eventually, infiltration slows to a continuous steady state as the ***hydraulic gradient*** in the wetted region is cancelled.

The permeability of soil is controlled by the least permeable layer in the profile. Usually this is the most compacted layer or that of the highest clay content. Information about soil permeability is one of the most important hydraulic factors affecting land-use suitability. Irrigation and drainage system design, and that for septic tank leach fields, all depend on accurate knowledge of the soil's permeability. The most commonly used indicator of soil permeability is *hydraulic conductivity* (K), which is a quantitative measure of the ability of a porous media to transmit water. The U.S. Department of Agriculture (USDA) divides soils into eight interpretive permeability classes based on permeability rates for each horizon (**Table 7.12.2**). These are qualitative indicators for soil survey purposes and should not be confused with hydraulic conductivity.

Table 7.12.2 USDA Interpretive Soil Permeability Classes.

Permeability Class	Criteria Estimated, mm/hr (in./hr)
Impermeable	<0.038 (<0.0015)
Very slow	0.038-1.5 (0.0015-0.06)
Slow	1.5-5.1 (0.06-0.2)
Moderately slow	5.1-15 (0.2-0.6)
Moderate	15-51 (0.6-2.0)
Moderately rapid	51-150 (2.0-6.0)
Rapid	150-510 (6.0-20.0)
Very rapid	>510 (>20.0)

Source: National Soil Survey Center, Field Book for Describing and Sampling Soils. NRCS USDA, 2002.

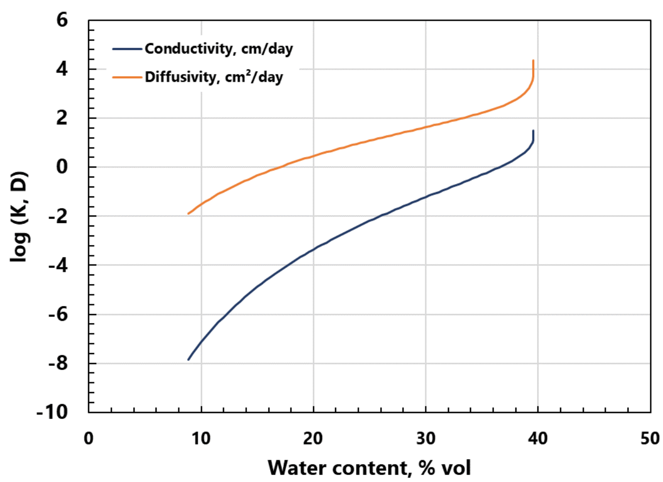


Figure 7.12.8 Soil water diffusivity and unsaturated conductivity curves for a cultivated Wedowee sandy loam soil in North Carolina, USA. The curves illustrate decreasing convective flow (increasing time per unit distance or area) with rising degree of saturation. Note that diffusivity changes less than conductivity over the range of volumetric water contents. *Data source: Walters-NCSU.*

The knowledge of hydraulic conductivity is extremely important to understanding soil conditions affecting water redistribution. **Figure 7.12.9** portrays the saturated hydraulic conductivity (K_s) measured at two depths beneath two contrasting land management systems: no tillage and conventional moldboard plow tillage with disking. The crop was maize, the test location situated in the crop interrow, and the timing of the measurements late in the season around R5 (dent).

Two important points can be taken from this data:

1. Under no tillage, the saturated hydraulic conductivity to 30 cm deep was "slow" according to the permeability classes in **Table 7.12.2** but K_s did not differ significantly with depth.
2. Under moldboard plow tillage with disking, K_s increased nearly 8-fold compared to no tillage and varied significantly with depth.

What explains the difference in behavior?

First, no tillage involves planting directly into the previous year's crop residue. There is zero soil disturbance apart from the narrow slit created by the seed openers, which is sealed closed immediately following insertion. Over time, continuous no-tillage has caused the plow layer (Ap horizon) to undergo settlement, creating a large network of micropores and fewer macropores. In turn, saturated hydraulic conductivities are relatively low because the resistance to flow through narrow channels is high compared to wide channels. Because the surficial 0-15 cm and subsurface 15-30 cm zones are relatively homogenous in terms of porosity, K_s in the two zones is also similar. Compare this to moldboard plow tillage with disking, where the soil is inverted annually to 30 cm deep followed by disking to level and break up clods.

In **Figure 7.12.9** it can be seen that K_s values for moldboard plow tillage with disking were nearly two-fold greater in the surficial 15 cm zone compared to the 30 cm zone. This is because the soil has retained memory of plowing and disking 106 days after upheaval, where a residual network of relatively large macropores sustains higher K_s values compared to no tillage. Beneath 15 cm, the weight of the overburden coupled with alternate wetting and drying cycles has partially consolidated the soil mass, reducing conductivity in that zone. Also note that K_s measured in the surface 15 cm, 27.6 mm/hr, approximates K_s estimated by the SPAW software program shown in **Figure 7.12.10** using measured particle size distribution for the same soil so we know this K_s value is reasonable.

Of course, there are tradeoffs with conductivity: high K_s values readily transmit infiltrated water down in the profile, preventing waterlogging. On the other hand, a short residence time in the root zone means the water deficit will have to be replenished with greater frequency to avoid stress. Leaching of dissolved nutrients away from the root zone is also accelerated. A balance must be struck optimizing free drainage with retention. In many situations, artificial drainage is the only solution if the soil is not already in good balance. The soil's K_s value is a key parameter for evaluating a soil's drainage capability.

Here we pause to introduce, however gently, the famous equation known as Darcy's law. Named after Henry Darcy, a French engineer, who derived it empirically in the mid-19th century in a series of investigations on seepage rates through sand filters in the city of Dijon. No understanding of fluid flow through porous media is complete without appealing to Monsieur Darcy's peculiar insights.

Darcy's law is a mathematical statement relating the vertical flow rate across an area in a permeable media, with

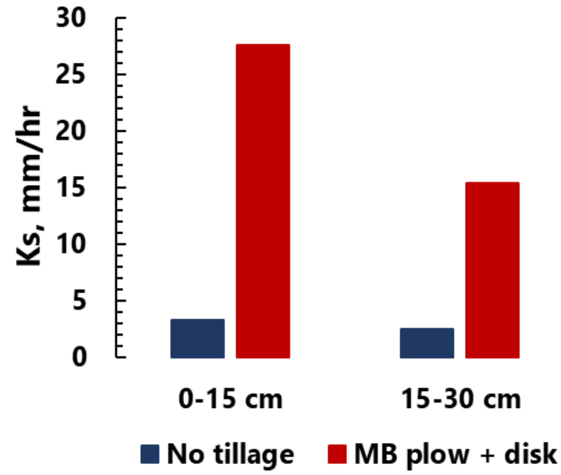


Figure 7.12.9 Average saturated hydraulic conductivity, K_s, measured beneath two tillage systems and two depth increments 106 days after planting maize in a Wedowee sandy loam soil, North Carolina, USA. Ks was measured with a constant head borehole permeameter in four replicates for each of two soil depths. *Source: Walters-NCSU unpublished data*.

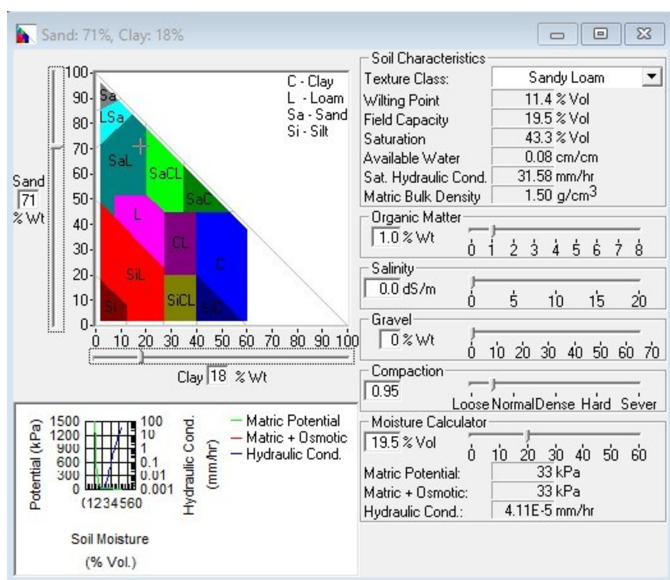


Figure 7.12.10 Soil water characteristics calculator, a component of the Soil-Plant-Air-Water (SPAW) software program (link).

hydraulic gradient. Recall that a gradient is simply a slope downward or upward along a path. A hydraulic gradient may be understood as the energy supplied or consumed to drive fluid flow along that path. A hydraulic gradient may develop due to elevation, pressure, or velocity heads, or a combination (**Figure 7.12.11**).

Darcy's Law is usually written as:

$$Q = \frac{AK\Delta P}{L} \qquad [\text{Eq. }1]$$

where

Q = volumetric water discharge ($L \times L \times L$: cubic units cm^3, m^3, etc.) per unit time

A = area through which the water passes (($L \times L$: square units cm^2, m^2, etc.)

K = hydraulic conductivity (L: length units cm, mm, etc.) per unit time (t = seconds, hours, days)

ΔP = difference in water potential energy level between two points: hydraulic gradient (unitless)

L=distance between two points in a hydraulic gradient (L: length units cm, mm, etc.)

Darcy's law states that the driving force of a liquid through porous media is proportional to the hydraulic gradient, often denoted by the lower case letter i, *and* also proportional to the ability of the media to conduct fluid. We have demonstrated that soil is a mixture of solid particles, liquid, and air voids (TN 7 Part II Section 7.4). As such, soil manifests the essential properties of porous media needed to evaluate water redistribution using the Darcy equation.

Let's break Darcy's equation down in sequence to better understand.

The first term Q in Eq. [1] relates to the volumetric discharge rate:

$$Q\left(\frac{cm^3}{\sec}\right)$$

with units of volume H_2O per unit time. Here we'll use cubic centimeters per second but the units can be scaled any way depending on the problem. Now, we'll add the second and third terms:

$$Q\left(\frac{cm^3}{\sec}\right) = A\left(m^2\right) \times K\left(\frac{m}{\sec}\right)$$

Here we see that Q equals the product of area A and hydraulic conductivity K. Hydraulic conductivity is understood as the *proportionality factor* in Darcy's equation. In the soil K values remain constant over time and depend on the size and configuration of pores and properties of liquid water:

$$K = k\frac{\gamma}{\eta}$$

where k is the *coefficient of permeability* with units of Darcys or cm^2; γ is the unit weight of water which is a product of its density ρ, and the acceleration due to gravity, g, with units

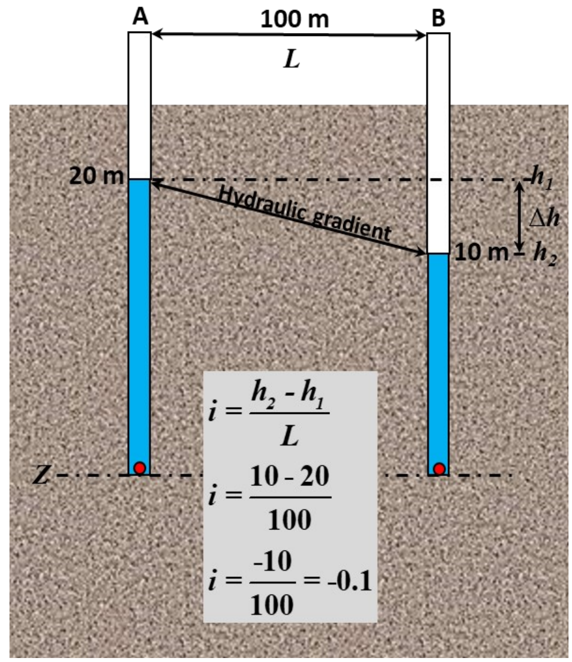


Figure 7.12.11 Illustration of a simple hydraulic gradient. The free water surface in well A is 20 m above the reference datum z, and 10 m above the same reference datum in well B. The hydraulic gradient i is given by $\Delta h/L$ or h_2 - h_1 divided by the distance L between well A and B, shown above. Water will spontaneously flow from well A to well B in response to the difference in apparent elevation head. The volume of flow is determined by magnitude of the gradient as well as the saturated hydraulic conductivity of the soil. Hydraulic gradients can develop in any direction.

g/cm^3 divided by η, its dynamic viscosity with units centipoise or Pascal-second. Note that the coefficient of permeability, k, represents an intrinsic property of porous media i.e. the size and configuration of its pores. Whereas the term

$$\frac{\gamma}{\eta}$$

represents the properties of water. As such, the proportionality constant K in Darcy's law accounts for the properties in both the media *and* liquid water. This is an important point. Typical steady-state values for K in saturated soil are given in **Table 7.12.1**.

Darcy's law requires a driving force, and this is where the last term on the right comes into play:

$$Q\left(\frac{cm^3}{\sec}\right) = A\left(m^2\right) \times K\left(\frac{m}{\sec}\right) \times \frac{\Delta P}{L}$$

The term $\frac{\Delta P}{L}$ is the hydraulic gradient, the prime mover.

This can be rewritten as $\dfrac{h_2 - h_1}{L}$

which the reader will recognize as identical to the equation for hydraulic gradient shown in **Figure 7.12.11**. We can also apply the notation of derivatives to express the magnitude of the driving force in terms of hydraulic gradient per unit length:

$$Q\left(\frac{cm^3}{\sec}\right) = A\left(m^2\right) \times K\left(\frac{m}{\sec}\right) \times \frac{\Delta P}{L}$$

The term $\Delta P/L$ can be understood as the *difference in total energy potential* between two points divided by the distance between those points. Thus, Darcy found that Q, the volumetric flow rate per unit time, was proportional to the difference in hydraulic gradient per unit length, $\Delta P/\Delta L$, multiplied by a constant factor K and cross-sectional area A. In mathematics the proportionality factor K is called the slope, the magnitude of which controls Q under the same hydraulic gradient as shown for two contrasting soils in **Figure 7.12.12**. The hydraulic gradient per se has no units, so it doesn't carry anything over to left side of Darcy's equation.

Alternatively, Q can be expressed in terms of volumetric ***flux***, denoted by lower case q

where q = the volumetric discharge per unit time, Q, divided by the area A with units of velocity as noted above. In turn,

Eq.[1] can be rearranged to solve for the volumetric flux, *q*, *as*

$$q = \frac{Q}{At} = \frac{L^3}{L^2 \times t} = \frac{L}{t} \ (velocity) \qquad [\text{Eq. 2}]$$

$$\frac{Q}{A} = -Ki \ or \ q = -Ki$$

where lower case letter i denotes the hydraulic gradient $\Delta h/\Delta L$ $(=\Delta P/\Delta L)$. The negative sign in front of K simply indicates the direction of flow.

As explained back in TN 7 Part I Section 7.3, water always flows from high to low potential, or down the hydraulic gradient. At the free water surface, the pressure potential is zero. Beneath a free water surface, such as the water table, the pressure is always positive. Negative pressure head, denoted 'suction', 'tension' or 'vacuum' pressure, indicates pressure *less than* that of the atmosphere, a condition that arises when the soil becomes unsaturated. When Δh = 0, no flow should be detected according to Darcy's law regardless of the soil's K value.

To further illustrate, the soil profile depicted in **Figure 7.12.13** is divided into idealized hydrologic zones, relating the pressure state in each zone with the usual range of water content. In well drained soils, pressure at or near the surface is usually less than atmospheric, i.e. under tension indicating unsaturated conditions. Plant roots contribute substantially to maintaining soil water tension in the ***rhizosphere***. The hydraulic gradient remains constant, more or less, through the ***transmission zone*** which is an extension of the unsaturated ***vadose zone*** from the top of Earth's land surface down to the water table. At the capillary fringe, weak

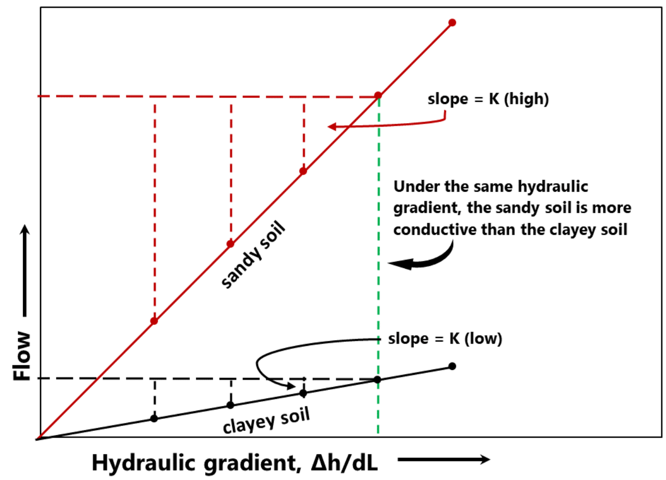


Figure 7.12.12 Illustration showing how the magnitude of K in Darcy's equation affects conductivity under the same hydraulic gradient. Darcy found that flow through porous media was proportional to the hydraulic gradient multiplied by a constant K. The proportionality constant K in this relationship is the hydraulic conductivity. *Source: adapted and redrawn from UDSA-NRCS.*

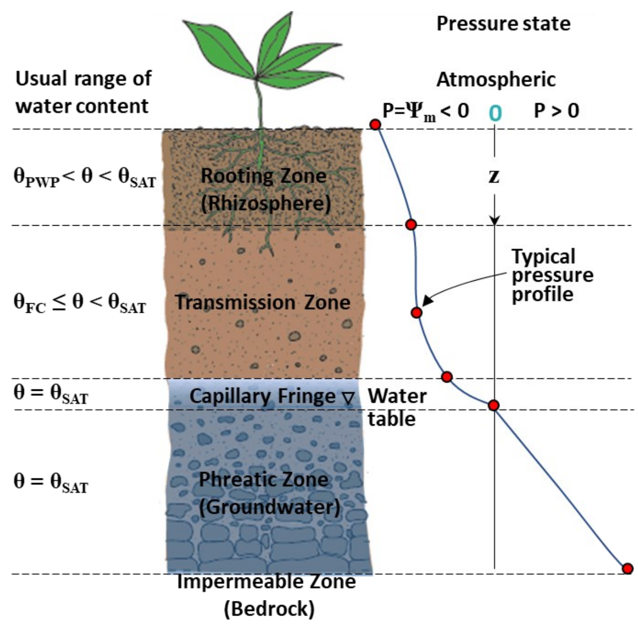


Figure 7.12.13 Soil profile divided into idealized hydrologic zones showing pressure state and usual range of water content. Water enters the root zone via infiltration and exits by evapotranspiration and gravity drainage. Key: Ψ_m = matric potential; z = depth; red dots = typical measured pressure state; θ = volumetric water content; θ_{PWP} = water content at the permanent wilting point; θ_{SAT} = water content at saturation. *Source: after Dingman 2002.*

matric forces dominate even though this zone is fully saturated ('tension saturated zone'). Atmospheric pressure prevails at the ground water surface. Below the ground water surface, in the ***phreatic zone***, hydrostatic pressures above atmospheric develop according to depth. In contrast, the usual range of soil water content diminishes with depth down to the water table.

Now, to put Darcy's law in practice, consider this simple use case.

Example 1: If we put a colored dye in well A in **Figure 7.12.11**, how long would it take before showing up in well B? The hydraulic gradient is -0.1 as noted. Assume uniform K_s of 27.6 mm/hr, equal to the measured 15 cm saturated hydraulic conductivity under disk tillage shown in Figure 7.12.9. We also determined the 15 cm effective[1] porosity was 38%.

Solution: Since a fluid velocity is needed, and movement is down in the soil without an explicit cross-sectional area, we want to solve for the flux per Eq.[2]:

$$q = -Ki$$

$$q = -K\frac{\Delta h}{\Delta L}$$

$$q = \frac{2.76\times 10^{-2} m}{hr}\times 0.1$$

$$q = \frac{2.76\times 10^{-3} m}{hr}$$

After one day, the dye will have moved a distance

$$\frac{2.76\times 10^{-3}\, m}{hr}\times\frac{24\, hr}{day} = \frac{6.62\times 10^{-2}\, m}{day}$$

or about 66 mm per day. The total time taken to move a distance of 100 m from point A to point B is:

$$100\, m\times\frac{day}{6.62\times 10^{-2}\, m} = 1,511\ days$$

Note that although the volumetric flux q is a velocity, it does not equal the true fluid velocity. The average fluid velocity v is found by dividing the volumetric flux by porosity:

$$v = \frac{q}{V_F}$$

where V_p fractional volume of area A represented by the pores through which flow is possible. The average fluid velocity is therefore:

$$v = \frac{\frac{2.76\times 10^{-3} m}{hr}}{0.38}$$

$$v = \frac{7.26\times 10^{-3} m}{hr}$$

or 574 days (1.6 years rounded).

[1] Pore space that contributes to flow, here assumed = 100% of the measured porosity.

Here it can be seen that dividing the flux q by % porosity increases the velocity. This is because as the area through which flow is passing decreases, the velocity increases. It's comparable to increasing the stream velocity by putting your finger on the end of a garden hose.

Even so, movement under these conditions would be slow compared to, say, surface flow. However, the situation would be different if, under the same conditions, harmful dissolved substances were introduced at point A and point B represented a stream only 10 meters away. Then, the time predicted for these substances to show up there would be about 57 days. Thereon, distribution in surface waters would be very rapid.

Darcy's law is also useful in studying leaching potential in the effective root zone under different cropping systems. This is important because nutrients that descend beneath this zone may enter shallow groundwater or worse, become pollutants. This is an active area of research that is likely to gain even more priority for sustainability against the rising cost of mineral fertilizers.

Down in the soil, underground fissures, earthworm channels and the like can increase flow rate and velocity ("preferential flow") or decrease by forcing through layers of lower hydraulic conductivity. Soil layering and two- and three-dimensional flow are more complex topics beyond the scope of this technical note but essential features to accurately describe water redistribution in the soil. Software programs like HYDRUS 2D/3D facilitate description of multi-dimensional soil hydraulic properties, heat, and solute movement that would otherwise be intractable by hand calculation.

In summary, Darcy's law embodies several important properties about water flow and redistribution through porous media, including:

- If no hydraulic gradient exists over a distance, no flow occurs. Hydrostatic conditions prevail.
- If there is a hydraulic gradient, flow will occur from high to low potential energy, i.e. down the hydraulic gradient in the direction opposite of increasing gradient. The right side of Darcy's equation carries a negative sign indicating the direction of flow.
- Driving force is proportional to the hydraulic gradient through the same porous media.
- Flow rate can speed up or slow down depending on the intrinsic permeability of porous media, even under the same hydraulic gradient in both cases.
- Hydraulic conductivity can be seen as the inverse of resistance driving flow across a hydraulic gradient, i.e. higher values of K increase flow whereas lower values of K decrease it.

Darcy's law has some limitations. One is that it applies only to laminar flow in porous media. This means that water is flowing roughly in parallel layers with no disturbance between the layers. Because the velocity of water in a saturated porous media like soil is relatively slow, it's assumed to be laminar. However, this may not always be the case, particularly in gravelly soil with large interparticle spaces where preferential flow occurs. And it certainly isn't the case in surface waters (see Manning equation).

Another assumption of Darcy's law is that flow through porous media continues in a steady state. This is embodied by K_s, the saturated hydraulic conductivity. In reality, where free drainage exists, the ground above the water table in the root zone does not stay saturated for long. Water movement takes place anyway in unsaturated soil but the environment in which it happens is more complex. What are the forces driving water movement where Darcy's law does not apply?

To answer this, consider the wetting patterns in a sandy loam and clay loam soil after irrigation shown in **Figure 7.12.14.** It can be seen that water movement is a combination of vertical and horizontal flow. It can also be seen that vertical movement is more rapid and pronounced in the sandy loam soil compared to the clay loam soil. As water descends in the profile, the largest pores empty first leaving the finer pores to assist movement. Descending motion is driven by the force of gravity when the soil is saturated until the forces of gravity and capillarity equalize. Then, soil water content is at field capacity.

The principle force driving water motion in soil drier than field capacity, i.e. unsaturated and thus non-Darcian, is the difference in water potential. This time, however, differences in the *matric potential gradient*, not hydraulic gradient, are driving the motion. Soil water in a region of high matric potential, e.g. –10 kPa will flow spontaneously to a nearby drier region of low matric potential, e.g. -100 kPa. Because the unsaturated region in the clay loam soil has more finer pores, water tends to move horizontally to a greater extent than in the sandy loam soil albeit more slowly. Some water movement in **Figure 7.12.14** is likely saturated flow assisted by the force of gravity and thus, Darcian in nature. Plant roots growing in the vicinity may also lower the matric potential, pulling water towards the root tips. The migration of water from the bulk soil into the plant root is discussed in TN 7 Part II Section 7.6.

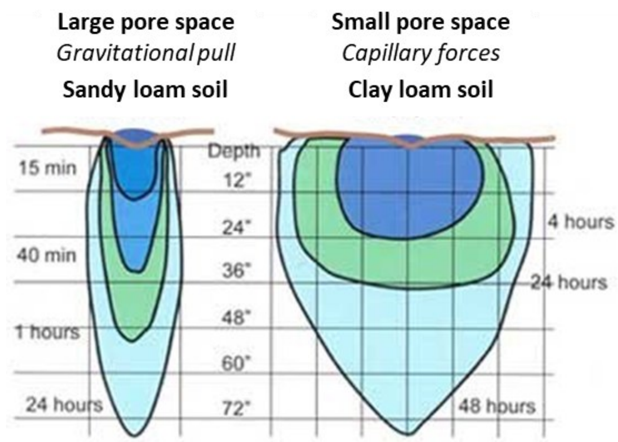


Figure 7.12.14 Comparison of wetting patterns in a sandy loam and clay loam soil after irrigation. *Source: Colorado State Univ.* link

7.13 Measuring Infiltration and Runoff

A simple method for measuring infiltration consists of one or two metal rings driven partially into the soil (**Figure 7.13.1 A** and **B**). For the single or double ring devices, there are two methods used for measuring infiltration rate: *constant head* and *falling head* methods. In the constant head method, water is ponded on the surface and the level in the ring maintained at a fixed level and the amount of water used to maintain this level is measured. The water level can be maintained manually or automated via float valve or ***Mariotte siphon*** device. In the falling head method, the time it takes ponded water in the ring to decrease a certain amount is recorded at intervals from the start of the test, until a constant rate is detected.

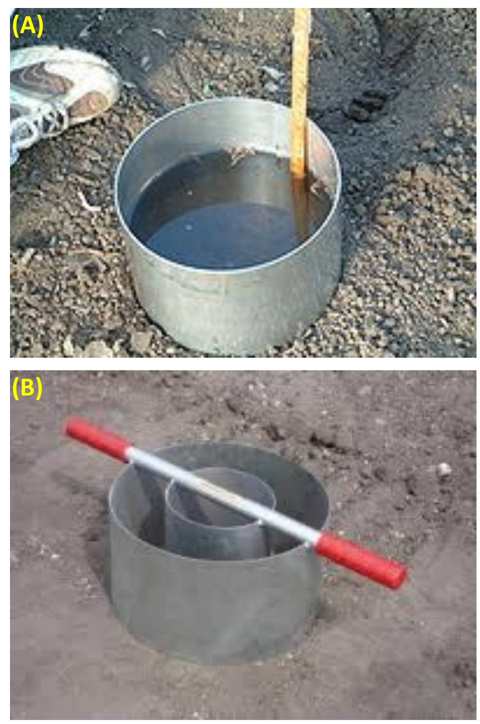


Figure 7.13.1 Single ring (**A**) and double ring (**B**) infiltrometers. Ring infiltrometers are typically constructed from narrow gauge sheet metal to facilitate pressing in the ground. The infiltrometers pictured above are the "falling head" type.

Another common device for measuring infiltration is the tension or disc infiltrometer (**Figure 7.13.2**). The device pulls water into the soil through a porous disc on the soil surface. Measurements of the amount of water passing through the disc are recorded at intervals in a manner similar to that of ring infiltrometers. Computations of water loss are then used to predict infiltration. Tension infiltrometers like the one pictured in **Figure 7.13.2** can be adjusted to pass water at different suction values, a clever design allowing the investigator to create infiltration models based on the distribution of pore size classes in the soil. Tension infiltrometers can also measure hydraulic conductivity but in this case, it's based on the applied tension and therefore denoted *unsaturated hydraulic conductivity*, K_Ψ. Because K_Ψ is not a constant value, but rather depends on matric potential (or conversely, water content), it can be seen as a continuous function similar to the conductivity curve depicted in **Figure 7.12.8**.

Rainfall simulators are widely used for modeling infiltration and surface runoff (**Figures 7.13.3A** and **B**). They are elaborate devices requiring significant resources to deploy and operate. However, rainfall simulators are better at capturing spatial variability than ponded ring or tension infiltrometers. Generally, rainfall simulators are deployed in situations where there's a need to augment or validate ponded infiltrometer tests or monitor water balance e.g. infiltration vs. runoff, particularly in environmental studies.

Two variables are usually computed when performing an infiltration test:

- Steady-state infiltration rate, depth H_2O per unit time
- Cumulative infiltration, depth H_2O per unit precipitation rate

Steady-state infiltration rate is needed to determine the maximum intake rate of the soil where water does not pond on the surface (and potentially runoff). Typically, infiltration rate is initially high and then decreases over time until a steady state is reached, as depicted in **Figure 7.12.3A**. The steady-state intake rate + soil surface storage (e.g. micro-depressions associated with surface roughness) is then taken as the maximum application rate for sprinkler irrigation systems, e.g. center pivot, linear move, solid set, landscape, etc. Typical saturated conductivity (steady state) infiltration rates are given in **Table 7.12.1.** Design of surface irrigation systems i.e. where water is purposely ponded on the soil surface (border, furrow, basin) also relies on a knowledge of steady-state infiltration rate and cumulative infiltration (wetted) depth parameters.

Data collected from field infiltration tests can be analyzed in several ways depending on the objectives of the investigator. Often, the data are fitted to empirical mathematical models describing infiltration over time as well as predicting steady-state infiltration and wetted depth. One popular model is the *Kostiakov equation*:

$$i = kt^a \qquad \text{[Eq. 3]}$$

where

i=depth of infiltration, mm

t=time, hr

k, a = constants

Another model, *Philip's equation*, is basically the same as the Kostiakov equation except the parameter a=0.5 and k is the soil water sorptivity. The *Horton equation* is a three-parameter model with some advantages over the Kostiakov equation, e.g. the final infiltration rate is a nonzero constant value. The Kostiakov equation can also be modified by

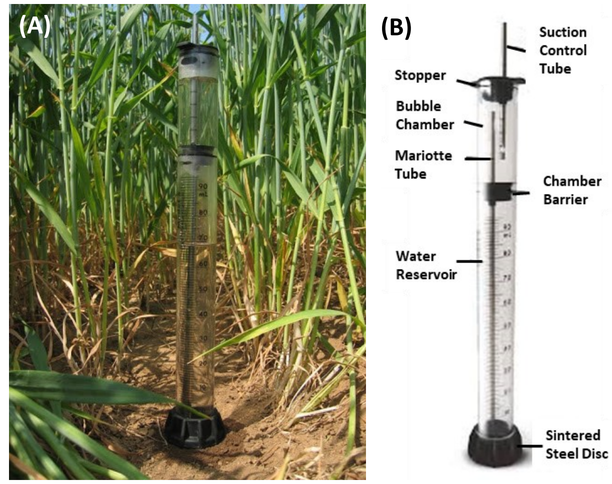
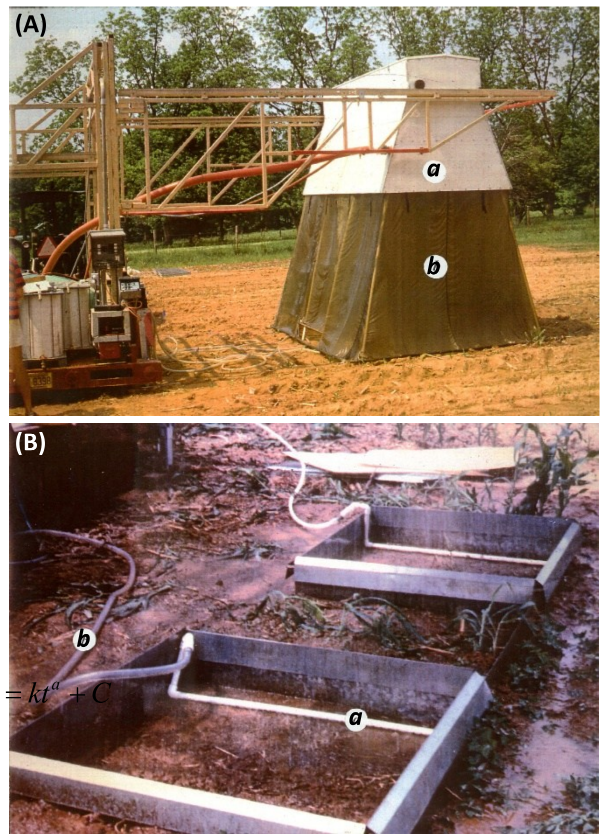


Figure 7.13.2 Tension infiltrometer (**A**) showing typical field installation; and (**B**) labeled device. *Image source: (A) hydropedologie.agrobiologie.cz; (B) Meter Group.*



$$i = kt^a + C$$

Figure 7.13.3 Variable-rate sprinkling infiltrometer. (**A**) Typical field set up with sprinkler hood (***a***) suspended on tracked reach arms with wind protection skirt (***b***) attached; and (**B**) steel frames with perforated PCV pick-up pipe (***a***) and suction hose (***b***) leading to collection tanks. *Image source: Walters NCSU.*

adding a constant value term

$$i = kt^a + C \qquad \text{[Eq. 4]}$$

where C can represent the final steady state infiltration rate reached at saturation (Mezencev, 1948) or, the initial infiltration rate into surface features like cracks and biopores (NRCS, 2012). Analytical solutions like the *Green-Ampt equation* are more complex and depend on measured (or estimated) soil properties.

The Kostiakov and modified Kostiakov equations are common models quantifying infiltration for irrigation design because they are more generalized and calculating the constants k and a is relatively straightforward.

Let's walk through one use case for the Kostiakov equation.

In this exercise, we'll use infiltration measurements collected from a ponded ring infiltrometer study in the same North Carolina Wedowee sandy loam soil under conventional plow + disk tillage described above in Section 7.12. The data are given in **Table 7.13.1**.

Table 7.13.1 Infiltration data.

Time (min)	Infiltrated depth (cm)
0	0
5	1.1
10	1.8
15	2.4
30	4.08
60	5.7
120	8.55
240	14.35

Readers trained in mathematics will see right away that the Kostiakov equation is a generalized power function $y = x^n$ with exponential properties. Thus, the constants a and k can be computed by taking the natural logarithm of each pair of observations in **Table 7.13.1** and plotting the result, which should yield something like **Figure 7.13.4.**

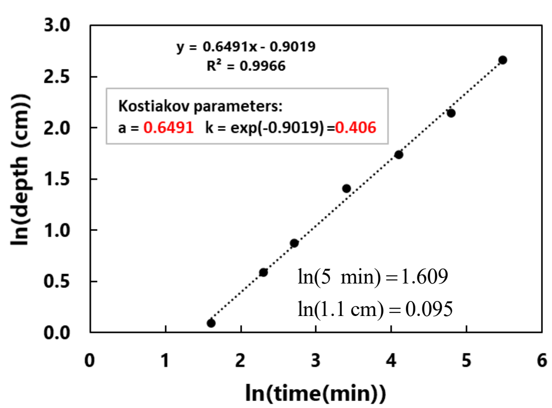


Figure 7.13.4 Linear regression of infiltration data.

The logarithm of infiltration rates is usually linear[2] and can be fit to the equation $y = mx + b$ as shown above.

[2] By the rule: log (AB) = logA + LogB

Spreadsheets like Excel make this very easy to do. For brevity, hereon we'll show a single sample calculation in each step of the Kostiakov procedure, but the reader should know that modeling infiltration from raw data involves lots of tedious, repetitive calculations best left to computer algorithms.

It can be seen in **Figure 7.13.4** that the slope, 0.6491, is the parameter "a" in the Kostiakov equation and the parameter "k" is computed by taking the inverse log (antilog) of the intercept (-0.9019). The depth of infiltration at 60 minutes is predicted as:

$$i = 0.406 \times 60^{0.6491}$$
$$i = 5.79\ cm$$

The depth of infiltration, i, agrees with the 60-minute infiltration shown in **Table 7.13.1**. Depth of infiltration is an important consideration in irrigation design because we want to place water in the prime or ***effective root zone*** where plants can easily find it. Shallow wetting is undesirable because it encourages surface evaporation whereas, water that moves below the effective root zone is mostly wasted. **Figure 7.13.5** shows the cumulative infiltration depth from beginning (t = 0) to finish (t = 240) in a Wedowee sandy loam soil. It can be seen that infiltration depth increased rapidly in the first 50 minutes but thereafter the wetted depth increased at a relatively steady rate. The final predicted wetted depth is 14.2 cm. This would be considered a good "soaking" depth of irrigation.

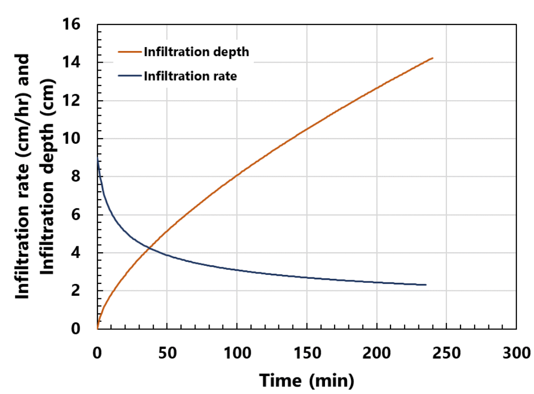


Figure 7.13.5 Kostiakov infiltration depth and rate for a North Carolina Wedowee sandy loam soil.

Next, we'll calculate the infiltration rate. The *instantaneous* infiltration rate is defined as the infiltration rate at any moment in time. The instantaneous Kostiakov infiltration rate is found through a bit of calculus sleight of hand, i.e. taking the derivative of Eq. [3] or [4]:

$$\frac{di}{dt} = ak(t)^{(a-1)}$$

where $\frac{di}{dt}$ = infiltration rate, cm/hr

t=time, minutes

Infiltration rate at 60 minutes is:

$$\frac{di}{dt}(cm/hr) = 0.264 \times 60^{(-0.3509)} \times 60$$

$$\frac{di}{dt} = \frac{3.76\ cm}{hr}$$

This infiltration rate is not the steady-state infiltration rate, however. The Kostiakov infiltration rate in Figure 7.13.5 indicates that intake rate equals 2.31 cm/hr after 240 minutes.

This is within range of a sandy loam soil per **Table 7.13.1**. This also looks to be the steady-state infiltration rate for this soil. If you plotted the curve out to 1,000 minutes the infiltration rate would be less than 2.31 cm/hr. This is an artifact of the Kostiakov equation (and certain power functions generally), not necessarily the soil behavior. Where free drainage exists, the soil approaches steady-state infiltration after about 60 minutes. Change in infiltration rate may be detected after 60 minutes but generally the magnitude is small. The Kostiakov equation also suggests that irrigation application rate should not exceed 2.31 cm/hr (~0.9 inch) to avoid runoff. Even so, this application rate may have to be scaled back to avoid runoff in trafficked lanes where there's compaction.

When water is infiltrating the soil, there is zero runoff. Runoff is the opposite of infiltration: it is water that does not enter the soil. Runoff occurs whenever precipitation exceeds the soil's infiltration capacity, or where a physical barrier prevents water from entering. Flooding from runoff is a growing concern worldwide. Urban development is covering more of the earth's surface with impermeable asphalt. Construction activities often compact and seal off soil surfaces to avoid water entry that would destabilize road and building subgrades, foundations, and other permanent structures (**Figure 7.13.6**). In agriculture, runoff is a serious threat to soil preservation and function, particularly soil surface sealing and crusting (**Figure 7.13.7**). Runoff exposes the soil to erosion, and sediment-laden flood water may contain dissolved chemicals (pesticides, fertilizer, etc.) that pose environmental hazards.

Variable intensity rainfall simulators that can simultaneously measure infiltration and surface runoff are invaluable tools for evaluating the water balance under different land management systems. **Figure 7.13.8** shows cumulative runoff collected at two different times from soybean plots under conventional plow and disk tillage at Reidsville, North Carolina. It can be noticed that runoff collected at full canopy was significantly greater than at early post-emergence, despite the initially lower soil water content at full canopy. This dramatically illustrates the extent of aggregate breakdown and crusting at the soil surface attributed to plowing and disking activities before planting. The presence of a living crop canopy is beneficial but no guarantee of runoff-free conditions as this study revealed.

Figure 7.13.6 Sheepsfoot roller compacting subgrade fill during road construction. *Image source: pavementinteractive.org*

Figure 7.13.7 Soil crusting increases runoff and delays or prevents seedling emergence. Tillage operations like plowing and disking contribute to the breakdown of aggregate structure at the soil surface. This leaves a hard, brittle, relatively impermeable surface crust. Water that does not infiltrate cannot restore the soil water deficit nor alleviate drought stress in crops. *Image source: A.J. Meijer NCSU.*

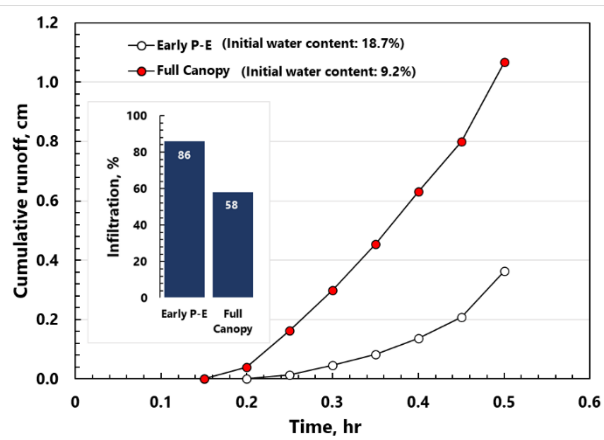


Figure 7.13.8 Cumulative runoff and percent infiltration collected over 30 minutes from a soybean field at early post-emergence (Early P-E) and full canopy growth stage on a Wedowee sandy loam soil under conventional disk tillage at Reidsville, N.C. In this study, a variable intensity rainfall simulator was calibrated to deliver 5.1 cm (2 inches) of rainfall per hour, typical of frequent high-intensity seasonal storms in the region. *Source: Walters NCSU*

Core Intelligence

Cumulative infiltration: Describes the accumulated depth, i.e. a length, of infiltrated water over a given period of time (t). Units are centimeters or equivalent length. Infiltration depth is typically larger than depth of applied water because it is spread out over a much larger area in the soil due to intermingling with solid particles.

Diffusivity: A hydraulic property relating the motion of water to relative differences in water content ('wetness" or "degree of saturation") rather than matric potential. In technical terms, diffusivity is equal to the product of the hydraulic conductivity and the slope of the water retention curve at a particular water content:

$$D(\theta) = K(\theta) \times \frac{d\psi}{d\theta}$$

where

$D(\theta)$ is the diffusivity or hydraulic diffusivity $D_h(\theta)$ with dimensions L^2 ($L \times L$) per unit time, t

$K(\theta)$ is the hydraulic conductivity, L per unit time

$\frac{d\Psi}{d\theta}$ = slope of the water retention curve at a particular water content with dimension L (centimeters or equivalent).

The motion of water in the soil described by diffusivity is one of mass flow (aka convection) under unsaturated conditions, not the classical notion of diffusion enshrined in Fick's Law.

Diffusivity is directly proportional to hydraulic conductivity and related non-linearly to volumetric soil water content.

Drainage: Describes the movement of gravitational water into open tile, channels, and ditches. The quantity of water that can be exhausted from the soil profile per unit time is called the *drainage coefficient* and is key to proper design of drainage systems. The *drainable porosity* is the pore space left after gravitational water has been exhausted from the soil profile. Numerically it is equal to the difference in volumetric water content at saturation (0 kPa) and field capacity.

Effective root zone: The upper portion of the root zone where plants get most of their water. Effective rooting depth is estimated as one-half (50%) of the maximum rooting depth. See "effective root depth" in TN 7 Part III Section 7.11.

Empirical: Based on observed measures or values from experimentation, not theory. An empirical model describes a type of object or system not attributing to knowledge of its composition or internal mechanisms. The parameters of an empirical model may have no physical meaning even though correlated with properties exhibited by an object or system. A *theoretical* model describes a type of object or system by attributing to it knowledge of its composition or internal mechanisms, such that reference to its behavior can be explained by that knowledge alone. Empirical vs. theoretical is an important distinction in probability, statistics, physics, and other areas of inquiry.

Field capacity: The amount of water remaining in the soil after free drainage. Water entering the soil from irrigation or natural precipitation initially moves downward due to the pull of gravity. The point at which drainage ceases (or becomes very small) is determined by soil particle shape and the packing density of the particles. Water remaining in the soil after free drainage is then held by capillary forces (adhesion and the surface tension of water molecules) and represents its water content at 'field capacity'. A sandy loam soil reaches field capacity when matric potential is near -10 kPa. Medium to fine-textured soils reach field capacity near –33 kPa matric potential. Field capacity is mainly used to infer soil physical attributes like workability or available water capacity as related to water content Field capacity is reached in most agricultural soils within 24-48 hours after wetting providing there are no restrictive layers.

Flux: This is the amount of "something" passing through an area (surface) per unit time. This "something" could be water, electromagnetic radiation, heat, molecules, or what have you. Because we're concerned with water in this technical note, flux is expressed as a volume, or volumetric flux. Flux could also refer to mass (molecules of a substance) or electromagnetic energy (photons absorbed, transmitted, or reflected by a surface). For Darcy's Law, the force driving water through a unit area is the hydraulic gradient (difference). The hydraulic gradient in a flux can be compared to a garden hose shooting water through an open window: the rate of flow per unit time, or volumetric flux, through the window area can be adjusted by opening or closing the hose valve. It can also be observed that the velocity of flow will increase as the area through which it passes decreases, i.e. by slowly closing off the hose valve.

Hydraulic conductivity: General term describing the ability of porous media to transmit water through pores or voids. Hydraulic conductivity, denoted K in Darcy's equation. is formally defined as

$$K = \frac{q}{\dfrac{\Delta h}{\Delta L}}$$

which describes the *flux* of water (or other substance) through a porous media per unit head gradient. The hydraulic conductivity of porous media is controlled by its permeability, which is governed, in turn, by pore size and connectivity. Soil is, by definition, a porous media by virtue of its ability to transmit water through pore channels or voids, a fraction of which are always present in any soil. Generally, sands and sandy loam soils have high hydraulic conductivity, loam and clay loam are intermediate, and silty clay and clay soils have low hydraulic conductivity.

Hydraulic Gradient: In physics, an increase or decrease in the magnitude of a property (e.g. temperature, pressure, or concentration) observed in passing from one point or moment to another is called a *gradient*. A hydraulic gradient is a gradient in which water potential is increasing or decreasing, i.e., water potential is the observed property. When the hydraulic gradient between two points is zero, no motion is possible. Hydraulic gradients can be positive or negative depending on the location of the reference plane.

Infiltration: describes the entry of water into the soil body at the surface. Infiltration is substantially affected by soil physical properties like porosity and aggregate structure, and initial water content. Direction of infiltration is generally vertical, i.e. descending in the soil profile. The forces driving infiltration are: gravitational, matric, pressure potential, or a combination.

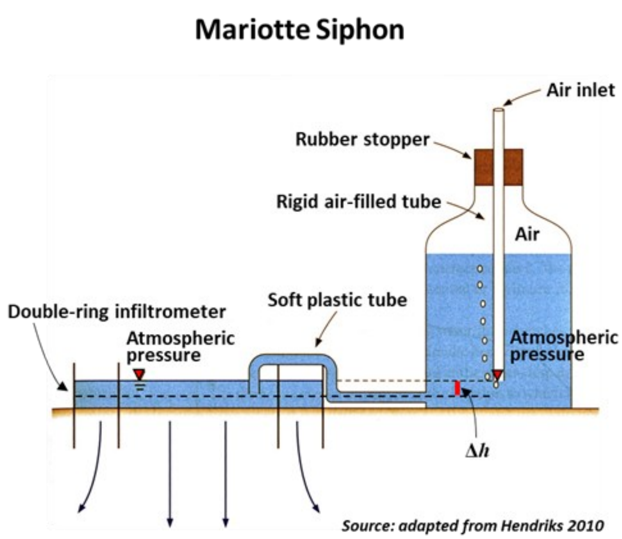
***Infiltrability**:* See infiltration capacity.

Infiltration capacity**:* Maximum rate at which infiltration can occur. Also called ***infiltrability. Units of measure are length (centimeters or equivalent) per unit time (seconds, hours, etc,). Infiltration capacity also describes in a given sense the hydraulic conductivity of the soil but the latter incorporates unit area in its calculation even though units for both terms, length per unit time, are identical. Flow via hydraulic conductivity can occur in any direction whereas flow via infiltration is generally downward.

***Infiltration rate**:* The rate at which water infiltrates the soil. Units of measure are length (centimeters or equivalent)) per unit time (seconds, hours, etc.). Normally, infiltration rate is high at the beginning of a rainfall or irrigation event and decreases over time as the soil gets wetter. Infiltration rate varies with soil physical properties and surface conditions (sealing, crusting, presence of biopores, etc.).

***Leaching**:* The downward flow of water transporting any soluble substances such as nutrients, chemicals, colloids, etc. in the unsaturated zone. The term *leachate* refers to any liquid that, in passing though porous media, extracts some soluble substances along the way.

***Mariotte siphon**:* A device for delivering water at a constant flow rate from a closed container or tank. Also called *Mariotte's bottle* or *Mariotte bubbler*, named after the French physicist Edme Mariotte (1620-1684) who discovered it. Curiously, the device lay dormant until described briefly in an article in Science magazine in 1934 by E.L. McCarthy. A typical use case for a Mariotte siphon is to maintain constant head in a double-ring infiltrometer, shown below.

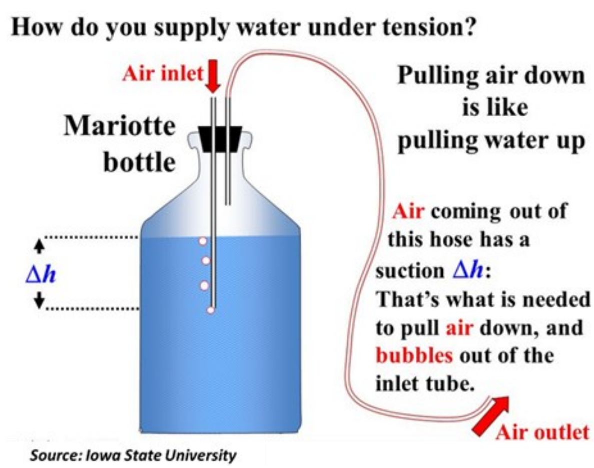


The principle behind Mariotte's device is based on the relative difference in height between the air inlet tube and the siphon tube. The difference, Δh, establishes the *hydraulic head* which determines the flow rate in the siphon tube. This can be seen in the above illustration where the upper horizontal dashed line marks the elevation at the bottom of the air inlet tube immersed in the reservoir of water, and also coincides with the height of the free water column in the inner infiltrometer ring. Because the bottom of the air inlet tube and the free water surface in the inner ring are both at atmospheric pressure, no flow occurs.

As water evacuates from the inner infiltrometer ring, Δh becomes non-zero and water flows out of the siphon to re-establish equilibrium. Water flowing out of the siphon pulls air into the reservoir, causing the air inlet tube to emit bubbles. Because the pressure of the water and air above the inlet tube is less than atmospheric, the bubbles move up to the top of the reservoir, and the rate of flow out of the siphon tube remains constant regardless of the water level in the reservoir above the air inlet tube. The pressures developed inside the Mariotte bottle are explained concisely in this video, albeit somewhat long-windedly.

The bubbling rate is proportional to rate of flow in the siphon, which is also proportional to Δh, indicated by the red bar in the above illustration. Therefore, a constant level, or "head" of water is maintained in the inner infiltrometer ring. Cumulative infiltration and infiltration rate are determined by measuring the length of water (centimeters or equivalent) exhausted from the supply bottle, and the time interval between successive measurements until bubbling reaches a slow, steady rate or ceases altogether (meaning contact with an impermeable layer).

The Mariotte siphon is also employed as a "constant head" permeameter for measuring hydraulic conductivity in the soil as explained in this laboratory exercise. A variant of Mariotte's siphon where the siphon tube is suspended in air inside the container under tension (less than atmospheric pressure) is the basis for measuring unsaturated hydraulic conductivity via tension or disk infiltrometer:



The tension of the air coming out of the hose is related to the hydraulic head, Δh. Hydraulic head, and therefore tension, are adjusted by increasing or decreasing the length of the air inlet tube. The actual design of tension infiltrometers is more elaborate than shown above but the operating principle remains the same.

Percolation: A general term for the downward flow of water in the unsaturated (vadose) zone. Deep percolation is sometimes used to describe the amount of water that drains vertically below the crop rooting zone, comprising a source of water that recharges the groundwater aquifer.

Permeability: In general, the property of a material that allows another substance (gas or liquid) to pass through it. In soil mechanics, permeability is understood as the velocity of water flow caused by a unit hydraulic gradient (difference). Permeability can refer to water flow in any direction and at any point whereas, infiltration describes the descending movement of water into the soil body at the surface. Permeability is symbolized by the lower-case letter k. Units are distance, or length (millimeters or equivalent) per unit time (minutes, hours, etc.).

Ponding: The accumulation of free water on the soil surface when the input rate exceeds the infiltration capacity. Ponding may also come from below if the groundwater level rises enough to saturate the soil. Ponding infiltrometers are devices that maintain a constant *ponding depth*, i.e. length of free water above the surface, during measurement.

Phreatic zone: The zone of saturation below the water table. Also generally referring in hydrology to the groundwater aquifer. The term 'phreatic surface' indicates the location where water pressure is under atmospheric conditions, i.e. the pressure head is zero. Usually this location coincides with the surface of the water table.

Redistribution: The movement of infiltrated water in the unsaturated zone of the soil. Redistribution can occur in any direction: toward the surface via evaporation from the upper layer of the soil; capillary rise into the unsaturated zone from the saturated zone; percolation from the unsaturated zone to the saturated zone; horizontal or downslope. Forces driving redistribution of water in the soil are gravitational, matric, pressure, or osmotic potential.

Rhizosphere: Generally referring to the soil and microbial communities found adjacent to the roots of living plants. Originally coined in 1904 by the German agronomist and plant physiologist Lorenz Hiltner, the term has been adopted by soil ecologist emphasizing the holistic view of the soil as a living body. In agronomics, the rhizosphere includes the surface, near-surface, and soil root zones.

Runoff: When rainfall intensity exceeds the capability of the soil to infiltrate (infiltration capacity), the fraction of water discharged away in overland flow is called runoff. Runoff generated on sloping land may be carried to nearby ditches, culverts, and waterways, causing water levels to rise. Flooding is a natural consequence of excessive runoff.

Run-on: Describes the relocation of runoff water within a field. Run-on water ponds in local depressions where it eventually infiltrates. This results in non-uniform distribution of water in the soil profile, and may result in differences in crop rooting, water availability, nutrient uptake, growth, and yield. Fields are typical "crowned" or graded in such a way to uniformly discharge runoff water to collection ditches.

Saturated hydraulic conductivity: The maximum rate at which water can be transmitted in the soil under saturated conditions. It is symbolized by K_s or sometimes K_{sat}. The saturated hydraulic conductivity is a constant value represented by steady-state flow conditions. However, the K_s value can vary substantially in the same field and in different soil layers. The opposite of saturated hydraulic conductivity is unsaturated hydraulic conductivity, symbolized by K_Ψ or K_h. The unsaturated hydraulic conductivity varies non-linearly with soil matric potential, Ψ_m. Both have dimensions of length per unit time.

Saturated flow: Saturated flow occurs when the soil pores are completely filled with water. Saturated flow occurs in aquifers (water-bearing sediments and rock strata), soil with limited drainage, seepage, or whenever soil water content rises above field capacity. Darcy's Law describes the movement of water in saturated porous media.

Saturated zone: That part of an aquifer below the water table, which is in direct contact with the subsoil or rock, and in which usually all pores and fractures are saturated with water. Above the water table is the unsaturated or vadose zone. Water in the saturated zone below the water table is usually above atmospheric pressure. In hydrology, the saturated zone is called the *phreatic* or *groundwater* zone. Its upper limit is usually denoted on maps by an inverted open triangle $\triangledown$.

Seepage: The water that drains below the irrigation canal or, under controlled drainage, ascends vertically via capillarity around the crop rooting zone or laterally via gravitational flow toward a drainage ditch.

Sorptivity: A physical process describing the ability of a medium to absorb or desorb liquid water, generally by capillarity (Philip, 1957). Sorptivity is the principal means driving the imbibition of water at atmospheric pressure into the soil matrix.

Transmissivity: This is related to hydraulic conductivity as:

$$T = Kb\left[\frac{L^2}{t}\right]$$

Where K is the hydraulic conductivity, i.e. the rate of flow under a unit hydraulic gradient, and b is the unit width of a layer. In hydrogeology, the layer may be an aquifer, confined or unconfined.

Transmission zone: An extension of the unsaturated zone ("vadose zone", qv.) beneath the rooting zone or rhizosphere, down to the capillary fringe of the ground water table where changes in water content are relatively small. Hence, the movement ("transmission") of water is this zone is largely driven by gravitational forces rather than differences in matric potential.

Unsaturated flow: Unsaturated flow occurs when the soil pores are not completely filled with water. Typically, this means that larger pores have emptied such that the contribution of hydraulic head (i.e. pressure head) or gravitational component to water flow becomes progressively less and the contribution of the matric (capillary) component increases. Soil water content is less than field capacity. The driving force for water flow in unsaturated porous media is the matric potential gradient, symbolized Ψ_m. The Richards equation approximates water flow in unsaturated soil:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[K\left(\Psi_m\right)\frac{\partial\Psi_m}{\partial z} + 1\right]$$

where

K is the hydraulic conductivity (q.v.),

Ψ_m the matric head arising from capillarity (or alternatively, volumetric water content θ),

z the elevation above a vertical datum,

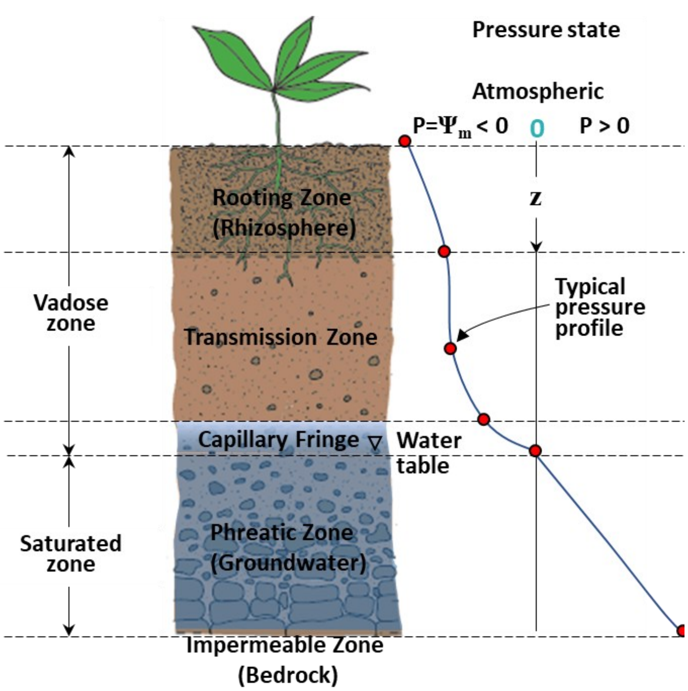
θ the volumetric water content, and

t is time.

The Richards equation represents a non-linear *partial differential equation*. In contrast, Darcy's Law is a linear relationship between water flow and hydraulic gradient which depends on the constant proportionality factor K_s, the saturated hydraulic conductivity. The Richards equation depends on the unsaturated hydraulic conductivity K_Ψ which is non-constant and related non-linearly to volumetric water content θ as depicted in **Figure 7.12.8**. The Richards equation is difficult to solve and involves considerable computational processing. We don't cover the Richards equation in this technical note. However, dirt hogs with lofty ambitions are encouraged to read more about it in Hillel (1994), Dingman (2002), and Hendriks (2010).

Unsaturated zone: Refers to the portion of the subsurface above the groundwater table. In hydrology, the unsaturated zone includes the capillary fringe, transmission zone, rhizosphere (rooting zone), and surface zone. Soil and rock in the unsaturated zone usually contain air as well as water in its pores. Another term for unsaturated zone is vadose zone. Water pressure in the unsaturated zone is usually less than atmospheric, i.e. under tension. Although the water holding capacity of Earth's unsaturated zone is enormous, water usually does not accumulate there. Thus, it holds only a minute fraction of earth's fresh water. It is, however, the zone mainly responsible for agricultural productivity.

Vadose zone: This describes the unsaturated part of Earth's terrestrial subsurface that extends from the top of the ground surface down to the water table. The vadose zone includes the surface soil, unsaturated subsurface materials, and the tension-saturated capillary fringe. Water in the vadose zone has a pressure less than atmospheric, i.e. it is usually under tension. The diagram below is a copy of **Figure 7.12.13** with unsaturated (vadose) and saturated zones delineated.



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