Predavanje 3-4-2018

9

Soil Physics

9.1 | Chapter Summary

Soils store a considerable amount of heat and water. The diurnal cycle of soil temperature and seasonal variation in soil temperature over the course of a year are important determinants of the land surface climate. The amount of water held in soil regulates evapotranspiration. This chapter reviews the physics of soil heat transfer and soil water relations. Heat flows from high to low temperature through conduction. Important soil properties that determine heat transfer are thermal conductivity and heat capacity. Two forces govern water movement in soil. Gravitational potential represents water movement due to the force of gravity. The second force, called matric potential, occurs because water is bound to soil particles. Water flows from high to low potential as described by Darcy's law. The Richards equation combines Darcy's law with principles of water conservation to describe the change in soil water content over time. Key hydraulic properties are porosity, matric potential, and hydraulic conductivity. These latter two properties vary with soil water. Soils differ in hydraulic properties in relation to the size and arrangement of pores. The pores in sandy soil are large, water loosely adheres to soil particles, water movement is rapid, and the soil drains rapidly. Pores are smaller in clay soil, water is tightly bound to soil particles, movement is slow, and drainage is impeded. Loams are intermediate, draining more slowly than sands and retaining more water.

9.2 | Soil Texture and Structure

Soils are composed of organic material, mineral particles, air, and water. A typical mineral soil is 55 percent solid particles and 45 percent air and water. Most soils have relatively low organic matter content, ranging from 1 percent to 10 percent. These soils are known as mineral soils. In contrast, organic soils are those in which more than 80 percent of the material is organic matter. These soils develop in swamps, bogs, and marshes where waterlogged conditions inhibit decomposition. The type, abundance, and arrangement of mineral and organic particles determine heat and water flow.

Mineral particles consist of three types determined by size. Sand particles are the largest, ranging in size from 0.05 mm (very fine sand) to 2 mm (very coarse sand). They are rounded or irregular, which creates large pore spaces between particles. Sand particles have a low capacity to hold water. Clay particles are the smallest mineral particles and are less than 0.002 mm in size. Clay particles are generally flat, plate-like, and fit closely together. They have the largest surface area, which facilitates the adsorption of water to the clay particles. In between, ranging in size from 0.002 to 0.05 mm, are silt particles.

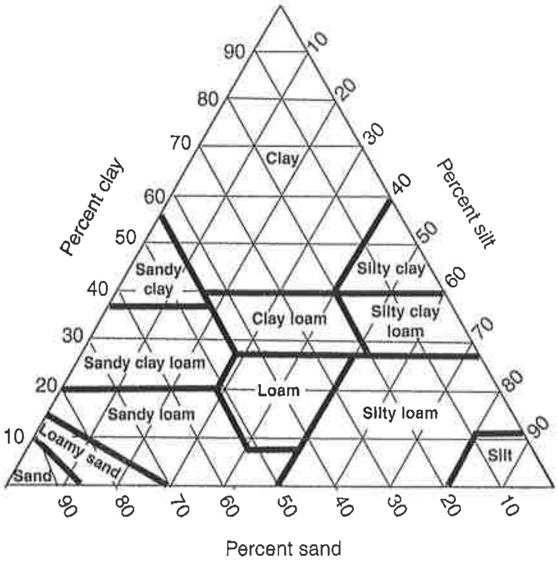

Fig. 9.1 Soil texture triangle showing the relationship between the 12 soil texture classes and percentage of sand, silt, and clay. The percentage of particles that are clay is determined by the lines running parallel to the sand side of the triangle. The percentage silt is determined by the lines running parallel to the clay side. The percentage sand lines run parallel to the silt side. For example, a soil that is 60 percent sand, 30 percent clay, and 10 percent silt is a sandy clay loam.

The relative abundance of sand, silt, and clay particles determines soil texture (Figure 9.1). There are three broad texture classes: sands - soils in which sand particles comprise more than 70 percent of the material by weight; clays - soils in which clay particles comprise at least 35–40 percent of the material; and loams - soils that are a mixture of sand, silt, and clay. A sandy soil has loose, individual grains that can be seen and felt. It readily falls apart when dry. If squeezed when dry, a loamy soil will form a molded shape that stays together with careful handling. When wet, the molded form is more durable and can be handled without breaking. A clay soil forms hard clumps when dry and is sticky when wet. Wet clay will form a long, flexible ribbon when squeezed.

Soil texture considers the size distribution of mineral particles. The arrangement of soil particles into large recognizable units, known as soil structure, is equally important. Loose, granular soil is much more porous than compacted soil. The development of good soil structure requires some agent that coheres and cements individual mineral particles together. Clay particles, because of their large surface area, are one such binding agent. Lack of clay is one reason why sandy soils crumble so easily. Organic matter is another cementing agent that allows individual particles to adhere to one another. Through this, organic matter has a great influence on the ability of soil to store water.

The sand, silt, clay, and organic particles that form soil lie in close contact, but because of their irregular shapes they do not fit evenly together. Instead, voids exist around the individual particles. The total volume of voids is known as pore space, or porosity. It is the volume of soil that is occupied by air and water. These pores can fill with water, such as happens when water infiltrates into the soil. Or when dry, the pores consist mainly of air. Most field conditions are in between, and the soil is a mix of solid particles, water, and air. In a typical soil, about 55 percent of the soil volume comprises solids and 45 percent of the soil volume is pore space comprising air and water.

9.3 | Soil Temperature

Soils are a large source or repository of heat that moderates the diurnal and seasonal range in surface temperature. During the day, when solar radiation heats the surface, the surface is warmer than the underlying soil and heat flows into the soil. This transfer of heat away from the surface cools the surface. At night, the surface is cooler than the soil and heat flows out of the soil. This gain of energy at the surface warms the surface. As a result, surface air temperature shows less of a diurnal range than if no heat were stored in the soil. The same behavior occurs annually, when soil stores heat in warm months and releases heat during cold months.

Figure 9.2a illustrates a typical summer temperature profile during the day and night. At night, temperatures increase with depth. Because heat flows from high to low temperatures, heat flows upward from deeper depths

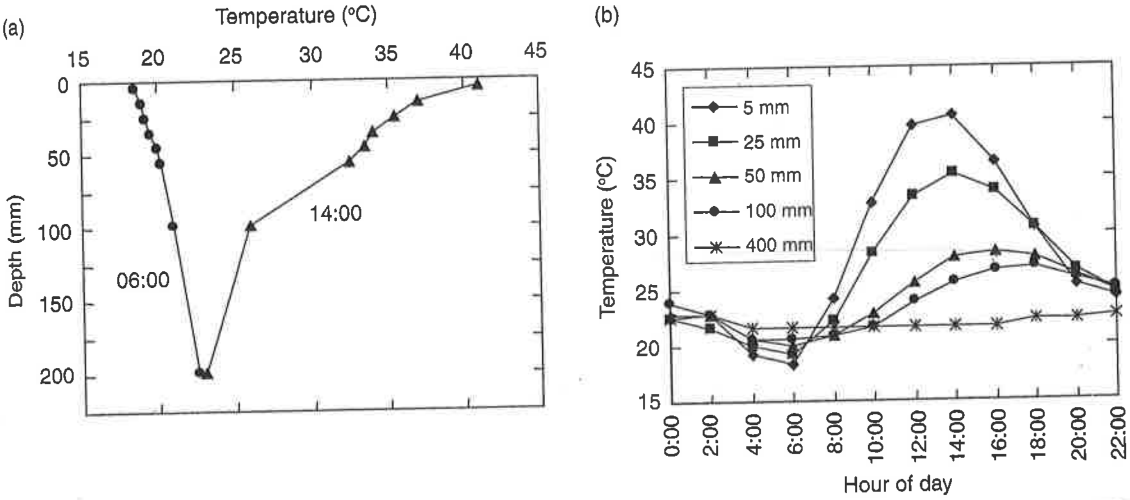

Fig. 9.2 Summer soil temperatures. (a) Typical night and day soil temperature profiles during summer. (b) Diurnal cycle of soil temperature at several depths on a typical summer day. Adapted from Hartmann (1994, p. 86).

to the surface. During the day, the soil profile warms, but the warming decreases with greater depth. The deep soil hardly warms at all from its nighttime temperature. As a result, daytime soil temperatures decrease with depth, and heat flows downward from the surface towards the deep soil. The diurnal cycle of soil temperature at different depths further illustrates this behavior (Figure 9.2b). The soil close to the surface, at a depth of 5 mm, warms rapidly as the Sun's radiation heats the surface, increasing from 18°C at 0600 hours to 41°C at 1400 hours. This upper soil also cools rapidly at night. Deeper soil layers are cooler than upper layers during the day (e.g., from 1200 to 1600 hours) and warmer at night (e.g., from 0400 to 0600 hours). The diurnal range in temperature decreases with depth, and maximum temperatures occur later in the day.

Two soil properties (thermal conductivity, heat capacity) determine the temperature profile for a given heat flux at the surface. First, heat flows from high temperature to low temperature. The rate at which heat flows between two points separated by a distance of Δz meters is equal to soil thermal conductivity times the temperature gradient:

$$F = -\kappa(\partial T / \partial z) \quad (9.1)$$

with F heat flux (W m^{-2}), κ thermal conductivity (W m^{-1} K^{-1}), and $\partial T / \partial z$ (K m^{-1}, or °C m^{-1}) the temperature gradient. This latter term is the change in temperature with depth in soil, approximated numerically by $\Delta T / \Delta z$. The negative sign denotes that the heat flux is positive for a negative temperature gradient; the heat flux out of the soil is positive, and the flux into the soil is negative. Thermal conductivity determines the heat flow in unit time by conduction through a unit thickness of a unit area of material across a unit temperature gradient.

Second, if more heat enters a volume of soil than exits, the soil gains heat and warms. Conversely, net loss of heat cools the soil volume. Heat capacity is a measure of the temperature change arising from this change in heat storage. It is the amount of heat required to change the temperature of a unit volume of material by 1°C. Energy conservation requires that the difference between heat coming into the top of a slab of soil at depth z and heat exiting the bottom of the slab at depth $z + \Delta z$ equal the rate of heat storage (Figure 9.3):

$$\rho c(\Delta T / \Delta t)\Delta z = -(F_z - F_{z+\Delta z}) \quad (9.2)$$

where ρc is heat capacity (J m^{-3} K^{-1}), $\Delta T / \Delta t$ is the change in temperature with time (K s^{-1},

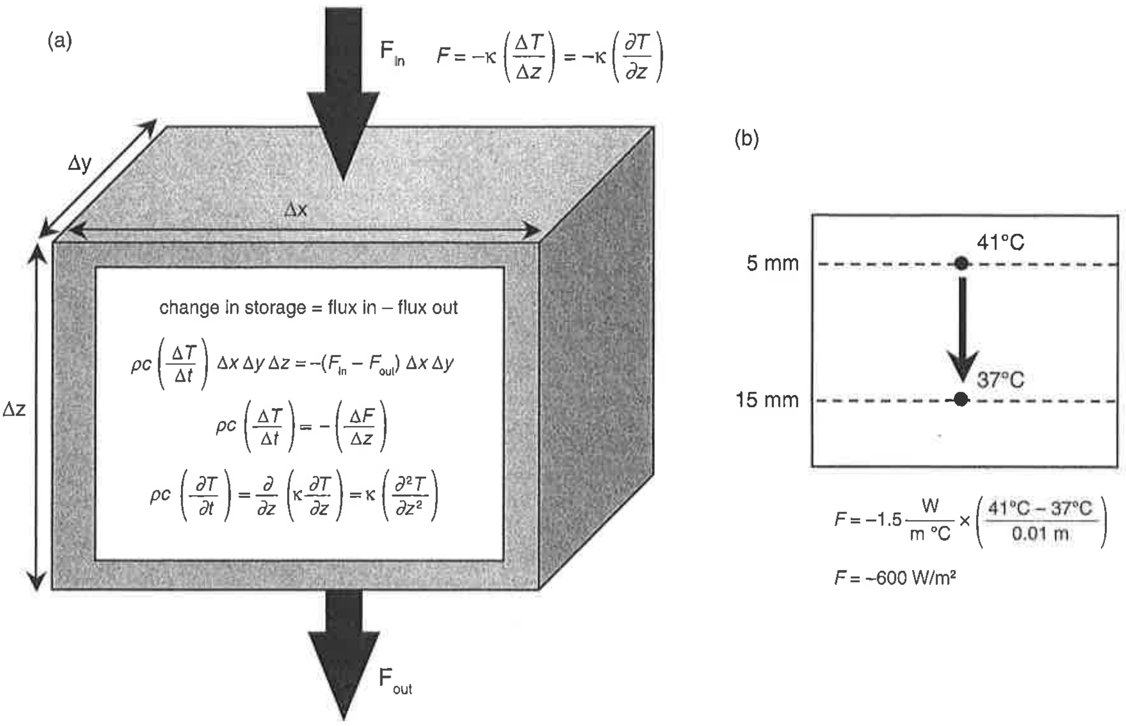

Fig. 9.3 Heat transfer in soil. (a) Heat balance of a volume of soil. Equations are shown in both their numerical finite difference form and in the notation of calculus. (b) Example of vertical heat transfer between two points with a temperature difference of 4°C separated by a distance of 10 mm with a thermal conductivity of 1.5 W m^{-1} K^{-1}. Hanks (1992) and Hillel (1998) review soil heat transfer.

or °C s^{-1}), and F_z and $F_{z+\Delta z}$ are the heat flux (W m^{-2}) into and out of the soil slab. The negative sign ensures that temperature increases when there is a net gain of heat.

Combining equations for heat flux, given by Eq. (9.1), and storage, given by Eq. (9.2), and assuming heat capacity and thermal conductivity do not change with depth gives the change in temperature over time (Figure 9.3), or in the notation of calculus:

$$\frac{\partial T}{\partial t} = \frac{1}{\rho c}\left[\frac{\partial}{\partial z}\left(\kappa\frac{\partial T}{\partial z}\right)\right] = \frac{\kappa}{\rho c}\left(\frac{\partial^2 T}{\partial z^2}\right) \tag{9.3}$$

The change in soil temperature over time is directly proportional to the thermal conductivity and inversely proportional to the heat capacity. Thermal conductivity determines the rate of heat transfer and heat capacity determines the temperature change as a result of this heat transfer. Soils with a high thermal conductivity gain and lose energy faster than soils with a low thermal conductivity. Soils with a low heat capacity warm and cool faster, for a given heat flux, than soils with a high heat capacity.

Thermal conductivity and heat capacity vary depending on mineral composition, porosity, organic matter content, and the water content of soils (Table 9.1). Soils consist of solid particles, air, and water. The overall thermal conductivity of a soil is a weighted average of the conductivity of its solid, air, and water fractions. Quartz has a very high thermal conductivity, and soils with high quartz content (e.g., sandy soils) have a high thermal conductivity. Clay minerals have a lower thermal conductivity, and clay soils have a lower thermal conductivity than sandy soils. Organic material has an extremely low thermal conductivity, and soils with high organic matter content have a thermal conductivity that is one-quarter to one-third that of mineral soils. Air and water

Table 9.1 | Thermal conductivity and heat capacity for soil components and for sand, clay, and peat soils in relation to soil water

	Thermal conductivity ($W\ m^{-1}\ K^{-1}$)	Heat capacity ($MJ\ m^{-3}\ K^{-1}$)
Soil component		
Quartz	8.80	2.13
Clay minerals	2.92	2.38
Organic matter	0.25	2.50
Water	0.57	4.18
Air	0.02	0.0012
Sandy soil (porosity = 0.4)		
0%	0.30	1.28
50%	1.80	2.12
100%	2.20	2.96
Clay soil (porosity = 0.4)		
0%	0.25	1.42
50%	1.18	2.25
100%	1.58	3.10
Peat soil (porosity = 0.8)		
0%	0.06	0.50
50%	0.29	2.18
100%	0.50	3.87

Note: Soil water content is expressed as a percentage of saturation.
Source: From Monteith and Unsworth (2013, p. 281).

occur in the voids, or pore space, around soil particles. Air and water have a lower thermal conductivity than mineral particles. Consequently, soils with a high pore space have a lower thermal conductivity, all other factors being equal, than soils that are less porous. Sandy soils are less porous than clay soils, which is another reason why they have a higher thermal conductivity. Organic soils are often extremely porous. Thermal conductivity of soil increases greatly with increasing soil water content because the thermal conductivity of water is more than 20 times that of air. Similarly, the heat capacity of water is 3500 times that of air, and the heat capacity of soil increases with water content.

The practical implications of these differences in thermal properties are clearer under idealized conditions. The diurnal and annual cycles of temperature at the soil surface can be represented as a sine wave in which surface temperature varies periodically between some maximum and minimum values. Mathematically, the temperature at the soil surface at some time t is:

$$T_s(t) = \overline{T}_s + A_s \sin(2\pi t / p) \tag{9.4}$$

where p is the period of oscillation (e.g., 86,400 seconds for a diurnal cycle), $\overline{T}_s$ is the average temperature over this period, and A_s is the amplitude (i.e., one-half the difference between maximum and minimum temperatures) over the same time period. This periodic behavior is seen for the near-surface temperature in Figure 9.2b, which has a minimum early in the morning, a maximum in early afternoon, and decreases again during the night.

With a periodic surface temperature and if thermal properties are constant with depth, temperature at a depth of z meters is:

$$T_z(t) = \overline{T}_s + A_s e^{-z/D} \sin\left(2\pi \frac{t}{p} - \frac{z}{D}\right) \tag{9.5}$$

where $D = \sqrt{\alpha p / \pi}$ and $\alpha = \kappa / \rho c$ is the thermal diffusivity ($m^2 s^{-1}$). The term $A_s e^{-z/D}$ describes the decrease in surface temperature amplitude with depth. At depth $z = D$ the amplitude is $0.37A_s$. This depth is called the damping depth. The amplitude at depth $z = 2D$ is $0.14A_s$ and at depth $z = 3D$ is $0.05A_s$. In other words, the temperature amplitude decreases with depth, exactly as seen in Figure 9.2b.

A typical soil diffusivity is $\alpha = 7 \times 10^{-7}$ $m^2 s^{-1}$. Over the course of a day (p = 86,400 seconds), the damping depth is 14 cm. That is, the diurnal range in temperature at a depth of 14 cm is 37 percent that at the surface. Over the course of a year (p = 86,400 × 365 seconds), the damping depth is 2.65 m. At a depth equal to three times the damping depth, the range in temperature is 5 percent that at the surface. So at a depth of 42 cm, the temperature is approximately equal to the average daily temperature. At a depth of about 8 m, the temperature is approximately equal to the average annual temperature.

The term z / D represents the shift in time with depth when maximum and minimum temperatures occur. For example, maximum temperature at the surface occurs at $t = 0.25p$, but maximum temperature at depth $z = D$ occurs at $t = 0.41p$. That is, over the course of a day the maximum temperature at the damping depth occurs almost 4 hours later than the maximum temperature at the surface. With deeper depths, maximum temperature occurs later. At depth $z = \pi D$, temperature is at a maximum when surface temperature is at a minimum. Again, this is exactly what is seen in Figure 9.2b.

Soils are often covered by organic material or snow. Forests, in particular, are typically covered with a layer of decomposing leaf litter several centimeters thick. Organic material has a heat capacity similar to mineral soil, but a much lower thermal conductivity (Table 9.1). As a result, decomposing organic material acts as an insulator, preventing soil from warming in the day and from cooling at night. Snow, with its low thermal conductivity (e.g., 0.34 W m^{-1} K^{-1}), has a similar insulating effect. A deep snow pack early in winter can keep soil warmer than if no snow was present.

In seasonally frozen soils, it is necessary to account for the different thermal properties of water and ice (Farouki 1981; Lunardini 1981). The heat capacity of ice (approximately 2 MJ m^{-3} K^{-1}) is one-half that of water, while its thermal conductivity (2.2 W m^{-1} K^{-1}) is almost four times that of water. Additionally, the change in phase of water consumes and releases heat. At 0°C, 334 joules are needed to melt one gram of water (Table 3.3). This energy (latent heat of fusion) changes the phase of water from ice to liquid rather than warming the soil. The same heat is released when water freezes. While water is freezing, its temperature remains at 0°C. The importance of phase change is seen in simulations of soil temperature with and without phase change. Without phase change, an unfrozen soil undergoing freezing cools too rapidly.

9.4 | Soil Water

The upper meter of soil typically holds 10–45 cm of water. One measure of soil moisture is the volumetric water content, which is defined as the volume of water per unit volume of soil (i.e., the fraction of the soil volume that is water). Figure 9.4 shows typical water contents for a loam soil. The maximum amount of water held in soil occurs when all the pore space is filled with water. The soil is a mixture of water and solid particles. For loam, pores comprise 45.1 percent of the volume of soil, and this is also the water content at saturation. In other words, 1 m^3 of loam holds 0.451 m^3 of water when saturated. Volumetric water content is also the depth of water per unit depth of soil; 1 m of loam holds 45.1 cm of water at saturation. This water is loosely held in the soil, and it quickly drains due to the force of gravity. The amount of water held in the soil after gravitational drainage is its field capacity. The soil is no longer saturated, but rather is a mixture of water, air, and solids. At field capacity, 39.3 percent of the soil volume (87.1% of the pore space) is water. As the soil dries, water movement becomes more difficult, and at some critical water content the water is so strongly bound to soil particles that it can no

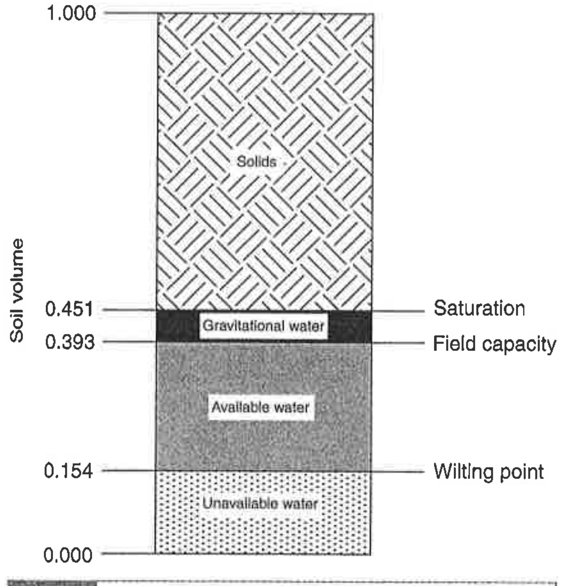

Fig. 9.4 Representative water contents for a loam soil.

longer be removed. This water content is known as the wilting point because a plant is likely to lose turgor and wilt if the soil is not replenished with water. The soil consists primarily of air and solid particles, and the pores are relatively devoid of water. At wilting point, 15.4 percent of the soil volume (more than one-third of the pore space) is water, but this water is unavailable to plants because it is tightly bound to the soil particles. The difference between the water content at field capacity and the water content at wilting point is the water available to support plant growth. The available water capacity is 23.9 percent, meaning that water can be extracted from only about one-quarter the volume of a loam soil. One meter of loam soil, therefore, only has at most 23.9 cm of available water. When dry, it still contains 15.4 cm of water, but plants cannot extract this water.

Two forces, or potentials, govern water movement in soil. Gravitational potential represents water movement due to the force of gravity. Gravitational potential is the elevation above some arbitrary reference height. A point one meter higher than another point has a gravitational potential that is 1000 mm greater, and because water flows from high potential to low potential water flows downhill to the lower point. For wet soil, gravity is the dominant force causing water to move.

The second force, called the matric potential, occurs because water is attracted to the solid surfaces of mineral and organic particles. This attraction creates a negative pressure, or suction, that binds water to the soil matrix. Matric potential depends on water content. For saturated soil, relatively weak matric forces are exerted on water. As the soil dries, its matric potential decreases and strong pressures bind water to the soil matrix (i.e., suction increases). Water movement is more difficult because of this binding force. This can be described mathematically (Clapp and Hornberger 1978):

$$\psi = \psi_{sat}\left(\theta / \theta_{sat}\right)^{-b} \tag{9.6}$$

where ψ is matric potential, ψ_{sat} is matric potential at saturation, θ is volumetric water content, θ_{sat} is water content at saturation (porosity), and b is an empirical parameter. Soils differ in hydraulic properties and matric potential (Table 9.2). The same amount of water is held more strongly in clay than in sand (Figure 9.5). van Genuchten (1980) provides another commonly used soil water retention function.

Field capacity typically occurs at a suction of 1000 mm. Because water is held more tightly by clay than by sand, field capacity occurs at higher water content for clay than for sand (Figure 9.5). The water content at which wilting occurs depends on plant physiology, but it generally occurs at a suction of 150,000 mm (Figure 9.5). Soil water contents at saturation, field capacity, and wilting point vary with texture (Table 9.2). At wilting point, clay holds more water than loam or sand and water fills more of the pore space. Sand and clay have low available water, and loam has high available water.

The sum of gravitational and matric potential governs water movement, and water flows from high to low potential. Except when wet, the adsorptive force binding water to the soil matrix exceeds the gravitational force pulling water downward. For example, consider a column of loam one meter deep. The difference in gravitational potential between water at the top of the column and water at the bottom is

Table 9.2 Hydraulic properties in relation to soil texture

Soil texture	Porosity, θ_{sat} (fraction)	Percentage of saturation			Hydraulic conductivity at saturation, K_{sat} (mm hr^{-1})	Matric potential at saturation, ψ_{sat} (mm)	Exponent
		Field capacity	Wilting point	Available water			b
Sand	0.395	59	17	42	634	−121	4.05
Sandy ilovi loam	0.435	73	26	47	125	−218	4.90
Loam	0.451	87	34	53	25	−478	5.39
Clay loam	0.476	95	52	43	9	−630	8.52
glina Clay	0.482	92	59	33	5	−405	11.4

Note: See also Cosby et al. (1984) for relationships with sand and clay content.
Source: From Clapp and Hornberger (1978).

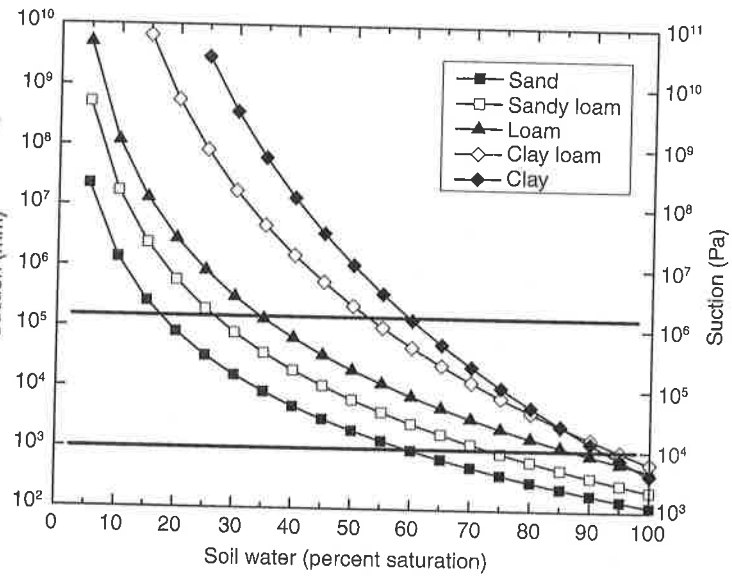

Fig. 9.5 Adsorptive forces binding water to soil particles as a function of water content for sand, sandy loam, loam, clay loam, and clay soils. Suction is a negative pressure (i.e., suction is the negative of matric potential). Suction and matric potential can be expressed in units of pressure (Pa = kg m^{-1} s^{-2}) or height of a column of water under that pressure (mm). These units are related by the density of water (ρ_w, 1000 kg m^{-3}) and the gravitational constant (g, 9.8 m s^{-2}) as 1 mm = (0.001 m) × $\rho_w g$ = 9.8 Pa. Field capacity generally occurs at a suction of 1000 mm. Wilting point typically occurs at a suction of 150,000 mm. Data from Clapp and Hornberger (1978). See also van Genuchten (1980) for another soil water retention function.

1000 mm. When the soil is more than 85 percent saturated, the force binding water to the soil matrix (Figure 9.5) is less than the gravitational force and the soil readily drains. At a water content less than about 85 percent of saturation, however, the matric suction exceeds the gravitation potential.

Water flow is governed by Darcy's law:

$$F = -K\left[\frac{\partial(\psi + z)}{\partial z}\right] \quad (9.7)$$

where F is water flux (mm s^{-1}), K is hydraulic conductivity (mm s^{-1}), ψ is matric potential (mm), and z is gravitational potential (mm), which is defined as the height above some arbitrary height. The term $(\psi + z)$ is the total potential, and the term $\partial(\psi + z)/\partial z$ is the change in total potential with depth, approximated numerically by $\Delta(\psi + z)/\Delta z$. The negative sign ensures that downward water flow is negative and upward water flow is positive. Darcy's law is analogous to soil heat transfer. Vertical water

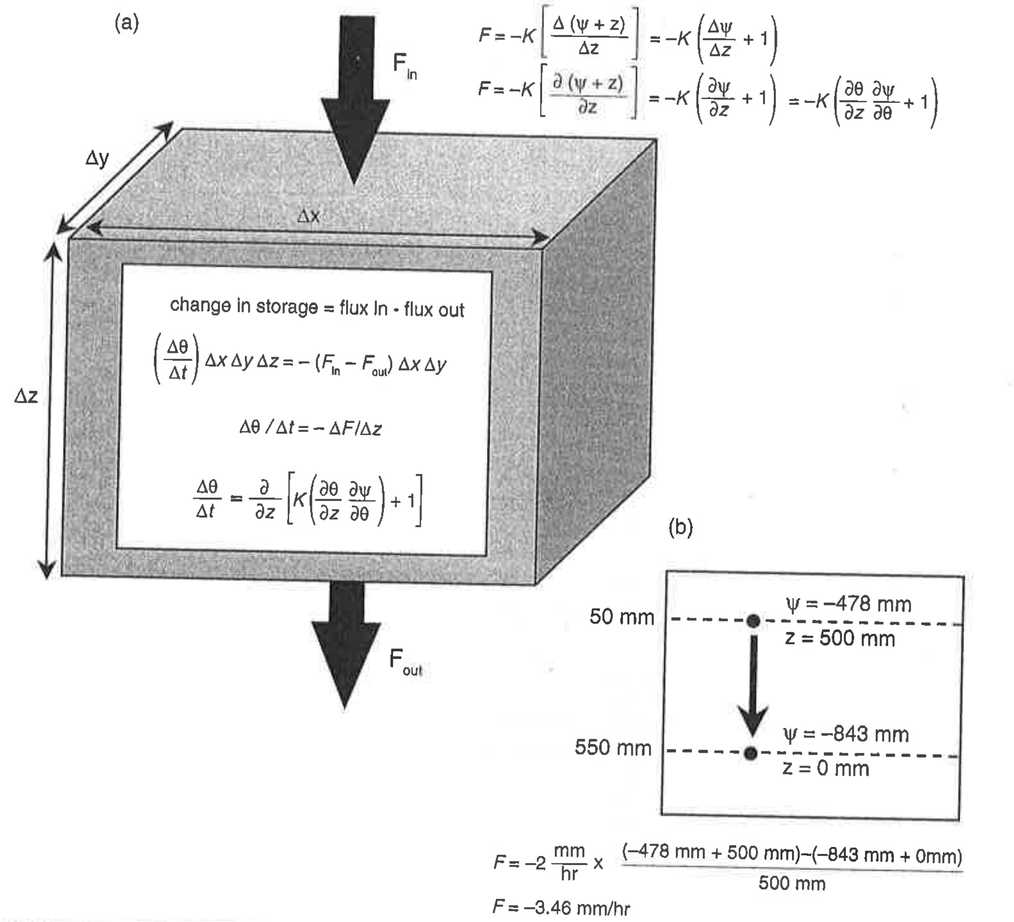

Fig. 9.6 Water flow in soil. (a) Water balance of a volume of soil. Equations are shown in both their numerical finite difference form and in the notation of calculus. (b) Example of vertical water flow between two points with a matric potential difference of 365 mm separated by a distance of 500 mm with a hydraulic conductivity of 2 mm hr^{-1}. Hanks (1992) and Hillel (1998) review unsaturated soil water flow.

flow is related to a hydraulic conductivity times a potential gradient, flowing from high to low potential.

The change in soil water over time is given by combining Darcy's law with principles of water conservation (Figure 9.6). The change in water storage in a volume of soil equals the difference between the flux of water into the volume and the flux of water out of the volume. The volume of water in a cube with the dimensions $\Delta x\Delta y\Delta z$ is $\theta\Delta x\Delta y\Delta z$. At the top of the cube, F_{in} mm s^{-1} of water flows into the soil across a cross-sectional area $\Delta x\Delta y$ mm^2. Likewise, $F_{out}\Delta x\Delta y$ mm^3 s^{-1} of water flows out at the bottom. Combining equations for water flux, given by Eq. (9.7), and change in storage gives the water balance $\Delta\theta / \Delta t = -\Delta F / \Delta z$, or in the notation of calculus:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[K\left(\frac{\partial\theta}{\partial z}\frac{\partial\psi}{\partial\theta}\right) + 1\right] \tag{9.8}$$

This equation, known as the Richards equation, relates the change in water content over time to hydraulic conductivity and matric potential. The left-hand side of the equation is the rate of change of water in a volume of soil and the right-hand side is the difference between inflow and outflow rates, each expressed on a per unit volume basis. The term $\partial\psi/\partial z = (\partial\theta/\partial z)(\partial\psi/\partial\theta)$

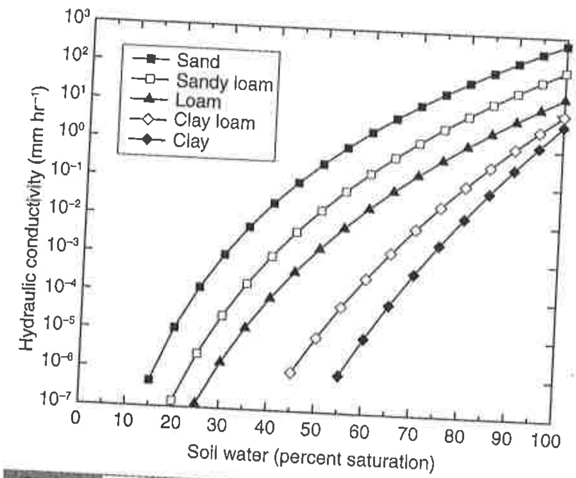

Fig. 9.7 Hydraulic conductivity in relation to soil water for sand, sandy loam, loam, clay loam, and clay soils. Data from Clapp and Hornberger (1978). See also van Genuchten (1980) for another function.

arises because water flux is given in terms of matric potential (ψ) whereas water storage is in terms of volumetric water content (θ). The relationship between ψ and θ provides common terms, but makes the Richards equation difficult to solve numerically because it is non-linear (Figure 9.5).

An additional complication is that hydraulic conductivity is a non-linear function of soil water. Hydraulic conductivity decreases as soil dries, described mathematically (Clapp and Hornberger 1978):

$$K = K_{sat}\left(\theta / \theta_{sat}\right)^{2b+3} \tag{9.9}$$

where K_{sat} is hydraulic conductivity at saturation (Figure 9.7). When saturated, a loam has a hydraulic conductivity of 25 mm hr^{-1}. At a water content of 90 percent of saturation, the hydraulic conductivity is only 6 mm hr^{-1}, and at 80 percent of saturation it is only 1 mm hr^{-1}. As soil dries, not only is water held more tightly to the solid particles (i.e., suction increases and matric potential becomes more negative), but the pore space filled with water becomes smaller and discontinuous. As a result, movement of water is hindered and hydraulic conductivity decreases. In general, sand has the highest hydraulic conductivity for a given water content, loam has moderate hydraulic conductivity, and clay has extremely low conductivity.

Table 9.2 compares hydraulic properties for several soil textures. In addition to total pore space, physical properties of soil such as hydraulic conductivity and the soil moisture retention curve determine how fast water drains in a soil, how much water it can hold at field capacity, and how much water is available to plants. Total pore space, expressed on a volumetric basis, ranges from 40 percent in sand to 48 percent in clay. Although porosity is low in sand, the individual pores are large because of the large mineral particles. Water loosely adheres to mineral particles (high matric potential when saturated), water movement is rapid (high saturated hydraulic conductivity), and sand drains rapidly. As a result, sands have low water content at field capacity. On the other hand, fine-textured soil such as clay has high porosity. Although the pores are smaller than those of sand, clay has many more pores. Strong adsorptive forces bind water to the soil matrix (low matric potential when saturated), movement is slow (low saturated hydraulic conductivity), and drainage is impeded (high field capacity). Loams are intermediate, draining more slowly than sands and retaining more water.

Infiltration is the vertical flow of water into soil. The maximum amount of water that can infiltrate is the infiltration capacity. This depends on soil wetness at the start of infiltration and the length of time. In general, infiltration rates are high initially and decrease as the soil becomes wet. Many methods are available to estimate infiltration. The Green–Ampt equation uses the notion of a horizontal wetting front that moves downward in the soil (Hillel 1998). From Darcy's law, the infiltration rate (i, mm s^{-1}) is:

$$i = -K_{sat}\left(\frac{0-\psi_{sat}}{z_{sat}}+1\right) = -K_{sat}\left(\frac{-\psi_{sat}+z_{sat}}{z_{sat}}\right) \tag{9.10}$$

where z_{sat} is the depth (mm) of the wetting front. The surface is assumed to have a matric potential of zero, and the wetting front is saturated with a matric potential equal to ψ_{sat}. The total cumulative infiltrated water (I, mm) is given by the depth of the wetting front multiplied by the difference between saturated water content (θ_{sat}) and initial water content (θ_i):

$$I = z_{sat}(\theta_{sat} - \theta_i) \tag{9.11}$$

The infiltration rate is the negative of the time derivative of I (the negative occurs because a negative water flux means downward motion):

$$i = -\frac{dI}{dt} = -(\theta_{sat} - \theta_i)\frac{dz_{sat}}{dt} \tag{9.12}$$

Combining Eq. (9.10) and Eq. (9.12) gives an equation for z_{sat} with respect to time:

$$(\theta_{sat} - \theta_i)\frac{dz_{sat}}{dt} = K_{sat}\left(\frac{-\psi_{sat} + z_{sat}}{z_{sat}}\right) \tag{9.13}$$

Solution of this equation, substituting Eq. (9.11) for z_{sat}, provides the time (t, seconds) required for a given cumulative infiltration (I):

$$t = \frac{I}{K_{sat}} + \frac{\psi_{sat}(\theta_{sat} - \theta_i)}{K_{sat}}\ln\left[1 + \frac{I}{-\psi_{sat}(\theta_{sat} - \theta_i)}\right] \tag{9.14}$$

Equation (9.14) relates cumulative infiltration (I) to time (t) given the soil properties K_{sat}, ψ_{sat}, and θ_{sat} (Table 9.2) and initial water content (θ_i). The infiltration rate (i) at time t can be found from Eq. (9.10) and Eq. (9.11):

$$i = -K_{sat}\left[\frac{-\psi_{sat}(\theta_{sat} - \theta_i)}{I} + 1\right] \tag{9.15}$$

For example, if K_{sat} = 0.007 mm s^{-1}, ψ_{sat} = -478 mm, θ_{sat} = 0.451, and θ_i = 0.392 (values representative of a loam soil at field capacity), the time required for I = 25 mm of water to infiltrate into the soil is about 17 minutes. At this time, the depth of the wetting front is z_{sat} = 424 mm, meaning the top 424 mm of soil is saturated. Calculation of infiltration at different times shows a high initial infiltration rate that decreases towards saturated hydraulic conductivity over time (Figure 9.8). Sand has the greatest infiltration capacity, loam an intermediate value, and clay has the lowest infiltration capacity. The same soils have higher initial infiltration rates and accumulate more water over one hour when drier.

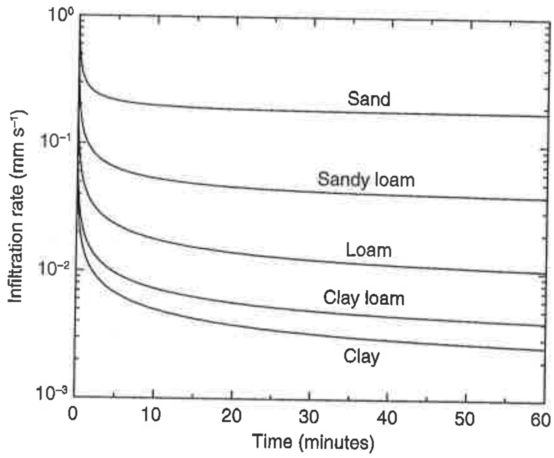

Fig. 9.8 Infiltration over a 60-minute period for sand, sandy loam, loam, clay loam, and clay soils at field capacity using the Green–Ampt equation with hydraulic properties from Table 9.2.

9.5 | Review Questions

1. On a cold winter day, which floor feels colder when walking on it barefoot - a wood floor with a thermal conductivity of 0.15 W m^{-1} K^{-1} or a tile floor with a thermal conductivity of 2 W m^{-1} K^{-1}?

2. The heat flux measured in soil at depth 50 mm is -150 W m^{-2}. The heat flux is -120 W m^{-2} at depth 100 mm. Calculate the time rate of change of temperature ($\Delta T / \Delta t$). The heat capacity is ρc = 2.9 MJ m^{-3} K^{-1}.

3. Calculate the heat flux in Figure 9.3b using the thermal conductivity of sand (2.0 W m^{-1} K^{-1}), clay (1.4 W m^{-1} K^{-1}), peat (0.5 W m^{-1} K^{-1}), and snow (0.34 W m^{-1} K^{-1}). How does the heat flux in organic soil and snow compare with that of mineral soil? Which material is likely to warm more?

4. Geothermal heating takes advantage of soil heat storage to heat and cool a house. A heat pump placed at a specified depth in soil transfers heat from the warm soil in the winter and conversely cools a building in summer. If the annual mean temperature is 11°C and the temperature range is 24°C, find the depth at which soil temperature does not drop below freezing throughout the year. Use $\alpha = 7 \times 10^{-7}$ m² s⁻¹. At what depth does soil temperature not vary by more than ±1°C from the annual mean?

5. Explain why the temperature in a stone cathedral is less than the outside air temperature on a hot summer day.

6. A volume of soil 10 cm wide × 10 cm long × 10 cm deep has a mass of 1.5 kg when wet. The soil mass is 1.25 kg when dry. Calculate the volumetric water content.

7. Find the available water holding capacity of sandy clay loam with ψ_{sat} = -299 mm, θ_{sat} = 0.42, and b = 7.12. How much water is available for evapotranspiration in a sandy clay loam soil that is 100 cm deep?

8. The flux of water into a volume of soil 50 mm deep is -2.5 mm hr^{-1}. The flux of water out of the soil column at the bottom is -1.0 mm hr^{-1}. Calculate the time rate of change in water content ($\Delta\theta / \Delta t$).

9. In the example problem given in Figure 9.6, calculate the water flux if the flow is horizontal rather than vertical. Both points are at the same height and are separated by a horizontal distance of 500 mm.

10. A soil 50 cm deep has a volumetric water content of 0.17. What is the depth of water that must be applied to the soil to raise the water content to 0.33?

11. A soil has an initial volumetric water content of 0.12 and water content at saturation of 0.42. How deep will 25 mm of rainfall saturate the soil?

12. From the Green–Ampt equation, what is the infiltration rate at infinity?

13. A column of soil is saturated with water. Use the data in Table 9.2 to calculate the initial rate of drainage from the bottom of the column for (a) sandy loam, (b) loam, and (c) clay loam. Which soil has the highest initial drainage rate?

9.6 | References

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