9
Water

The hydrosphere, with a strong influence on the pedosphere, comprises the sum total of all water bodies (oceans, rivers, lakes), groundwater (renewable and fossil), and soil water. Although water is the most abundant of all resources covering 70% of the earth's surface, freshwater is a scarce resource. The data in Table 9.1 indicate that 97.2% (volume basis) of the world water is in oceans and seas (1370 M Km^3). Freshwater accounts for merely 2.8% of the total volume, of which groundwater is 0.6%, and soil water accounts for less than 0.1% of the total (Fig. 9.1).

Soil is a major reservoir of freshwater, which accounts for about 50 times that in rivers and streams (Table 9.1). Some hydrologists classify the freshwater pools using simple nomenclature that reflects their functional characteristics. For example, *blue water* refers to water in water bodies that is lost from the land as runoff or seepage flow. This is the water that is temporarily lost for use by humans, animals, or plants. Freshwater, usable by primary producers (and comprising soil water, groundwater, and other irrigable sources), can be termed *green water*. The fraction of freshwater that is lost to the atmosphere through direct and soil evaporation may be termed *red water*. Fossil water is difficult to assess, is not renewable, and may be termed *gray water*. While simple and easy to comprehend, such terminology is vague, subjective, and arbitrary.

TABLE 9.1 Global Water Resources

Reservoir/Pool	Quantity (Km^3)	Percent of total
Water bodies		
Oceans	1,370,000,000	97.2
Freshwater lakes	125,000	<0.1
Saline lakes and inland areas	104,000	<0.1
Rivers and streams	1,300	<0.1
Ice sources		
Polar ice cap and glaciers	29,200,000	2.2
Lithosphere		
Soil water	67,000	<0.1

Groundwater	8,350,000	0.6
Atmosphere	13,000	<0.1

Source: Nace, 1971; Edwards et al., 1983; Goldman and Home, 1983; Van der Leeden et al., 1990; Alley et al., 2002.
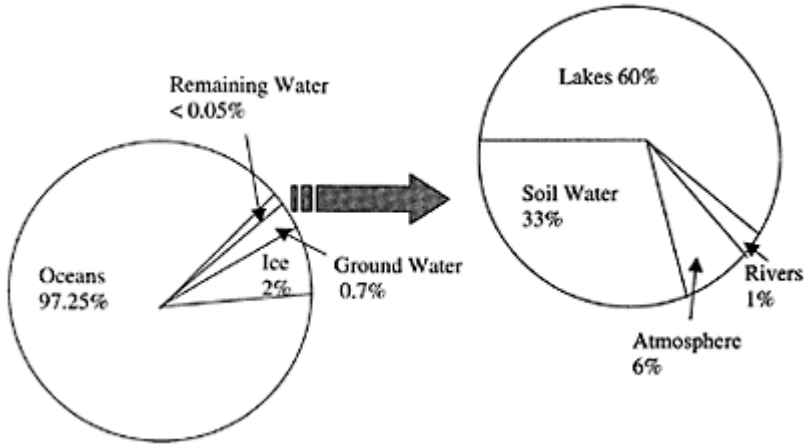
FIGURE 9.1 Different types of natural water.

9.1 PROPERTIES OF WATER

Principal properties of H_2O relevant to soil physical properties and processes are listed in Table 9.2. Some specific properties are described below.

9.1.1 Water Molecule

A single water molecule has a radius of 1.38 Å, at the center of which lies the oxygen nucleus. The oxygen and hydrogen protons in the water molecule

TABLE 9.2 Properties of Water Relevant to Soil Physical Properties and Processes

Property	Value
Density at 20°C	998.20 Kg m^{-3}
Density at 3.98°C	1000.0 Kg m^{-3}
Viscosity at 0°C	1.787 centipoise
Viscosity at 20°C	1.002 centipoise

| Surface tension against air at 0°C	75.6 dynes cm^{-1}
Surface tension against air at 20°C	72.75 dynes cm^{-1}
Boiling point at NTP	100°C
Freezing point at NTP	0°C
Heat of vaporization	590 cal g^{-1}
Heat of freezing	80 cal g^{-1}

Source: Adapted from Weast, 1987.

are about 0.97 A apart, and hydrogen protons are about 1.54 A apart. Two hydrogen atoms are at an angle of about 105° from each other (Fig. 9.2), giving water an electric dipole of about 1.87×10^{-18} esu.

The dipole moment produces electric field in the vicinity of each molecule. The electric field of adjacent water molecules creates an attractive force creating relatively weak intermolecular hydrogen bond between the proton of the hydrogen atom of one molecule and the oxygen atom of the other. Therefore, water molecules are joined together through hydrogen bonding (Fig. 9.3). These bonds are weaker than covalent bonds.

One mole of water, about 18 cm^3 (18 g), contains 6.02×10^{23} molecules. Therefore, 1 cm^3 of water contains 3.3×10^{22} individual molecules. When water crystallizes at 0°C, it develops an open crystalline structure. Therefore, ice is less dense than liquid water at the same temperature because water expands on freezing. It is this expansion of water on freezing that causes changes in soil structure by repeated cycles of freezing and thawing (refer to Chapter 4).

Water has a very high boiling point, a very high melting point, and low density in the liquid phase (Table 9.2; see also Appendix 9.1). The liquid water molecules are freer to move, and they have greater internal energy. About 80 calories (334 joules or 3.34×10^8 ergs) of heat energy per g of water is liberated when water changes from liquid to solid. Therefore, the entropy of water is higher in the liquid than in the more orderly crystalline, or solid state. The heat of vaporization of water, the heat absorbed to change from liquid to vapor state, is about 590 calories (2463 joules, or 2.45×10^9 ergs) per g of water. Therefore, entropy of water is higher in the vapor than in the liquid state (see Chapter 17).
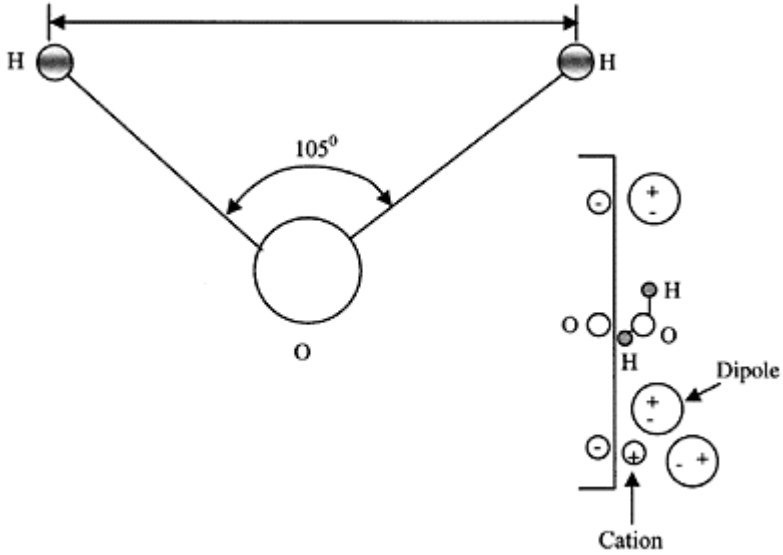
FIGURE 9.2 Water adsorption.
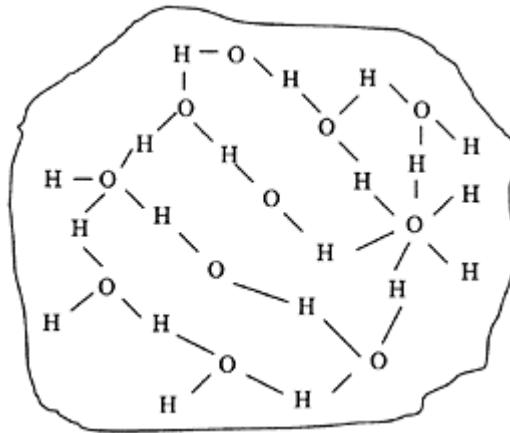
FIGURE 9.3 Water molecules joined together through hydrogen bonding.

9.1.2 Surface Tension

Water molecules at the air-water, solid-water, or another fluid-water interface are subjected to different forces than molecules within the bulk volume of the fluid. Water molecules within the bulk volume are hydrogen bonded to adjacent molecules and the cohesive forces are the same in all directions. At the air-water interface (or solid-water

interface), the force pulling into the air is much different than force within the bulk volume. For a molecule on the surface, there is a resultant attraction inward (Fig. 9.4).
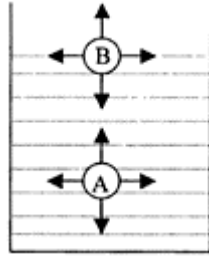
FIGURE 9.4 Forces acting on a molecule resting inside the liquid (A) and on the surface (B). The molecule on the surface has an unbalanced force making the water surface behave like a stretched membrane.

This imbalance in the force has the net effect of pulling the molecules in the one or two molecular layers near the surface into the bulk volume. The result is an orientation of molecules at the surface in such a way that pressure beneath the surface is much greater than above it, the surface behaves as if it were a stretched membrane, and the surface of the liquid always tends to contract to the smallest possible area. That is why the drop of liquid and bubbles of gas in a liquid become spherical. For a sphere, the surface is minimum for the given volume. In order to extend the surface, work has to be done to bring the molecules from the bulk of the liquid into the surface against the inward attractive force. This is called *free surface energy*. The difference in pressure is the cause of surface tension *(γ),* which is expressed in units of force per unit length or dynes/cm.

If a solid is immersed in water, the interfacial tension is due to the forces of adhesion (e.g., the forces of attraction for the water molecules by the solid and vice versa). These forces are the reasons for the work to be done to separate the solid from the liquid. The amount of work required is given by Eq. (9.1).

$$\frac{W_{sw}}{A_s} = \gamma_{sa} + \gamma_{wa} - \gamma_{sw} \tag{9.1}$$

The work *(W$_{sw}$)* is expressed in units of energy, and γ_{sa}, γ_{wa}, and γ_{sw} represent surface tension at the solid-air, water-air, and solid-water interfaces, and A_s is the area of the solid surface. Eq. (9.1) is called the Dupré equation (1969).

9.1.3 Contact Angle

Soil is a three-phase system: solid, liquid, and gas (see Chapter 2). Assuming that two fluid phases (liquid and gas) are in contact with soil solid, the
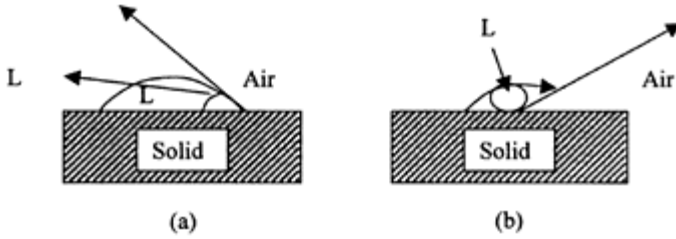
FIGURE 9.5 Angle of contact (a) is acute in a liquid that wets the solid, and (b) is obtuse in a liquid that does not wet the solid (L refers to liquid).

interface between air and water, forms a definite angle called the *contact angle* (Fig. 9.5). This angle is determined by Eq. (9.2) or Young's equation.

$$\cos \alpha = \frac{\gamma_{sa} - \gamma_{sw}}{\gamma_{wa}} \qquad (9.2)$$

The contact angle α thus depends on three interfacial tensions. However, whether it is acute (<90°) depends on the relative magnitude of γ_{sa} and γ_{sw}. If γ_{sa} exceeds γ_{sw}, then cos α is positive, and α is less than 90°. This is generally the case with most mineral soils and water, because water wets soils. If γ_{sw} exceeds γ_{sa}, then cos α is negative, and α is between 90° and 180°. This is the case of mercury and soil, because mercury does not wet the soil. When the liquid wets the solid (soils) the contact angle is acute and the liquid meniscus is convex, when it does not, the contact angle is obtuse and the meniscus is concave (Fig. 9.5).

Hydrophilic Versus Hydrophobic Soils

If the adhesive forces between the soil and water are greater than the cohesive force inside the water, and greater than the forces of attraction between the air and the soil, then the soil-water contact angle is acute and water will wet the soil. Therefore, hydrophilic soil can be defined as having the following characteristics:

Adhesive force (water-soil) >cohesive force (water-water) >adhesive force (soil-air)

A contact angle of zero implies complete flattening of the drop and perfect wetting of the soil surface by the water, and soil has absolute preference for the water over air.

A contact angle of 180° would mean a complete nonwetting or rejection of the water by the air-full soil. The water drop would retain its spherical shape without spreading over the soil surface.

When water is wetting the soil, the contact angle is low or acute. This low angle is called "wetting" or "advancing" angle. When the soil is drying and the water film is receding, the contact angle is different. It is called "receding" or "retreating angle." This

difference in wetting and retreating angle is also the cause of soil-water hysteresis, which will be discussed in Chapter 10. Soil hydrophobicity is affected by some organic substances coated on aggregate surfaces, such as in the case of the formation of algal crust. In such cases, the angle of contact can be modified through management. Plowing and physically rupturing the crust can improve wetting. In irrigated soils, wettability can be improved by use of surfactants.

9.1.4 Capillarity

A capillary tube in a body of water forms a meniscus as a result of the contact angle of water with the walls of the tube. The curvature of this meniscus will be greater (i.e., radius of curvature smaller), the narrower the tube (Fig. 9.6a vs. 9.6d). The height of capillary rise depends on the diameter of the section that corresponds with the pressure difference (Fig. 9.6b vs. 9.6c). Because of the difference in the contact angle, the water rises in the glass tube but mercury falls in the glass tube (Fig. 9.7).

A liquid with an acute angle will have less pressure inside meniscus than atmospheric pressure [Eq. (9.3)].
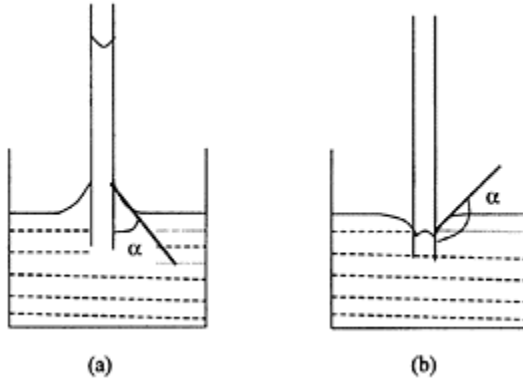
$$P_i < P_o$$

(9.3)

For a capillary of uniform radius r, at equilibrium the forces per unit area pulling down.

$$F_{\downarrow_g} = \pi r^2 h \rho_w g = \text{force of gravity}$$

(9.4)

(a)   (b)   (c)   (d)
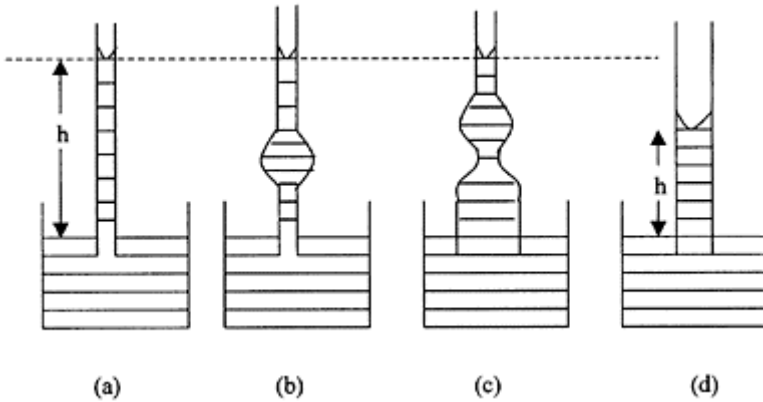
FIGURE 9.6 Rise of water in capillary tubes of different diameters.

FIGURE 9.7 Angle of contact for (a) glass and water is acute (<90°), and (b) glass and mercury is obtuse (>90°).

$$F_{\uparrow_\gamma} = 2\pi r \gamma \cos \alpha = \text{surface tension}$$

(9.5)

At equilibrium $F_{\downarrow_g} = F_{\uparrow_\gamma}$

$$\pi r^2 h \rho_w g = 2\pi r \gamma \cos \alpha$$

(9.6)

$$h = \frac{2\gamma \cos \alpha}{r \rho_w g}$$

(9.7)

$$\frac{2\gamma \cos \alpha}{r} = \rho_w g h = -\Delta p$$

(9.8)

where ΔP is the pressure difference across the interface. Eq. (9.8) is the equation of Young and Laplace for a spherical surface. If the contact angle between solid and liquid is zero (glass, mineral soil particle) than Eq. (9.8) is as simple as Eq. (9.9).

$$\Delta P = \frac{2\gamma}{r}$$

(9.9a)

$$h = \frac{2\gamma}{r \rho_w g}$$

(9.9b)

$$r = \frac{2\gamma}{h \rho_w g}$$

(9.9c)

Equations (9.8) or (9.9a) state that as a consequence of the existence of surface tension at a spherical surface of radius of curvature r, mechanical equilibrium is maintained between two fluids (water and air in soil) at different pressures.

Substituting the appropriate values for H_2O at 20°C in Eq. (9.9c) (γ=72.75 dynes/cm or g/s^2, ρ_w=0.9982 g/cm^3, and g=980 cm/s^2), we can solve for r assuming that α is zero [Eq. (9.10)].

r=(0.1487/h) cm

$$(9.10)$$

Soil being an extremely heterogeneous mass, there are numerous radii to influence the pressure across the water film. In Figs. 9.6b and c, there are two radii to be considered. The radius *(r)* is within the liquid phase and forms a convex surface. If the bubble is not spherical and has two principal radii (r_1 and r_2), and assuming that the contact angle is zero, the pressure difference across an interface with two principal radii is given by Eq. (9.11).

$$\Delta P = \gamma \left(\frac{1}{r_1} + \frac{1}{r_2} \right)$$

$$(9.11)$$

The rise or fall of liquids in capillary tubes is used to calculate the pore size distribution (Chapter 6). Whether a liquid rises in a glass capillary (as water) or is depressed (as mercury) depends on the relative magnitude of the forces of cohesion (between the liquid molecules themselves) and the forces of adhesion (between the liquid and the wall of the tube). These forces determine the contact angle α.

The occurrence of a concave meniscus leads to the capillary rise, whereas a convex meniscus leads to capillary depression. As soon as the concave meniscus is formed, the pressure in the liquid under the curved surface is less than the pressure in the air. The liquid thus rises in the tube until the weight of the liquid column just balances the pressure difference (ΔP=$2\gamma/r$) and restores the hydrostatic equilibrium. The liquid column acts as a manometer to register the pressure difference across the meniscus.

Example 9.1

What is ΔP at the surface of a droplet of water and mercury with r of 1 mm?

Solutions

(a) For the air-water interface

$$\Delta P = \frac{2\gamma}{r} = 2 \left(72.75 \frac{\text{dynes}}{\text{cm}} \middle/ 0.1 \, \text{cm} \right) = 1455 \, \text{dynes/cm}^2$$

(b) For the mercury-air interface=8600 dynes/cm^2

　　The difference calculated across the interface does not refer to the difference due to vapor pressure, which may be substantial.

Example 9.2

What vacuum is needed to draw all the water out of a sintered glass funnel if the minimum pore size is 4 μm in diameter?

Solution

As the water is withdrawn from the pores, the maximum pressure is reached when a hemispherical bubble is formed with a radius just equal to that of the pore. Therefore, the

$\Delta P = (2\gamma/r) = 2(72.75 \text{ dynes/cm})/2 \times 10^{-4} \text{ cm} = 72.75 \times 10^{4} \text{ dynes/cm}$

9.1.5 Osmotic Pressure

Osmotic pressure is a property of solutions, expressing the decrease of the potential energy of water in solution relative to that of pure water (see Chapter 20).

When an aqueous solution is separated from pure water (or from a solution of lower concentration) by a membrane that is permeable to water alone, water will tend to *diffuse* or *osmose* through the membrane into the more concentrated solution, thus diluting it or reducing the potential energy difference across the membrane. The osmotic pressure is the counter pressure that must be applied to the solution to prevent the osmosis of water into it.

In dilute solutions, the osmotic pressure is generally proportional to the concentration of the solution and to its temperature according to Eq. (9.12).

$P_s = KTC_s$

$$(9.12)$$

where P_s is osmotic pressure, T is absolute temperature, and C_s is concentration of solute.

An increase in the osmotic pressure is usually accompanied by a decrease in the vapor pressure, a rise of the boiling point, and a depression of the freezing point.

9.1.6 Solubility of Gases

The concentration of gases in water generally increases with pressure and decreases with temperature. According to Henry's law, the mass concentration of gas C_m is proportional to the pressure of gas P_i [Eq. (9.13)].

$$C_m = S_c \frac{P_i}{P_o}$$

$$(9.13)$$

when C_m is mass concentration of gas, S_c is solubility coefficient of the gas in water, P_i is pressure of the gas, P_o is total pressure of the atmosphere, and C_m is mass of dissolved gas relative to the mass of H_2O. The volume concentration is similarly proportional to Eq. (9.14).

$$C_v = S_v \frac{P_i}{P_o}$$

$$(9.14)$$

where S_v is solubility expressed in terms at volume ratio, and C_v is volume of dissolved gas relative to the volume of H_2O.

9.1.7 Viscosity

Viscosity of a fluid is its resistance to flow. For example, water has lower viscosity than syrup or honey. When fluid is moved in shear (adjacent layers of fluid are made to slide over each other), the force required is proportional to the velocity of shear. The proportionality factor is called viscosity (η). It is the property of fluids to resist the rate of shearing, and can be visualized as an internal friction.

The coefficient of viscosity η is defined as the force per unit area necessary to maintain a velocity difference of 1 cm/sec between two parallel layers of fluid that are 1 cm apart. The viscosity equation is shown in Eq. (9.15).

$$\tau = \frac{F_s}{A} = \eta \frac{du}{dx}$$

(9.15)

where τ is shearing stress, F_s is force, A is area of action for the force, and du/dx is velocity gradient normal to the stressed area.

Kinematic Viscosity (η_κ)

The ratio of the viscosity to the density of the fluid is called the *kinematic viscosity* (η_κ). It expresses the shearing-rate resistance of a fluid mass independently of the density.

While η of water is about 50 times more than that of air, $\eta\kappa$ of water is actually lower. Viscosity has the units of poise or centipose (see Appendix 9.1 and Appendix L).

9.1.8 Newtonian and Non-Newtonian Fluids

Newtonian fluids obey Newton's law of viscosity, which states that shear stress (τ) is proportional to shear rate, with the proportionality constant being the coefficient of viscosity (η) as shown in Eq. (9.15) and Fig. 9.8.

For solids, shear stress divided by shear strain gives an elastic modulus [refer to Eq. (7.17)]. For viscous liquids, since the strain is increasing all the time, shear stress divided by the rate of shear strain gives the viscosity coefficient. Newtonian fluids have a constant viscosity at a given temperature. Examples of Newtonian fluids are water, salt solution, milk, mineral oil, etc. In general, all gases and most liquids with simpler molecular formula and low molecular weight (e.g., water, benzene, ethyl alcohol, CCl_4, hexane, and most solutions of simple molecules) are Newtonian fluids.

Non-Newtonian fluids do not obey Newton's law of viscosity. Such fluids have a variable viscosity at a constant temperature $\eta = f(t)$, and viscosity depends on the force applied (time and temperature).

$$\tau = \eta \frac{\partial v}{\partial x}$$

(9.16)

where v is the velocity and x is the distance, η is the apparent viscosity and is not a constant (Fig. 9.8). Examples of non-Newtonian fluids are a syrupy mixture of cornstarch and water, quicksand, slurries, pastes, gels, polymer solutions, etc.

In some non-Newtonian fluids, properties are independent of time under shear. Such fluids include the following:

Bingham Plastic

These fluids resist a small shear stress but flow easily under larger shear stresses e.g., toothpaste, jellies, and some slurry.

Pseudoplastic

Viscosity of the fluids decreases with increasing velocity gradient (e.g., polymer solutions, blood). Pseudoplastic fluids are also known as shear thinning fluids. At low shear rates *(dv/dx)* the shear thinning fluid is more viscous than the Newtonian fluid, and at high shear rates it is less viscous. Most non-Newtonian fluids fall into this group.
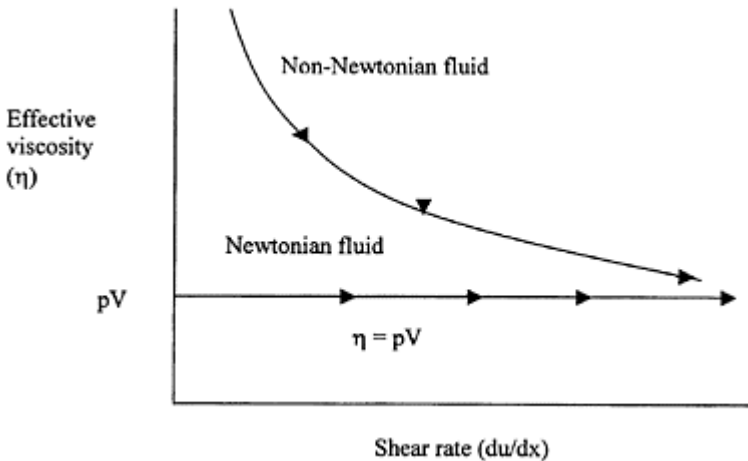
FIGURE 9.8 Newtonian and non-Newtonian fluids.

Dilatant Fluids

Viscosity of these fluids increases with increasing velocity gradient. They are uncommon, but suspensions of starch and sand behave in this way. Dilatant fluids are also called shear thickening fluids.

In other non-Newtonian fluids, properties are dependent upon duration of shear. Such fluids include the following:

Thixotropic Fluids

For thixotropic fluids the dynamic viscosity decreases with the time for which shearing forces are applied (e.g., thixotropic jelly paints).

Rheopcctic Fluids

For Rheopectic fluids the dynamic viscosity increases with the time for which shearing forces are applied (e.g., gypsum suspension in water).

Viscoclastic Fluids

Viscoelastic fluids have elastic properties, which allow them to spring back when a shear force is released (e.g., egg white).

9.1.9 Fluidity

Fluidity is the reciprocal of viscosity and has the units of 1/poise or 1/centipose.

$$F = \rho g / \eta = \frac{g}{cm^3} \cdot \frac{cm}{s^2} \cdot \frac{s \cdot cm}{g} = \frac{1}{cms} = \frac{1}{poise} \tag{9.17}$$

Fluids of lower viscosity flow more readily than those of high viscosity. The fluidity of water increases by about 3% per 1°C rise in temperature. The fluidity is also affected by the type and concentration of solutes.

9.1.10 Vapor Pressure

The change of state of water from liquid to vapor phase is related to the kinetic theory. The molecules in a liquid move past one another in a variety of speeds. A molecule in the upper regions of the liquid with a high speed may leave the liquid momentarily and fall back. Others with a critical speed may escape. The number of molecules with high KE increases with increase in temperature.

Evaporation $E \propto T$

Water evaporation is an endothermic process. When H_2O molecules with high energy escape, the velocity and kinetic energy of those remaining is less, the lesser the velocity the lower is the temperature. The liquid is, therefore, cool.

Saturated Vapor Pressure

The vapor is in equilibrium with a liquid because the rate of molecules escaping and those returning back by condensation is equal.

The equilibrium exists when the space is saturated. The pressure of the vapor when it is saturated is called the "saturated vapor pressure." The saturated vapor pressure does not depend on the size of the container. It depends on:

1. Pressure of water
2. Temperature of water
3. Chemical condition (solutes)

Boiling Point

Water boils at a temperature when vapor pressure becomes equal to atmospheric pressure. The saturated vapor pressure is related to the temperature *(T)* as per the simplified version of the Clasius–Clapeyron equation [Eq. (9.18)].

$$\ln P_o = a - \frac{b}{T} \qquad (9.18)$$

where In P_o is logarithm to the base e of the saturation vapor pressure P_o, T is absolute temperature, and a and b are constant.

Pressure of the liquid water also affects vapor pressure. Water in soil is a dilute solution of various electrolytes. The vapor pressure of electrolytes is lower than that of pure water, soil-water also has a lower vapor pressure even when the soil is saturated. In an unsaturated soil, capillary and adsorptive effects further lower the potential and the vapor pressure.

Vapor pressure is expressed in units of pressure, e.g., dynes/cm^2, bar, mm of Hg, or water. The vapor pressure of atmosphere can also be expressed in the following different ways:

1. $$RH = \frac{\text{partial pressure of water}}{\text{saturated pressure of water}} \times 100$$

2. $$(\rho_v)\ \text{Vapor density} = \frac{\text{mass of water vapor}}{\text{volume of air}}$$

3. $$\text{Specific humidity} = \frac{\text{mass of water vapor}}{\text{mass of air}}$$

4. Saturation (or vapor pressure) deficit=the difference between the existing vapor pressure and the saturation vapor pressure
5. Dew point temperature: The temperature at which the existing vapor pressure becomes equal to the saturation vapor pressure, i.e., the temperature at which a cooling body of air with a certain vapor content will begin to condense dew.
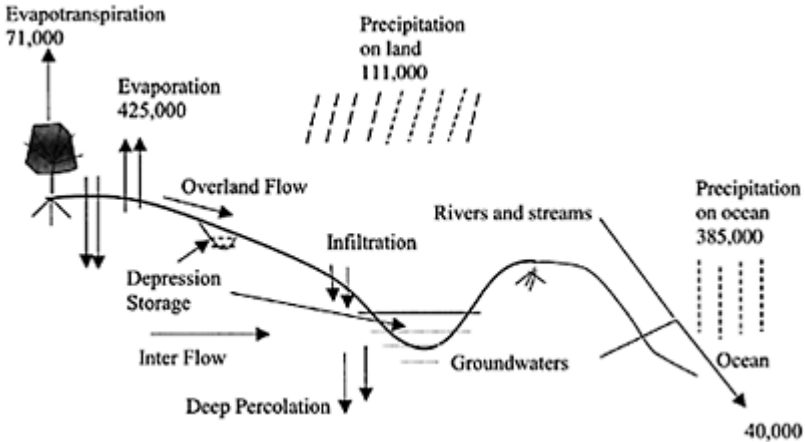
9.2 THE HYDROLOGIC CYCLE

Water is a completely renewable resource. It changes from one form to another and from one environment to another. Water transfer or movement from one form and/or one environment to another governs the hydrologic cycle (Fig. 9.9). The hydrologic cycle involves interchange (fluxes) between principal pools. These fluxes are: (i) evaporation, transpiration, or evapotranspiration (red water) by which water enters the atmosphere, (ii) precipitation by which returns to the land and ocean, and (iii) infiltration, percolation,

interflow, and runoff or overland flow by which water is returned from land to streams, rivers, lakes, and oceans (blue water). Soil water (green water) is a principal pool of the freshwater reserves. The magnitude of these pools and fluxes is shown in Fig. 9.9.

The data in Fig. 9.9 and Table 9.3 can be used to calculate the mean resident time (T_r) which is equal to the mass/flux. The T_r for water in the atmosphere, streams/rivers, and oceans is given by Eqs. (9.19) to (9.21).

T_r atmosphere=13,000 Km^3/496,000 Km^3 per yr=0.026/yrs

$$(9.19)$$

FIGURE 9.9 Schematic of global transfer rates (km^3/ yr) for water movement in the hydrologic cycle. (Flux rates are from Spiedel and Agnew, 1982 and Alley et al., 2002).

TABLE 9.3 Major Water Pools and the Mean Residence Time (Tr) of Water in Specific Pool

Pool	Capacity (Km^3)	Flux (Km^3/yr)	Tr (yr)
Oceans	1,370,000,000	425,000	3223.5
Freshwater lakes	125,000		
Saline lakes and inland seas	104,000		
Rivers and streams	1,300	40,000	0.0325
Glaciers and ice caps	29,200,000		
Soil water	67,000	111,000	0.604
Groundwater	8,350,000		

Atmosphere	13,000	496,000	0.0262

Source: Calculated from Spiedel and Agnew, 1982, and Alley et al., 2002.

T_r streams/rivers=1,300 Km^3/40,000 Km^3 per yr=0.033 yrs

(9.20)

T_o oceans=1,320×106 Km^3/485×103 per yr=3,100yrs

(9.21)

Therefore, atmospheric and stream/river pools are highly dynamic and can transfer contaminants or pollutants from one pool to another very rapidly. The T_r of water in soil is highly variable, and depends on soil properties.

9.3 SOIL AS A RESERVOIR OF WATER

The total pool of soil-water is estimated at about 67,000 Km^3. In terms of the freshwater reserves, soil is, in fact, a very efficient storage system. Assume that a one-hectare area of soil has water content of 20% by weight in the top 1-m depth with an average bulk density of 1.25 Mg/m^3. The total amount of water in the soil is 0.25 hectare-meter, 2.5×10^3 Mg, 2.5×10^6 Kg, or 2.5×10^6 L. This is indeed a large quantity of water. If human consumption of water is about 100 L/day, this water is enough for one person for 2.5×10^4 days or 68.5 years or for 25,000 people for one day. If half of this water were available for plant uptake at the consumptive use rate of 0.5 cm/day, it can support plant growth for 25 days.

Rather than the absolute quantity, it is often change in the soil-water pool that is of major interest. The change in the soil-water pool can be computed from the water balance Eq. (9.22).

$\Delta S=P+I-(R+D+ET)$

(9.22)

where ΔS is the change in the soil-water pool, P is precipitation, I is irrigation, R is surface runoff, D is deep drainage, and ET is evapotranspiration. Different components listed in Eq. (9.22) are determined by lysimetric evaluation.

9.4 COMPONENTS OF THE HYDROLOGIC CYCLE

Different components of the hydrologic cycle are outlined in Eq. (9.22), which is normally written in a form to solve for ET [Eq. (9.23)].

$ET=P+I-(R+D)\pm AS$

(9.23)

Therefore, different components of the hydrologic cycle include: (i) precipitation *(P)* including rain, snow, hail, fog, mist, (ii) irrigation *(I)* is not a component in natural ecosystem but is an important factor in the hydrologic cycle of managed and especially

agricultural ecosystems in arid and semi-arid regions, (iii) *R* is surface runoff, (iv) *D* is deep drainage leading to groundwater recharge, (v) ΔS is change in soil water storage, and (vi) ET is evapotranspiration. Methods of measurement and estimation or prediction of evapotranspiration are described in detail by Monteith (1985), and standard methods of measuring precipitation are discussed in texts on climatology or any hydrologic manual (USDA, 1979).
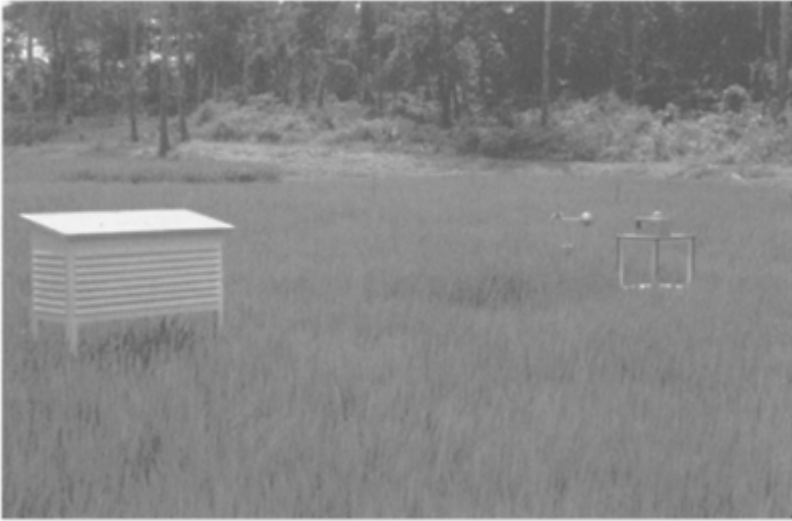
9.4.1 Precipitation

Accurate measurement of precipitation is important for reliable assessment of the water balance. In addition to simple or non-recording and recording rain gauges normally used at the meteorological stations (Figs. 9.10–9.13), rainfall measurement under a vegetation cover involves measurement of: (i) through fall using a spider gauge (Fig. 9.14a), and (ii) stem flow (Fig. 9.14b). Measurement of through fall and stem flow can be highly variable depending on the tree canopy and foliage characteristics.

9.4.2 Runoff

There are numerous methods of measuring surface runoff for different scales. The scale may range from a microplot of a few square meters to a watershed of several Km^2 or more (Table 9.4). Hydrologic parameters that are measured to compute surface runoff include total volume stage or water level, velocity, discharge, and their variation over time. Installation, measurements, and calibration procedures of these devices are described in USDA (1979), and shown in Figs. 9.15–9.18.

9.4.3 Lysimetric Analysis

A lysimeter is a confined volume of soil, in which input, output and change in water storage can be quantified. The size, shape, and material used in constructing lysimeters vary widely.

FIGURE 9.10 A meteorological
station installed within a rice paddy.

Lysimeters may be square (Fig. 9.19) or circular (Figs. 9.20–9.22), and made of steel, galvanized material, fiberglass, or plastic. Hydrologic inputs comprise precipitation and supplemental addition of water depending upon the management systems imposed. Hydrologic output comprises deep drainage or percolation water.

FIGURE 9.11 Recording and non-recording rain gauges.

FIGURE 9.12 A snow gauge.

(a)

(b)

FIGURE 9.13 (a) Class A pan evaporemeter; (b) a device to measure evaporation in a lake.

Changes in soil-water storage can be measured by using neutron moisture meter or gypsum blocks.

There are several types of lysimeter depending on the method of construction, and evaluating hydrologic balance. Common types of lysimeters are outlined in Table 9.5.

The drainage is facilitated by using about a 5 cm thick layer of gravel, sand, or diatomaceous clay at the base


(a)

(b)

FIGURE 9.14 (a) Spider gauge to measure through-fall and (b) stem flow.

TABLE 9.4 Method of Measuring Surface Runoff

| Technique/plot	Size	Equipment
Microplots	1–10m^2	A drum with a capacity of about 200 liter, or a small flume with water stage recorder
Field runoff plots	0.0025–100 ha	Multidivisor tanks, flume, water stage recorder
Small watersheds	1–10 ha	Flume, water stage recorder, proportional samplers
Large watersheds	>10ha	Weirs, waterstage recorders

Source: Adapted from Lal, 1990.
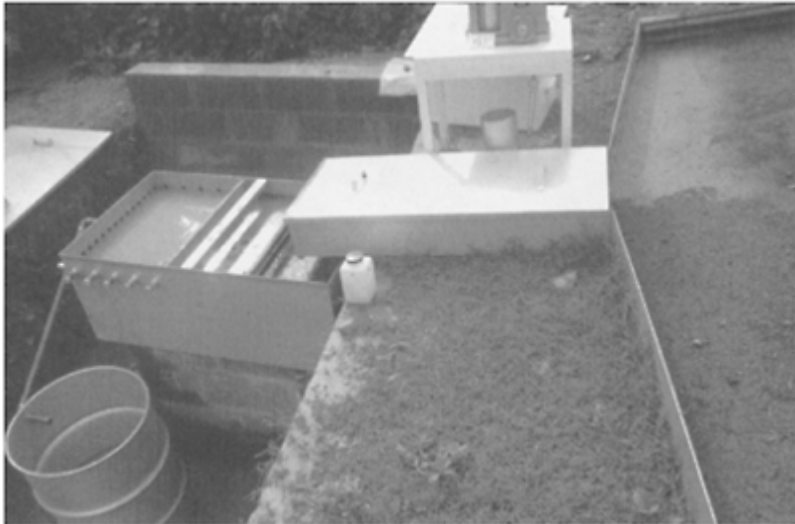
FIGURE 9.15 A multidivider tank and a flume with water stage recorder to measure runoff from a plot.

(Fig. 9.23). Lysimeters may be cited or different landscape positions in the field, or constructed at one cite to facilitate specific measurement (Figs. 9.24 and 9.25).

Lysimetric data are used to compute consumptive water use by plants or crops grown. An example of the method to use these data is shown below. Consider the data in Table 9.6 for 30-day period from a lysimetric experiment:

Consumptive use or ET per day=16 cm/30 days=0.53 cm/day
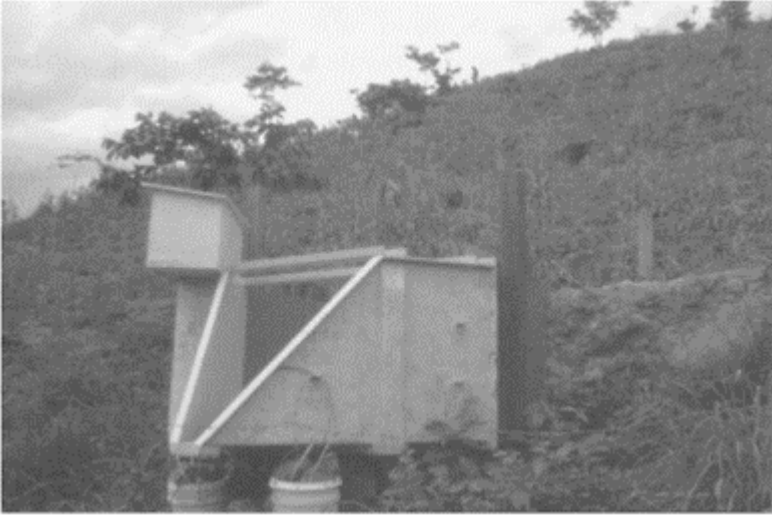
FIGURE 9.16 An H-flume and a
water stage recorder to measure runoff
from a steep agricultural watershed.

FIGURE 9.17 A wier with a slot-pipe
to collect runoff sample.

There are numerous uses of lysimetric experiments, with the primary use of measuring
the components of hydrologic cycle, especially deep drainage, soil-water storage, and
evapotranspiration. In addition, chemical analyses of the deep drainage or percolation

water can be extremely useful to study transport of chemicals applied to the soil, e.g., fertilizers and pesticides. Temporal changes in concentration of NO_3-N, PO_4-P, organic P, dissolved organic carbon can provide useful information on the risks of contamination of groundwater. Fate and pathways of pesticides can also be studied by lysimetric analyses.

Lysimetric studies are also useful to evaluate transport of clay from surface to the subsoil by the process of illuviation (Roose, 1977). The

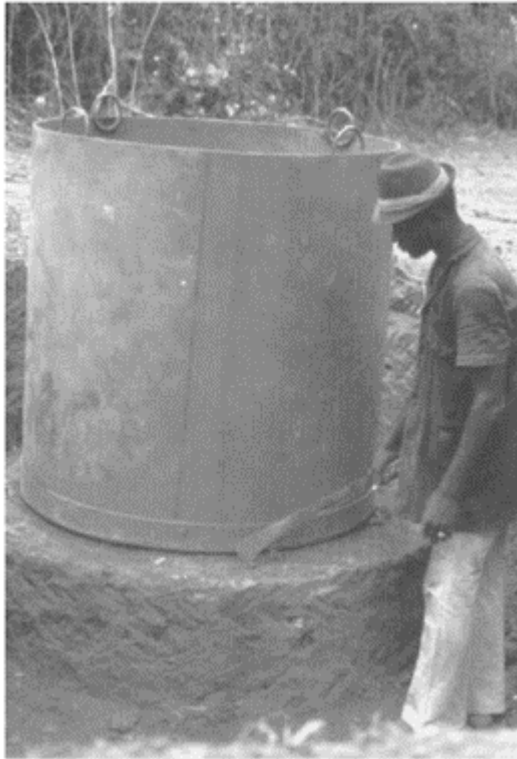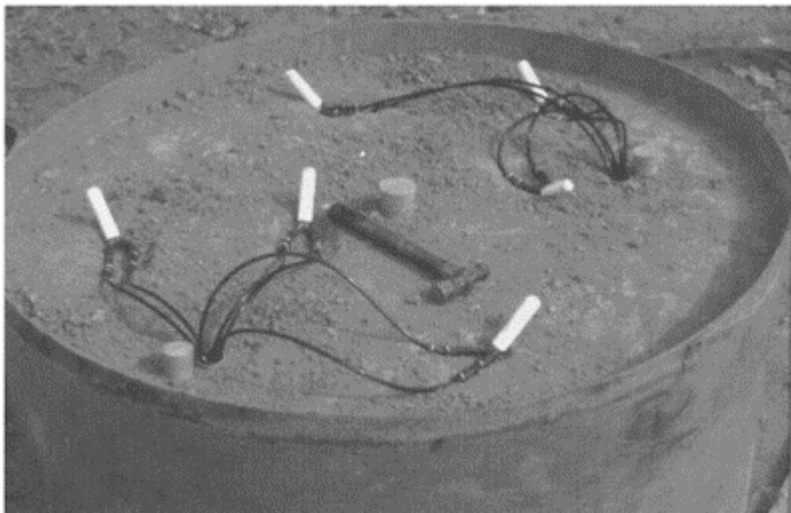FIGURE 9.18 A Coshocton wheel sampler to obtain runoff sample.
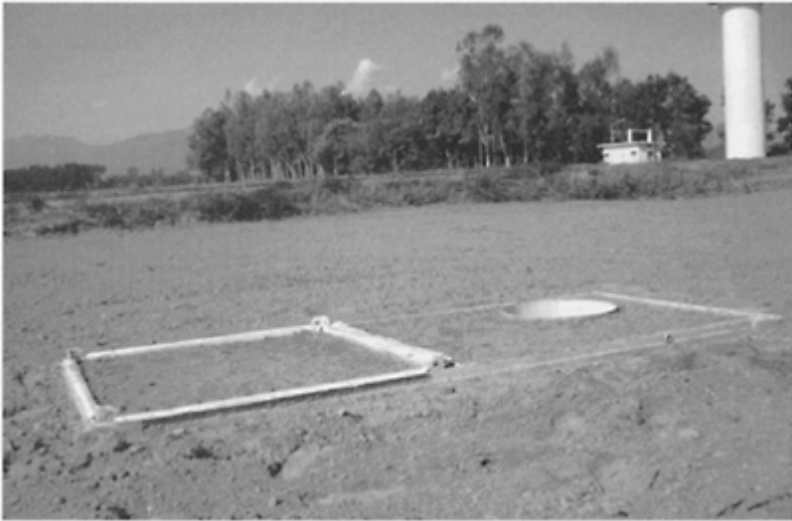
(a)

FIGURE 9.19 A square filled in lysimeter (a) method and (b) with removable cover.

(b)

FIGURE 9.20 Installation of a circular monoleith lysimeter.

FIGURE 9.21 A suction cup and
neutron probe access tube are installed
at the base.

FIGURE 9.22 Suction cups are
embedded in the diatomaceous clay.

information on solution weathering or rate of new soil formation can also be obtained by
chemical analyses (Al^{+3}, Si^{+4}, cations) of the percolating water. For these measurements,
lysimeters must be deep enough and include bedrock as a part of the monolith or soil
solum being studied.

TABLE 9.5 Types of Lysimeters Used for
Evaluating Components of the Hydrologic Cycle

Basis		Lysimeter types
Soil disturbance	(i)	Filled in, where disturbed soil is packed layer by layer at ρb similar to the field situation
	(ii)	Monolith, where a block of undisturbed soil is encased under natural conditions
Weighing		Nonweighing or drainage lysimeter in which water balance is obtained by carefully measuring the volume of water drained
		Weighing lysimeters monitor changes in total weight on a continuous basis or at regular time intervals. Weighing lysimeters may use a mechanical balance or a hydrologic weighing technique
Drainage	(i)	Gravity drainage

(ii) Suction drainage

Location   (i) In situ, constructed with soil in place

(ii) Constructed with soil transported from different regions

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Example 9.3

A runoff plot has a dimension of 25 m×4 m. The runoff collection system involves a Coshocton Wheel Sampler, which collects 1 % of the runoff. Total runoff collected after 2.5 cm of rainfall is 10 liters. The sediment load in runoff is 5 g/liter. Calculate runoff and erosion.

Solution

Total runoff volume=10 liters×100=1000 liters

$$\text{Runoff depth} = \frac{\text{volume}}{\text{area}} = 10^3\,\text{L} \times \frac{10^3\,\text{cm}^3}{\text{L}} \times \frac{1}{100\,\text{m}^2} \times \frac{10^4\,\text{cm}^2\,\text{m}^2}{10^4\,\text{cm}^2} = 1\,\text{cm}$$
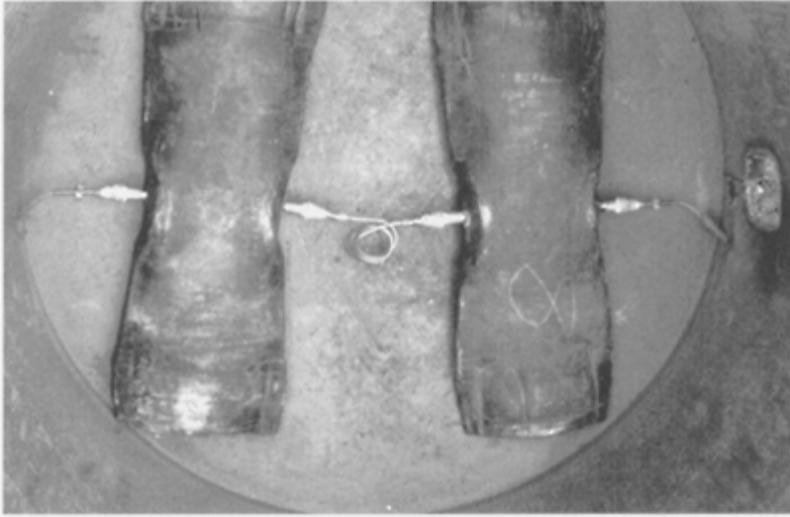
$$\text{Runoff C\% of rainfall} = \frac{1\,\text{cm}}{2.5\,\text{cm}} \times 100 = 40\%$$

$$\text{Total sediments} = 1000\,\text{L} \times \frac{5\,\text{g}}{\text{L}} \times \frac{\text{kg}}{10^3\,\text{g}} = 5\,\text{kg}$$
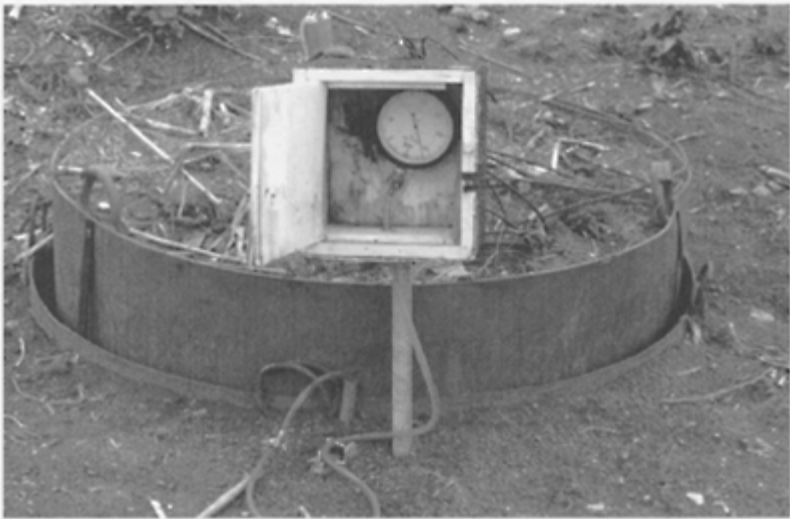
$$\text{Soil erosion} = \frac{5\,\text{kg}}{100\,\text{m}^2} \times \frac{10^4\,\text{m}^2}{\text{ha}} = 500\,\text{kg/ha} = 0.5\,\text{Mg/ha}$$

PROBLEMS

1. A lake has a capacity of 1200 Km^3. The steady state evaporation flux is 200 Km^3 y^{-1}. What is the mean residence time of water in the lake?

FIGURE 9.23 (a) A hydraulic weighting device may involve water-filled pillows placed beneath the lysimeter, and (b) connected to a pressure gauge.

2. A one hectare field contains 0.2 gg^{-1} of water to 10 m depth. Assuming a uniform soil bulk density of 1.5 Mg m^{-3}, calculate the total water content of soil in liters and equivalent depth.

3. Draw a landscape, and list principle components of the hydrologic cycles.

4. Tabulate methods of monitoring components of a hydrologic cycle along a hill slope.

(a)

(b)

FIGURE 9.24 A battery of drainage lysimeter (a) with a trench to collect seepage; (b) an underground weighing and seepage collection facility.

5. Draw up a table or a nomograph comparing different units of measuring water capacity and flux, and compute conversion factor to change from one unit to another.

6. Calculate the height of capillary rise in a soil pore of 50 μm inner diameter in winter (0°C), spring (10°C), early summer (20°C), and tropics (40°C).

7. Compute the pressure difference at the air-water interface in Question 1 above.

FIGURE 9.25 A series of lysimeters under a plastic shelter.

TABLE 9.6 Lysimetric Measurements

Period (days)	Precipitation	Irrigation	AS	Runoff	Deep drainage
0–10	0	5	−1	0	0
10–15	12	0	+4	3	2
15–30	5	0	−2	0	0

Calculate ET:
Solution:
ET=P+I−(R+D+AS)
ET For Period 1=0+5− (0+0−1)=6 cm
ET For Period 2=12+0−(3+2+4)=3 cm
ET For Period 3=5+0−(0+0−2)=7 cm
Total ET=16 cm

8. Consider the following equation of the height of capillary rise:

$$r = \frac{2}{h\rho g}$$

where γ and ρ refer to the surface tension and density of the fluid, respectively. What is the difference in the height of capillary rise in 20 μm diameter pore for water and alcohol at 20° C?

9. Write a brief essay on "surface tension." As a diagram, explain interactive forces, and define units.

10. The 0–50 cm layer of a lakebed soil in northwestern Ohio has a field capacity of 30% by weight, soil-water content of 15% by weight, and bulk density of 1.2 Mg m^3. A rainfall of 4 cm was received of which 75% was lost as runoff. Calculate the following:

1. What is the volume of runoff from a test plot of 25 m×40 m?

2. What is soil erosion (t/ha) if the runoff contained sediments of 25 g/liter?

3. What is the total NO_3 loss if concentration in runoff is 5 g/liter?

11. Why are some soils more wettable than others? Why does burning crop residue or any biomass make a soil hydrophobic?

APPENDIX 9.1 SOME PHYSICAL PROPERTIES OF WATER AT ATMOSPHERIC PRESSURE

Temperature (°C)	Density ρ (Mg/m^3)	Specific weight γ (N/m$^3\times10^3$)	Dynamic viscosity μ (N$\times$s/m$^2\times10^{-3}$)	Kinematic viscosity ($\eta\kappa$) (m^2/s$\times10^{-6}$)
0	1.0	9.810	1.79	1.79
5	1.0	9.810	1.51	1.51
10	1.0	9.810	1.31	1.31
15	0.999	9.800	1.14	1.14
20	0.998	9.790	1.00	1.00
25	0.997	9.781	0.891	0.894
30	0.996	9.771	0.797	0.800
35	0.994	9.751	0.720	0.725
40	0.992	9.732	0.653	0.658
50	0.988	9.693	0.547	0.553
60	0.983	9.643	0.466	0.474
70	0.978	9.594	0.404	0.413
80	0.972	9.535	0.354	0.364
90	0.965	9.467	0.315	0.326
100	0.958	9.398	0.282	0.294

0.001 N$\times$s/m^2=0.001 Pa$\times$s=−0.01P=1 cP=1 centipose
Source: Adapted from Weast, 1987; Julien, 1998.

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