15
Soil Water Evaporation

Soil water evaporation is an important component of surface energy balance. The rate and quantity of evaporation from a soil surface is a complicated process affected by many soil characteristics, tillage, and environmental interactions. Evaporation also affects plant available water content of soil and causes salinization in irrigated lands. It is known that energy and water availability largely dominate the process of evaporation, thus on an average these broad principles can be used to estimate direct soil water evaporation.

15.1 INTRODUCTION

Evaporation is the process of change of the state of water from a liquid to gaseous phase. It is a principal process of water cycling in the hydrosphere. Evaporation from a landscape may occur from plant canopies, free water surface, or soil surface. Evaporation of water from bare soil (i.e., in the absence of vegetation) is the process by which water is lost from the soil to the atmosphere. If the evaporation process is not controlled, a considerable amount of water can be lost from an irrigated or a rainfed cropland. Evaporation takes place from plowed land, shallow farmland, from soil between tree and row crops, and agricultural lands with no vegetation.

During planting and germination period, evaporation can reduce soil water content significantly and can hamper plant growth.

There are four conditions for evaporation from soil to occur. One, the evaporation from a bare soil takes place continuously, provided there is a continuous supply of energy. The amount of energy required is the latent heat of water for evaporation, which is about 590 cals/g of water evaporated at 15°C. The soil body itself, which gets cooler after rather than before evaporation, can supply this energy. Alternately, it can come from the advected or radiated energy from the surroundings. The second and third physical conditions for evaporation from bare soils are: the vaporpressure gradient between the soil and the atmosphere, and the transport of vapor away from soil by diffusion and/or convection. The energy for evaporation and vapor removal are generally external to the evaporating soil and are greatly influenced by meteorological factors (i.e., air temperature, humidity, wind velocity, and radiation). The evaporation rate is determined by the external evaporativity or water conductivity of the soil. The fourth condition is that there is a continuous supply of water within the soil body to the evaporating surface. This condition depends on both physical and conductance properties of soil. Some of them are: water content of soil body, soil water potential, hydraulic conductivity of soil, texture, compaction, soil horizonation or layering, and depth to the water table.

An important condition in poorly drained soils is evaporation in presence of a shallow ground water table. The evaporation occurs at a nearly steady state level and water content of profile remains almost the same. In the absence of a water table, the evaporation will dry the topsoil and the process is mostly in an unsteady state. The flow domains during the evaporation process may be one-dimensional or three-dimensional. The flow processes may be isothermal or non-isothermal and there may be the interactions between liquid flow and temperature gradients, conduction of heat, and vapor in the soil domain. The cracks inside the soil matrix form the secondary evaporation planes. The evaporation also depends upon the environmental conditions, which can be regular (i.e., diurnal, seasonal) or irregular (i.e., spells of cool and warm weather and of rewetting and drying). The presence of surface mulch, depth of soil, and degree of homogeneity of soil also alter evaporation from the soil surface.

15.2 THE EVAPORATION PROCESS

Following three processes summarize the evaporation of water from soil surface.

15.2.1 Transport of Water to the Soil Surface

As the evaporation process begins at the soil surface, a suction gradient is established between the surface soil and the layer beneath. This gradient forces the water to move upwards through capillary rise and supplies the water to the soil surface. This process continues as long as the soil underneath has enough water storage. The transfer of water from soil underneath is facilitated easily and for a much longer duration immediately after irrigation or a rainfall event, when soil water content is high.

15.2.2 Uninterrupted Supply of Heat to Change the State of Water

Solar energy is the most predominant source of heat for water evaporation. There are other minor sources of energy also, e.g., exothermic reactions, microbial activity, etc. The energy balance of the soil depends on physical and thermal properties. The latent heat of vaporization changes the water from a liquid to vapor state.

15.2.3 Transfer of Water Vapor from Soil Surface to Atmosphere

The vapor pressure immediately above the evaporating soil surface is lower than the vapor pressure inside the soil. This differential pressure creates a vapor pressure gradient, which enables water vapor to escape from soil to the atmosphere through the process of convection and diffusion.

15.3 SOIL DRYING DURING EVAPORATION

Evaporation leads to the loss of water, with attendant drying and depletion of the soil moisture reserves. This process of soil drying occurs in three distinct stages (Fig. 15.1) (Fisher, 1923; Pearse et al., 1949).

15.3.1 Initial Stage

When soil is very wet, evaporation of soil water is governed by external atmospheric conditions rather than soil properties. The soil has enough water, therefore, conductivity and supply of water to soil surface are at the potential rate. The evaporation rate during this stage is denoted as "potential evaporation." This stage is sustained over time because as the water content of soil profile decreases the hydraulic conductivity also decreases. However, hydraulic gradient increases and compensates for the reduction in hydraulic conductivity. This situation is analogous to the

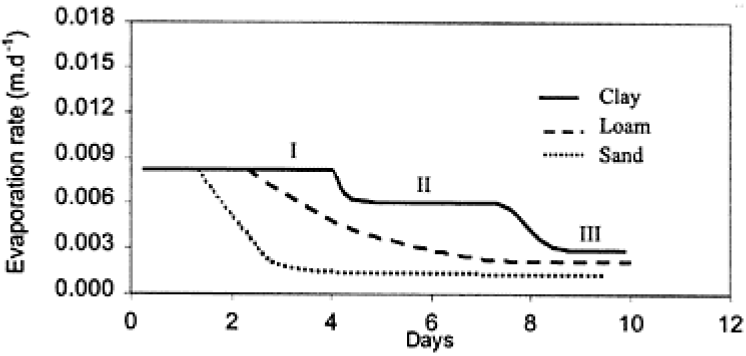

FIGURE 15.1 Three stages of evaporation during simultaneous drainage and evaporation from initially saturated profiles of sand, silt, and clay. (Modified from Hillel and van Bavel, 1976.)

flux-controlled stage of water infiltration into soil. Some soil properties, which influence the meteorological or atmospheric factors, include soil surface reflectance, mulch, ground cover, etc. The duration of the first stage of the drying process is lower for coarse-textured than fine-textured soils because fine-textured soils retain high water content and have more conductivity than coarse-textured soils. Figure 15.1 shows that the duration of the first stage of drying is in the order clay > loam > sand. The duration for first stage is also lower for a structureless than for a structured soil. Mulching increases the water content profile but shortens the duration of first stage of drying (Fig. 15.1). The duration of first stage is inversely proportional to soil water diffusivity (D_0).

15.3.2 Intermediate Stage

The evaporation rate during this stage is no longer at the potential rate but starts decreasing gradually with time. Soil starts to heat up and is not able to conduct water to the surface at the potential rate. The water content of the soil profile is decreased further as is the hydraulic conductivity. The hydraulic gradient can no longer increase significantly because the soil water pressure head is close to the partial water vapor pressure. The time at which the decrease in hydraulic conductivity is not compensated by hydraulic gradient denotes the end of first stage of drying. The depth of dry zone increases as does the hydraulic resistance of soil to water transport. The rate of evaporation during this stage is directly proportional to soil water diffusivity.

15.3.3 The Final Stage

The evaporation rate during this stage is relatively steady at a low rate and can continue up to several days. During this stage the liquid-water conductance totally ceases. This stage is also known as the vapor diffusion stage, since water transmission is primarily due to a slow process of vapor diffusion. The evaporation rate is determined by soil properties (affinity of the soil for water) rather than the evaporative demand of the atmosphere.

15.4 THEORY OF EVAPORATION

Specific processes of evaporation depend upon the presence of the water table at shallow depth, horizonation, and soil temperature region.

15.4.1 Steady Evaporation in the Presence of a Water Table

The essential and necessary condition for steady state flow is that the rate of change of flux density with depth is zero (d*q*/dz=0). Moore (1939) first studied the vertical steady state flow of water from a water table through soil profile during evaporation. Philip (1957), Gardner (1958), Ripple et al. (1972) discussed theoretical solutions for steady state evaporation. Mathematically, the steady upward flow can be described by the Darcy– Buckingham equation, for boundary condition z=0, Φm=0 and z is positive upward from the ground water table. The steady state upward flow or evaporation (q_e) can be expressed as follows [see also Darcy law for unsaturated flow] and see Eq. (13.3):

$$q_e = K(\Phi_m)\left(\frac{\mathrm{d}\Phi_m}{\mathrm{d}z} - 1\right) \qquad (15.1)$$

In terms of soil-water diffusivity [see Eq. (13.24)], the steady state evaporation is given by

$$q_e = D(\theta)\frac{\mathrm{d}\theta}{\mathrm{d}z} - K(\theta) \qquad (15.2)$$

where $D(\theta)$ is hydraulic diffusivity, θ is water content, $K(\theta)$ is hydraulic conductivity, z is height above the water table, and Φ_m is suction head. Eq. (15.1) shows that for $\mathrm{d}\Phi_m/\mathrm{d}z=1$, q=0. Eq. (15.1) can be rearranged as follows

$$\frac{q_e}{K(\Phi_m)} + 1 = \frac{d\Phi_m}{dz} \tag{15.3}$$

separation of variables results in

$$dz = \frac{K(\Phi_m)}{q_e + K(\Phi_m)} d\Phi_m \tag{15.4}$$

Integrating Eq. (15.4) to depths between 0 to z and suction gives the following expression in terms of depth of soil profile

$$z = \int \frac{K(\Phi_m)}{q_e + K(\Phi_m)} d\Phi_m \tag{15.5}$$

Similarly, Eq. (15.2) can be integrated for depths between 0 to z and following relationship in terms of z is obtained

$$z = \int \frac{D(\theta)}{q_e + K(\theta)} d\theta \tag{15.6}$$

Solutions of Eqs. (15.5) and (15.6) require prior knowledge of the functional relationships between $K(\Phi_m)$ and Φ_m and $K(\theta)$ and $D(\theta)$. For solving Eq. (15.5), Gardner (1958) proposed following relationship between Φ_m and K:

$$K(\Phi_m) = \frac{a}{\Phi_m^n + b} \tag{15.7}$$

where parameters a, b, and n are constants and are functions of type of soil. These parameters are soil-specific and need to be determined for each soil separately. Transferring Eq. (15.7) into Eq. (15.1) gives the following equation for evaporation rate (e) estimation

$$e = q_e = \frac{a}{\Phi_m^n + b}\left(\frac{d\Phi_m}{dz} - 1\right) \tag{15.8}$$

Transferring Eq. (15.7) into (15.5) gives the following equation, which provides the suction distribution with depth of soil for different fluxes

$$z = \int \frac{a/\left(\Phi_m^n + b\right)}{q_e + a/\left(\Phi_m^n + b\right)} d\Phi_m = \int \frac{a}{q_e\left(\Phi_m^n + b\right) + a} d\Phi_m \tag{15.9}$$

Steady rate of upward flow and evaporation rate from water table as a function of the suction prevailing at the soil surface for a sandy loam soil with n=3 is presented in Fig. 15.2 (Gardner, 1958). Fig. 15.2 shows that a

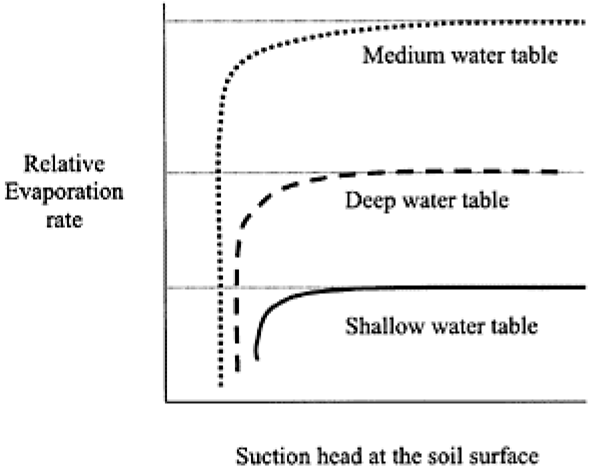

FIGURE 15.2 Steady rate of upward flow and evaporation from a water table as a function of the suction prevailing at the soil surface. The soil is sandy loam with *n*=3 (Modified from Gardner, 1958.)

steady rate of evaporation depends on the depth of this water table. The maximum possible evaporation for a ground water table at *z* will be for the lowest water content at the soil surface. Under this situation suction will tend to be infinite (pressure head infinity with negative sign). However, the extraction of water from soil profile is limited by the capacity of soil profile to transmit water or in other words by the hydraulic conductivity of soil. Ignoring the constant *b*, in Eq. (15.7), which leads to $K(\Phi_m) = a[\Phi_m^n]$. Gardner (1958) derived the following relationship between the depth of water table *(d)* and the maximum or limiting rate of transmission of water *(*q_{max}*)* by soil to the surface layer

$$d = \left(\frac{Aa}{q_{e\text{-max}}}\right)^{1/n} \tag{15.10}$$

where *A* is a constant and is a function of *n* and $q_{e\text{-max}}$, *a* and *n* are constants from Eq. (15.7). Equation (15.10) shows that evaporation and depth of water table are inversely related. Gardner (1958) showed that the evaporation rate from the soil is dependent on soil texture and is greater from medium textured soils as compared to the coarse textured soils (Fig. 15.3). Value of *n* is greater in coarse textured soils as compared to fine textured soils and therefore maximum evaporation rate

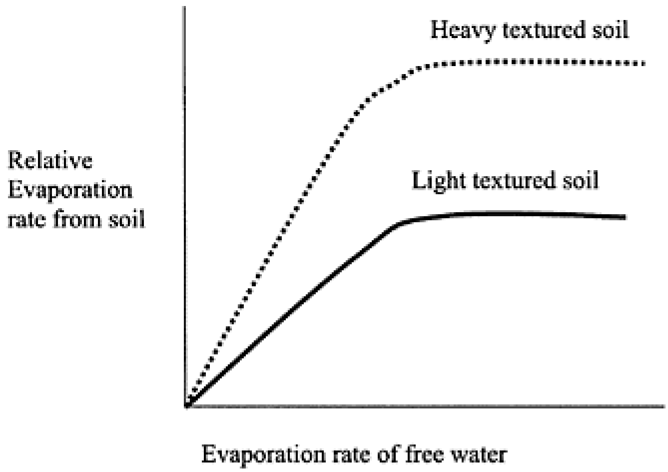

FIGURE 15.3 Schematic of evaporation rate as affected by texture of soil (water table depth, 60 cm). (Modified from Gardner, 1958.)

decreases more rapidly with depth in coarse textured soils as compared to the fine ones.

15.4.2 Evaporation in the Absence of a Water Table

Evaporation from soils in the absence of a water table is a transient process. Steady evaporation from soils is not always true because water table depths do not always remain constant for very long time. Therefore, a transient process describes the evaporation from soil surface more realistically. The transient condition implies that water content of soil profile does not remain constant, instead it decreases with evaporation as soil becomes drier. Another common assumption for steady state evaporation is that the external conditions (i.e., atmospheric evaporativity) remain constant, which is seldom true. However, for the sake of simplicity, constant atmospheric evaporativity is generally assumed for describing transient evaporation from the soil surface.

The process of transient evaporation or drying is already described as having three stages: initial constant rate stage, falling rate stage, and slow rate stage (Fig. 15.4). It is clear from Fig. 15.4 that the transition from the first to second stage of drying is sharp. However, transition from second to third stage is gradual and difficult to separate. In the initial stage of drying soil moisture depletion on the soil surface is compensated by soil underneath and evaporation remains by and large constant. Gradually, suction gradient inside the soil becomes larger with corresponding decrease in soil

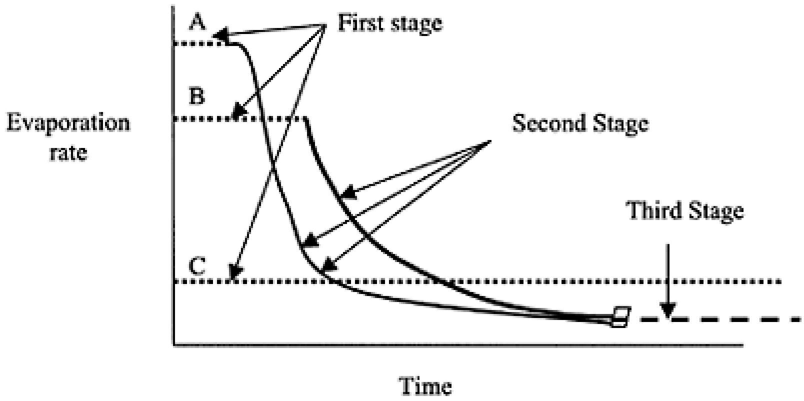

FIGURE 15.4 Stages of evaporation from a soil during steady atmospheric condition.

conductivity. This results in a decrease in evaporation rate with respect to time. The length of time the initial stage of drying can go on depends on the evaporativity. A low evaporativity will increase the duration of first stage of drying.

15.4.3 Evaporation from Layered Soils

Steady evaporation from layered soils can be determined similar to that from a homogeneous profile. Willis (1960) carried out the analysis by assuming that steady flow through layered profile depends upon the transmission property of soil. He further assigned that the suction or matric potential is continuous through the entire soil profile, although water content and conductivity are discontinuous using the relationship between $K(\Phi_m)$ and Φ_m [Eq. (15.7)] (Gardner, 1958) and assuming that each layer is internally homogeneous, he proposed the following relationship:

$$\int_0^{d_2} \mathrm{d}z + \int_{d_2}^{d_1+d_2} \mathrm{d}z = \int_{\Phi_{m0}}^{\Phi_{mL}} \frac{\mathrm{d}\Phi_m}{1+e/K_1(\Phi_m)} + \int_{\Phi_{mL}}^{\Phi_{m(L+d)}} \frac{\mathrm{d}\Phi_m}{1+e/K_2(\Phi_m)} \tag{15.11}$$

where d_1 and d_2 are the thickness of top and bottom layers respectively. Eq. (15.11) relates depth of water table to suction for a given evaporation rate. The limiting evaporation rate for a known water table depth can be calculated from above equation by assuming the suction (Φ_m) to

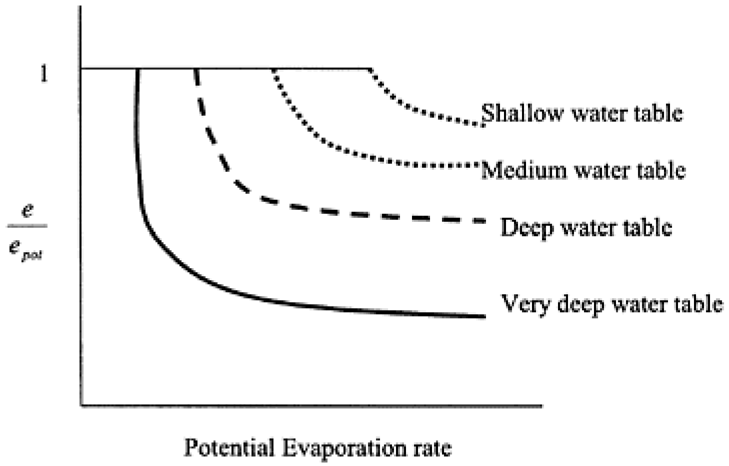

FIGURE 15.5 Dependence of relative evaporation rates, *(e/e_{pot})* upon potential evaporation rate (evaporativity, e_{pot}) for a clay soil. Numbers labeling the curve indicate the depth to water table (cm). (Modified from Ripple et al., 1972.)

be infinite at soil surface. Ripple et al. (1972) proposed a graphical method to measure the steady state evaporation from a multilayer soil profile. They included both the soil properties (i.e., water retention and transmission, vapor flow, depth of water table) and the meteorological factors (i.e., humidity, air temperature, and wind velocity) (Figs. 15.5 and 15.6).

15.4.4 Mathematical Modeling of Stages of Drying

The difference in suction at soil surface and a location with the soil body supplying water is much higher as compared to the depth of soil involved in the process of drying. Therefore, gravity effects are generally neglected for evaporation calculations. Most analysis is based on soil water content and the hydraulic diffusivity relationship. The first and second stages of drying depend upon the hydraulic diffusivity. In order to derive approximate description of drying in the first stage (Fig. 15.6), Gardner (1959) assumed that the evaporation rate from a soil profile of depth *(L)* could be expressed as

$$e = -L\frac{\partial\theta}{\partial t} \tag{15.12}$$

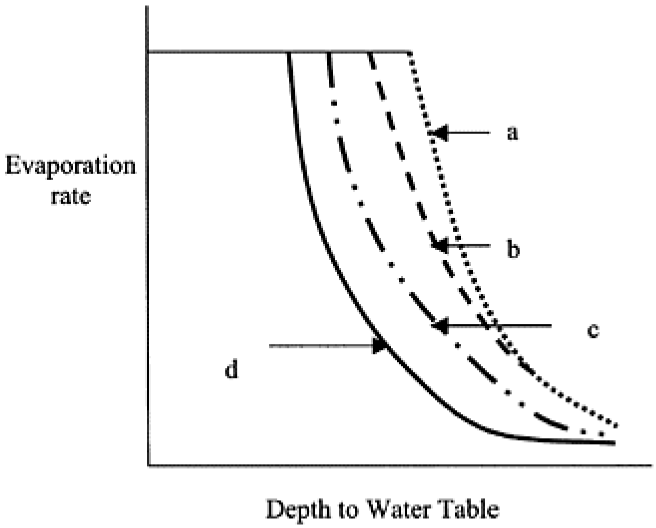

FIGURE 15.6 Effect of horizonation and water table depth on the evaporation rate: (a) limiting curve for soil water evaporation from for homogeneous soil; (b) a two-layer soil with the upper layer thickness of 3 cm; (c) thickness 10 cm; (d) a three-layer soil with thickness of intermediate and uppermost layers equal to 10 cm each. (Modified from Ripple et al., 1972.)

if the soil water diffusivity $(D(\theta))$ can be expressed by the following relationship (Gardner and Mayhugh, 1958)

$$D(\theta)=D(\theta)_0\exp[\beta(\theta-\theta_0)] \quad (15.13)$$

where $D(\theta)_0$ correspond to θ_0 and β ranges from 1 to 30. Gardner (1959) combined Eqs. (15.12) and (15.13) and after further approximation proposed that the total water content of soil profile can be approximated by the following relationship:

$$W=\frac{\beta W}{L} \quad (15.14)$$

and

$$D(\theta)=\frac{2D(\theta)_0}{e_{\mathrm{pot}}\beta L} \quad (15.15)$$

where W is the water storage in the entire soil profile at the end of first stage $t = t_1$.

$$W = \int_0^L \theta(z, t_1)\mathrm{d}z \tag{15.16}$$

Gardner and Hillel (1962) also assumed that the evaporation rate from soil profile is given by Eq. (15.12) and the flow equation as follows

$$-\frac{e}{L} = \frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left(D\frac{\partial\theta}{\partial z}\right) \tag{15.17}$$

where z is the height above the bottom of soil profile. Eq. (15.17) was integrated once. The constant of integration was assumed zero since flow through bottom of the soil profile (z =0) is zero. Assuming D can be represented by Eq. (15.13), Gardner and Hillel (1962) found that actual evaporation rate ceases to be equal to potential rate when at z=L, $\theta=\theta_0$. They proposed the following equation for total water content *(W)* of profile.

$$W = \frac{L}{\beta}\ln\left(1 + \frac{e\beta L}{2D_0}\right) \tag{5.18}$$

For a long soil column dependence of e on t is only approximately valid. For a soil column of finite length *(L)*, Gardner and Hillel (1962) proposed the following relationship to calculate evaporation during second stage (Fig. 15.5).

$$e = -\frac{\mathrm{d}W}{\mathrm{d}t} = D(\bar{\theta})\frac{W\pi^2}{4L^2} \tag{15.19}$$

where $\bar{\theta}$ is average water content, the $D(\theta)$ is known diffusivity function. Eq. (15.19) can be integrated to obtain cumulative infiltration.

Gardner (1959) presented the analytical solution for the second stage of evaporation by using the solution for diffusion by Crank (1956). According to Crank (1956), the weighted mean diffusivity for desorption $(\bar{D}(\theta))$ is

$$\bar{D}(\theta) = \frac{1.85}{(\theta_i - \theta_0)^{1.85}}\int_{\theta_0}^{\theta_i} D(\theta)(\theta_i - \theta)^{0.85}\mathrm{d}\theta \tag{15.20}$$

$\bar{D}(\theta)$ where for sorption process is higher than that for desorption process. The weighing is done differently because in infiltration maximal flux occurs at the wet end of column, where diffusivity is the highest. However, in drying the greatest flux is through the dry end, where diffusivity is the lowest. This is also the reason, why sorption processes are faster as compared to desorption. Gardner (1959) assumed that initial evaporation is infinitely high and soil surface is instantaneously brought to the final stage of drying. Therefore, $e_{pot}\rightarrow\infty$, at t=0 when second stage of drying starts. Using the diffusivity form of Richards' equation and assuming that influence of gravity is negligible, the evaporation rate for a semi-infinite soil column can be given as

$$e=(\theta_i-\theta_0)\left(\frac{\bar{D}(\theta)}{\pi t}\right)^{1/2} \tag{15.21}$$

and cumulative evaporation *(E)* can be given as

$$E=2(\theta_i-\theta_0)\left(\frac{\bar{D}(\theta)t}{\pi}\right)^{1/2} \tag{15.22}$$

Using the sorptivity concept of Philip (1975), Rose (1966) presented the following relationship for evaporation calculation in the second stage of drying:

$$E=S' t^{1/2}+A't \tag{15.23}$$

or

$$e=\frac{1}{2}S't^{-1/2}+A' \tag{15.24}$$

where S' is the soil evaporativity (which is equivalent to soil water sorptivity in infiltration, since the process is drying it can be termed as desorptivity, $LT^{-1/2}$) and A is a constant (LT^{-1} comparable to trasmissivity [see Eq. (14.24)]). The value of S' is positive whereas b is negative. The assumption of a zero flux at the bottom of the profile or at depth L, although simple, implies that in the absence of this condition evaporation will accompany redistribution. This will reduce both the evaporation rates and the duration of the first stage. Solutions of evaporation considering isothermal conditions differ from the nonisothermal condition. The concept of three stages of evaporation does not strictly hold in field conditions (Jackson et al., 1973). The diurnal temperature fluctuations and other atmospheric process largely affect the evaporation rate. When air temperatures are low the upward heat flow is accompanied with water flow. When temperatures are high, the downward heat flow is accompanied with water flow and/or vapor flow. All these effects make sure that the second stage of drying starts well before the moisture content of soil has reached hygroscopic coefficient or the final dry value. Another factor, which can influence evaporation by as much as 50% is the presence of cracks in the soil. The cracks or similar soil inhomogeinities have totally different thermal fields compared to homogeneous soils. Downward vapor flow due to thermal gradients is observed within the cracks of small sizes (Hatano et al., 1988). The cracks may not increase the evaporation rate during the early stage of drying, but can increase the duration of that stage. Cracks can also increase the evaporation rate of subsequent profile controlled drying period (or second stage).

15.4.5 Nonisothermal Evaporation

The isothermal flow equation is assumed to predict the constant and falling rate of evaporation reasonably well. The role of nonisothermal conditions is explained by comparing the solutions of an isothermal process to the solutions of nonisothermal process (Milley, 1984). According to Jackson et al. (1974), in wet soils the thermal and isothermal vapor fluxes are approximately equal and opposite in direction for diurnal variation of temperature. For a dry soil surface layer, the thermal vapor pressure increases the evaporation from soil profile during night. However, neglecting thermal effects over a month introduce only about 1% error.

15.5 MANAGEMENT OF EVAPORATION

Evaporation from bare soil surface needs to be reduced so that moisture status of soil can be maintained at a stage favorable for crop growth and production. The evaporation management can be done by: (i) reducing the total amount of incident radiations or sources of energy responsible for evaporation; (ii) modifying the color of soil by applying amendments and changing the albedo parameters; and (iii) reducing the upward flux of water by either lowering the water table, or decreasing the diffusivity and conductivity of the soil profile. The methods of evaporation reduction from bare soils depend on the stage of drying. The first stage requires modifications, which will alter meteorological conditions of the surroundings. The second stage requires measures, which will change water transmission properties of the soil profile. Covering or mulching the surface with vapor barriers or with reflective materials can reduce the intensities of the incoming radiations and reduce the evaporation in the first stage of drying. A deep tillage may change the variation of diffusivity with changing water content of soil profile and may change the rate at which water can be supplied to the soil surface from underneath for evaporation.

15.5.1 Mulching

Mulch is any material placed on a soil surface primarily to cover the surface for the purpose of reducing evaporation, controlling weeds, and obtaining beneficial changes in soil environment. The other benefits of mulching are: (i) reducing soil erosion; (ii) sequestering carbon; (iii) providing organic matter and plant nutrition; (iv) regulating and moderating soil temperature; (v) increasing earthworm population and improving soil structure; and (vi) reducing soilborne diseases.

Mulches can consist of many different types of materials, such as sawdust, manure, straw, leaves, crop residue, gravels, paper, and plastic sheets, etc. (Fig. 15.7) (Lal, 1991). Paper or plastic mulches, especially light colored, are effective in reducing the effects of meteorological variables, which influence the evaporative demand during the first stage of soil evaporation (Figs. 15.8 and 15.9). Black paper and plastic mulches are effective in weed control (Fig. 15.10). The temperature of the soil under plastic mulch can be 8 to 10°C higher than under straw mulch. Soil thermal regime is a function of the contact coefficient, which is a product of thermal conductivity and volumetric heat capacity of the soil (refer to Chapter 17). A mulched plot with dry crop residue is equivalent to a two-layered profile of which the upper layer has a lower contact coefficient. Therefore, temperature variations in the soil underlying the mulched layer are reduced (Figs. 15.11 and 15.12). High temperature may be beneficial to the crops on temperate regions during germination in spring. However, high temperature during summer and in the tropics may adversely affect the growth of temperature-sensitive crops. Other mulch materials may include preparations of latex, asphalt, oil, fatty acids, and alcohols. These materials can be used as mulches for reducing evaporation from soil surface. Hillel (1976) proposed that uppermost layer of soil be formed by clods or a rough seedbed, which are treated with water proofing materials (e.g., silicones). These waterproof clods act as dry mulch and reduce evaporation and erosion from soil surface.

Vegetative mulch must have sufficient thickness to be effective in reducing evaporation and risks of soil erosion. The porosity and hydraulic conductivity of the vegetative mulches are high, and therefore diffusion or

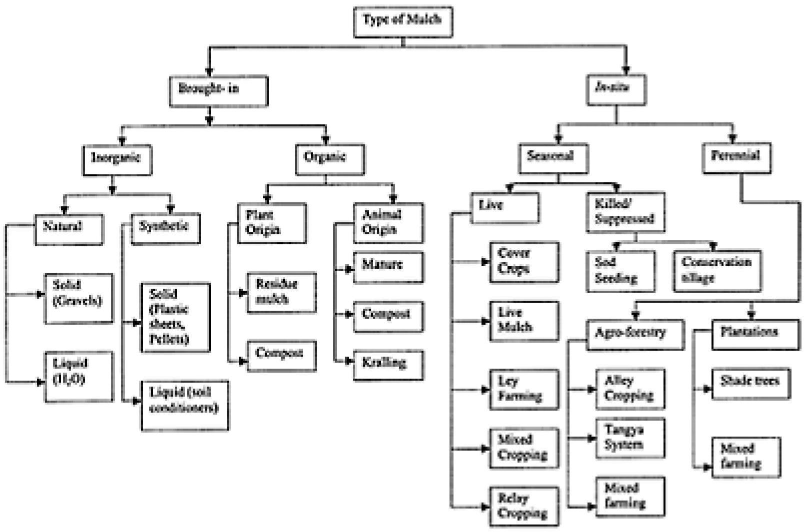

FIGURE 15.7 Type of mulches on the basis of the source of the material. (Modified from Lal, 1991.)

FIGURE 15.8 Clear plastic mulch used on cassava grown at IITA in western Nigeria to conserve soil water.

FIGURE 15.9 Clear plastic mulch used on a ridged seed bed. Note holes in the plastic for seedling emergence.

FIGURE 15.10 Black plastic mulch to conserve water and control weeds in strawberries grown in California.

airflow through the vegetative mulch is also high. A mulch of small thickness may be mostly ineffective. Vegetative mulches are light colored and reflect most of the incident radiations. Therefore, the initial evaporation rate under mulch is generally less. Gravel mulching is a common practice of water conservation, as it enhances the infiltration and simultaneously suppresses evaporation and reduces erosion of soil. Disadvantages of gravel mulch are that gravel cannot be removed from the field after application and can adversely affect future land uses.

15.5.2 Tillage

Among the various soil management practices for weed control and seedbed preparation, tillage is an important technique of soil manipulation. Tillage operations generally result in opening up of soil, changes in structure, loosening of tilled soil, and compaction of soil immediately below the tilled layer (Fig. 15.13) (Lal 1989, 1990). The opening of the topsoil enhances the evaporation from the tilled soil layer. However, the compaction of layers underneath might reduce the upward transmission of water and subsequently make the water availability limiting and reduce evaporation. The reduction of diffusivity in the soil layer also reduces the evaporation. The discontinuity of pore channels due to the

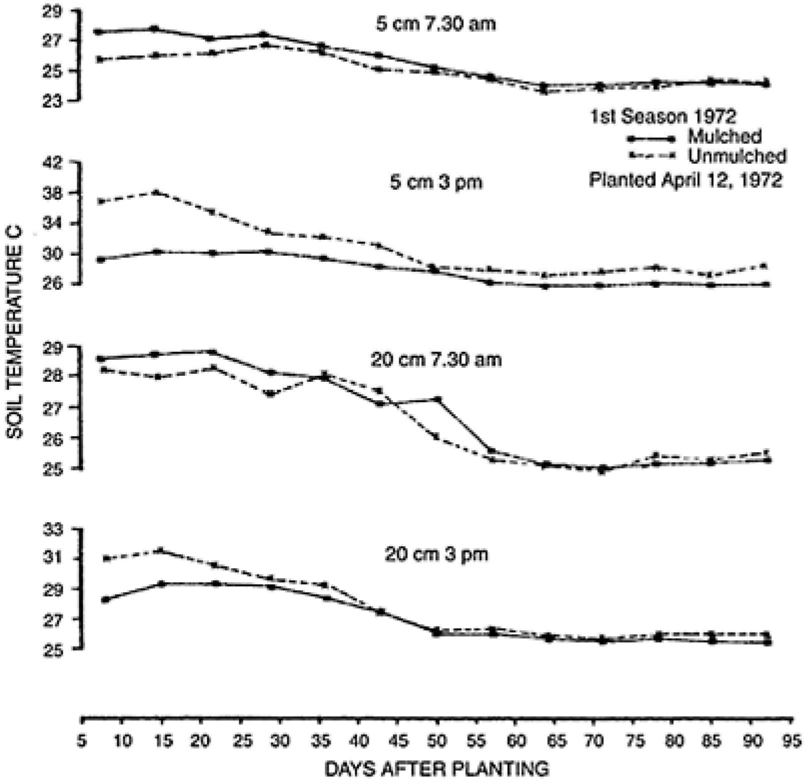

FIGURE 15.11 Effects of mulching on soil temperature under maize. (From Lal, 1974.)

tillage operations does not reduce the upward flow of water and does not reduce the total evaporation. More recent trends have indicated that management practices involving minimum tillage are better for efficient soil management. The tillage is beneficial under two situations: (i) in soils with high swell-shrink capacity and where frequent wetting and drying produces cracks. These cracks are the sources of secondary evaporation from soil. Cultivation may prevent development of or help obliterate cracks, (ii) Tillage eliminates weeds and may reduce the rate of application of herbicides. Burning crop residue and the presence of ash on the soil surface can influence soil temperature by altering albedo and soil moisture regime (Figs. 15.14 and 15.15).

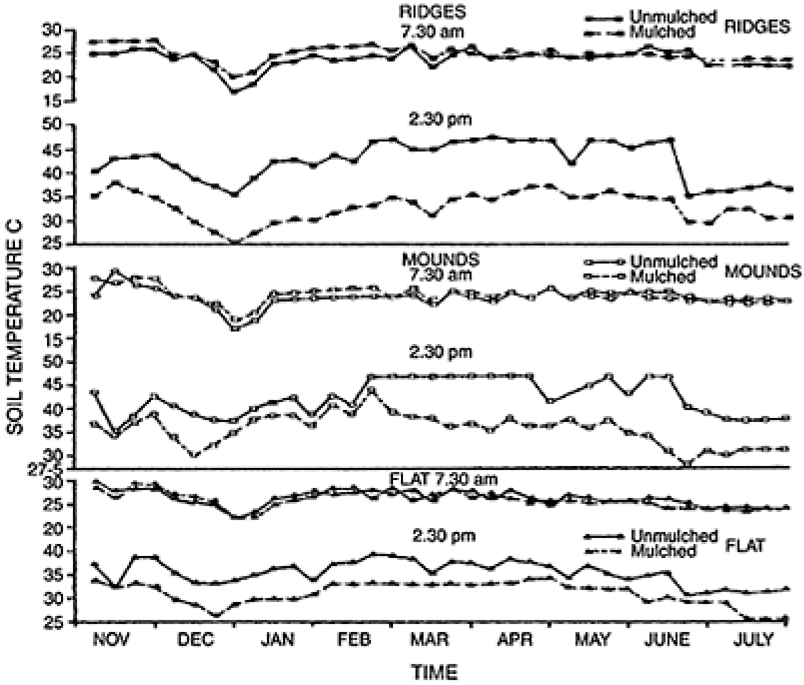

FIGURE 15.12 Effect of mulching and methods of seedbed preparation on soil temperature under yams. (From Lal, 1974.)

15.5.3 Conservation Tillage

Conservation tillage practices leave a high percentage of the residues from previous crops on the soil surface (Fig. 15.16). Plant residues left on the soil surface are effective in reducing evaporation and conserving soil moisture. A conservation tillage practice widely used in semiarid and humid regions is stubble mulching where wheat stubbles or corn stalks from previous crops are uniformly spread over the soil surface. The land is then tilled with special implements, which leave most of the residue on the soil surface. The next crop is planted through the stubble, which results in a healthy environment (temperature, water, and air) for seed germination. No tillage, or zero tillage, is another conservation tillage system that leaves residue on the soil surface and a new crop is planted directly through the residue of the previous crop with no plowing or disking (Lal, 2003).

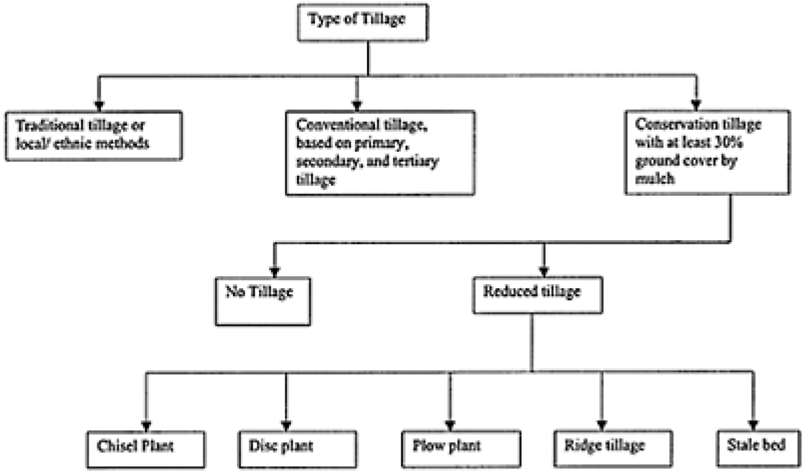

FIGURE 15.13 Types of tillage methods. (Modified from Lal, 1989; 1990.)

FIGURE 15.14 Burning crop residues in a mounded seed bed in Ethiopia. Mounded seedbed alters soil temperature and affects evaporation rate.

FIGURE 15.15 A mulch cap on yam mounds decreases soil temperature and reduces evaporation (right), while ash from crop residue alters albedo and soil temperature.

FIGURE 15.16 No-till farming with crop residue mulch reduces soil evaporation.

Example 15.1

Assume average daily steady state evaporation is 1 cm in a saturated loam soil in a high water table area. Estimate (a) threshold depth beyond which water table must be lowered, (b) water table depth at which evaporation will fall to 20% of potential value, and (c) plot daily evaporation rate with respect to water table depth. Use Eq. (15.10), assuming Aa to be equal to 4.5 cm^2.sec and n=3.

Solution

According to Eq. (15.10)

$$q_{max} = \left(\frac{Aa}{d^n}\right) \Leftrightarrow d = \left(\frac{4.5}{1/86400}\right)^{1/3} = 72.99 = 73\,\text{cm}$$

where d is the maximum depth of water table below the soil surface, which can supply water to maintain a steady flux for evaporation. Hence

(a) Threshold water table depth is 73 cm.
(b) The water table depth ($d_{0.2}$) at which evaporation rate falls by 20% can be calculated from again Eq. (15.10) as follows:

$$d_{0.2} = \left(\frac{4.5}{1 * 0.2/86400}\right)^{1/3} = 124.8 = 125\,\text{cm}$$

(c)

D (cm)	q_{max} (cm)
0–73	1
80	$4.5/80^3$=0.76
90	$4.5/90^3$=0.53
100	0.39
120	0.23

Example 15.2

Consider an infinite sandy loam soil profile, which is initially saturated with water. The initial moisture content of soil is 0.52 cm3 cm^{-3} and final moisture content of 0.2 cm^3 cm^{-3}. If weighted mean diffusivity of soil is 80 cm^2 d^{-1}, calculate evaporation and the evaporation rate for each day during the next 10 days.

Solution

From Eq. (15.21) the evaporation rate *(e)*, and from Eq. (15.22), the cumulative evaporation (E), can be calculated for days 1, 2, 3... 10 as follows:

Mid-day	e (cm d^{-1})	Day	E (cm)
0.5	2.28	1	3.23
1.5	1.32	2	4.57
2.5	1.02	3	5.59
3.5	0.86	4	6.46
4.5	0.76	5	7.22
5.5	0.69	6	7.91
6.5	0.63	7	8.54
7.5	0.59	8	9.13
8.5	0.55	9	9.69
9.5	0.52	10	10.21

PROBLEMS

1. If the composite coefficient Aa is 4.5 cm^2 /s, n=3, potential rate of evaporation is 8 mm/d to what depth must the water table be lowered for reducing evaporation? Also calculate the watertable depth at which the evaporation rate drops by 10%, 30%, and 70% of potential evaporation rate.

2. Assume an infinitely deep, saturated sandy loam soil profile under very high evaporativity. If initial volumetric water content of soil is 0.50, final volumetric water content is 0.10 and weighted mean diffusivity is $2\times10^4 mm^2 d^{-1}$. Calculate the evaporation and evaporation rate, for the next 6 days.

3. If an impermeable layer exists at the end of a uniform wetted soil of depth 1.2 m, initial volumetric water content (θ_o) 0.24, and initial diffusivity $(D(\theta_o))$ $4\times10^4 mm^2 d^{-1}$. If evaporativity is 10 mm/d, calculate evaporation rate during the first 10 days if diffusivity $(D(\theta))$ is given by Eq. (15.17) are assuming B=15, calculate $D(\theta_o)$ for the next 6 days.

4. Briefly outline techniques of regulating soil evaporation and explain the principle of their effectiveness in reducing evaporation.

5. What should be the irrigation strategy in arid environments and why?

REFERENCES

Hatano R., H. Nakamoto, T.Sakuma, and H.Okajima (1988). Evapotranspiration in cracked clay field soil. Soil Sci. Plant Nutr. 34:547–555.

Gardner W.R. (1958). Some steady state solutions of the unsaturated moisture flow equation with application to evaporation from a watertable. Soil Sci. 85(4): 228–232.

Crank J. (1956). The mathematics of diffusion. Oxford University Press, London and New York.

Fisher R.A. (1923). Some factors effecting the evaporation of water from soil. J. Agr. Sci. 13:121–143.

Gardner W.R. (1959). Solutions of the flow equation for the drying of the soils and other porous media. Soil Sci. Soc. Am. Proc. 23:183–187.

Gardner W.R. and D. Hillel (1962). The relation of external evaporative conditions to the drying of soils. J. Geophys. Res. 67:4319–4325.

Gardner W.R. and M.S.Mayhugh (1958). Solutions and tests of the diffusion equation for the movement of water in soil. Soil Sci. Soc. Am. Proc. 22:197–201.

Hillel D. (1976). On the role of soil moisture hysteresis in the suppression of evaporation from bare soil. Soil Sci. 122:309–314.

Hillel D. (1980). Fundamentals of soil physics. Academic Press, New York.

Hillel D. and C.H.M.van Bavel (1976). Dependence of profile water storage on soil hydraulic properties: a simulation model. Soil Sci. Soc. Am. J. 40:807–815.

Jackson R.D., B.A.Kimball, R.J.Reginato, and S.F.Nakayama (1973). Diurnal soil water evaporation: comparison of measured and calculated soil water fluxes. Soil Sci. Soc. Am. Proc. 38:861–866.

Lal R. (1974). Role of Mulching Techniques in Tropical Soil and water management. Tech. Bulletin no. 1, International Institute of Tropical Agriculture, Nigeria.

Lal R. (1989). Conservation tillage for sustainable agriculture. Adv. Agron. 42: 85–197.

Lal R. (1990). Soil erosion in the tropics: Principle and Management. McGraw Hill Book Co., 579 pp.

Lal R. (1991). Soil Structure and sustainability. Journal of sustainable Agriculture. 1(4): 67–92.

Lal R. (2003). Historical development of no-till farming. In R. Lal, P. Hobbs, N. Upof[l], and D.O. Hansen (eds) Sustainable agriculture and the rice wheat system. Marcel Dekker, New York.

Milley P.C.D. (1984). A linear analysis of thermal effects on evaporation from soil. Water Resour. Res. 20:1075–1086.

Moore R.E. (1939). Water conduction from shallow watertable. Hilgardia 12: 383–426.

Pearse J.F., T.R. Oliver, D.M. Newitt (1949). The mechanisms of the drying of solids: Part I. The forces giving rise to movement of water in granular beds during drying. Trans. Inst. Chem. Eng. (London) 27:1–8.

Philip J.R. (1957). Evaporation, moisture and heat fields in soil. J. Meteorol. 14: 354–366.

Ripple C.D., J. Rubin, and T.E.A. van Hylckama (1972). Estimating steady state evaporation rates from bare soils under conditions of high water table. US Geol. Surv. Water-Supply Paper 2019-A.

Rose C.W. (1966). Agriculture Physics. Pergamon Oxford.

Willis O.W. (1960). Evaporation from layered soils in presence of a watertable. Soil Sci. Soc. Am. Proc. 24:239–242.