ESR 320

Reading:
Soil Hydrology

4.6 INFILTRATION

Infiltration is the movement of water (precipitation usually) from the soil surface into the soil. The redistribution of infiltrated water examines the movement of that water in the unsaturated soil zone. The

fraction of precipitation that infiltrates on a global scale is about 76 per cent. However, on a regional or local scale it has a large seasonal and spatial variation, even within a few hectares of catchment area.

4.6.1 Elemental Properties of Soils

The elemental properties of soil in relation to infiltration are:

- Bulk density
- Particle density
- Porosity
- Volumetric water content
- Degree of saturation

Bulk density Bulk density ρ_b or the dry density of a soil is

$$\rho_b = \frac{M_d}{V_t} \tag{4.14}$$

where M_d = dry mass of a soil volume (dried at 105 °C for > 16 h)
and V_t = total volume (original undried)

Typical values of ρ_b are 0.7 kg/m^3 for peats to 1.7 kg/m^3 for sands or loams. Clays are typically about 1.1 kg/m^3.

Particle density Partical density ρ_m is

$$\rho_m = \frac{M_d}{V_d} \tag{4.15}$$

where V_d = dry volume (no air, no water)
typical values for ρ_m are 2.65 kg/m^3 for most soils

Porosity Porosity ϕ is the volume proportion of pore space,

$$\phi = \frac{V_a + V_w}{V_s} = 1 - \frac{\rho_b}{\rho_m} \tag{4.16}$$

where
V_a = volume of air
V_w = volume of water
V_s = volume of solids

Values of porosity range from about 35 to 45 per cent for fine sands to 50 to 55 per cent for clays, with peats at about 80 per cent.

Volumetric water content The water content θ is

$$\theta = \frac{V_w}{V_s} = \frac{M_{wet} - M_{dry}}{\rho_w V_s} \tag{4.17}$$

This is an important soil property and varies from 0 (when dry) to saturation (about 40 per cent for sands) and as we will see it varies over space and time. The most successful methods for determining field soil moisture are the neutron probe, the soil moisture capacitance probe or time domain reflectometry. Details of some of the techniques are found in Shaw (1994).

Degree of saturation The degree of saturation s is the proporation of water containing pores and is a measure of the 'wetness':

$$s = \frac{V_w}{V_a + V_W} = \frac{\theta}{\phi} \tag{4.18}$$

4.6.2 Soil Horizons

From Tables 4.1 and 4.2, it is seen that groundwater contributes only 0.5 per cent of what rivers contribute to the oceans. Also, these tables show that groundwater contains 30.1 per cent of the earth's freshwater supply whereas rivers and lakes contain only 0.266 per cent. Groundwater, while huge in volume, is also almost static with very slow movement in the horizontal direction. The level of the water table, however, rises and falls vertically, depending on climate and soil type. Soils play a major role in what happens to precipitation, as large volumes of water can be held in the soil matrix or none at all, depending on soil texture, porosity, structure, hydraulic conductivity and existing soil moisture. From Table 4.1, it is seen that the freshwater held as soil moisture is almost ten times greater than the freshwater held in rivers.

Figure 4.13 shows an idealized vertical profile through a series of soil layers. The top layer is usually a vegetation of grass, crop or trees, but it may be bare soil. Below this is the litter layer, more easily identifiable in forest areas as composed of dead leaves, bark and other decomposed growth. Below this is the soil proper and is described in *horizons* or layers. The upper or A horizon in mineral soils is generally friable and rich in humus. This layer corresponds to the surface soil (sometimes called the topsoil). It is that part of the soil in which living matter is most abundant and in which the organic matter is most plentiful. Being closest to the surface, it becomes more leached by rainfall than the lower layers. The middle level or B horizon, often called the subsoil, is mainly composed of well-weathered parent material interwoven with roots and micro-organisms. Lying between the A and C horizons, it has some of the properties of both, with fewer living organisms than A but more than C. By comparison with the A horizon, the B horizon has a high content of iron and aluminium oxides, humus or clay that have been partly leached from the A horizon. The lower C horizon is unconsolidated rock material composed of a wide range of stones of different sizes. Below the C horizon is the parent consolidated rock. The depth of each layer varies from millimetres to metres.

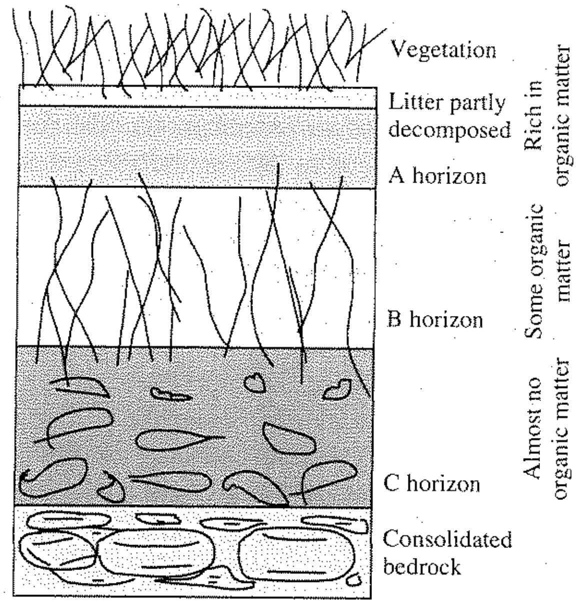

Figure 4.13 Idealized soil section (adapted from Hillel, 1980).

In environmental hydrology, there are two distinct zones above bedrock which may hold and transmit water. They are: the upper unsaturated zone and the lower saturated zone. They are shown in Fig. 4.14. Water movement in the unsaturated zone is more complex than that of the saturated zone. In the latter, the key parameter is hydraulic conductivity or the rate of movement of water. This is readily measured and tends to be reasonably constant. However, in the unsaturated zone the hydraulic conductivity can vary by orders of magnitude within a field, depending primarily on the degree of saturation and the current state of soil suction.

4.6.3 Soil Water Content

Soil moisture is a complex phenomenon well described but poorly quantified. All soils will have a maximum soil moisture magnitude when they are saturated. Similarly, if they are in extreme moisture

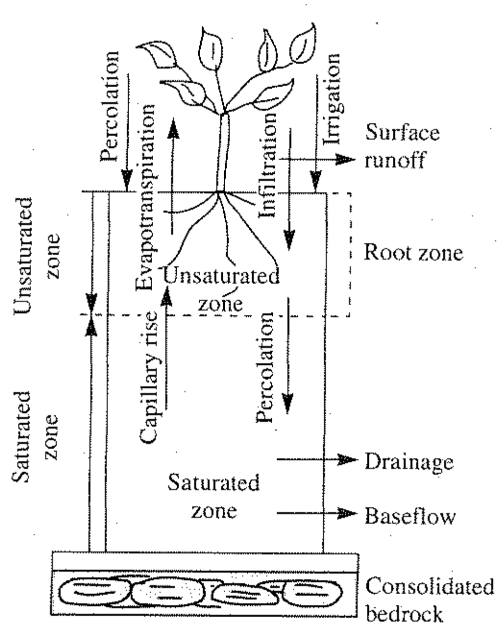

Figure 4.14 Unsaturated/saturated zone (adapted from Bras, 1990).

deficit, their soil moisture will be lowest (not zero). At any point in time, their soil moisture status will vary from close to zero to maximum. It is therefore a dynamic action and responds to the antecedent soil moisture conditions and the current rainfall and solar heat pattern. It is easy enough to quantify, in a vertical profile of a soil, the different levels of soil moisture (i.e. the percentage of moisture content). However, because of continuous activity beneath the surface and above the surface, the fluxes of moisture from one horizon to another are not constant. At times of rainfall, the movement of water in the soil column will be downwards under gravity or upwards towards the water table through capillarity. At times of drought, the direction of water movement will be upward towards the soil surface by capillarity from the groundwater. The fate of rainfall depends largely on:

- Climate zone
- Soil characteristics and
- Soil antecedent moisture conditions

Figure 4.15 is a schematic of water in soil. Within the soil column there are three zones: *aeration*, *capillary* and *groundwater*. The groundwater zone exists below the water table. The capillary zone is the zone through which water will rise through the soil pores by capillary action. The upper zone is the aeration zone where the pores are occupied by air. After rainfall events, the air may be expelled from the pores by hydrostatic pressure to allow the infiltration water to occupy the pores. The soil column is sometimes subdivided into two zones, the unsaturated upper zone and the lower saturated zone. The unsaturated zone is the subject of intense research by hydrologists and studies on hillslope hydrology help to elucidate the physics of unsaturated zone flow.

With respect to soil water, it occupies three different phases in a soil matrix. These are:

- Pore water
- Hygroscopic or adsorbed water
- Absorbed water

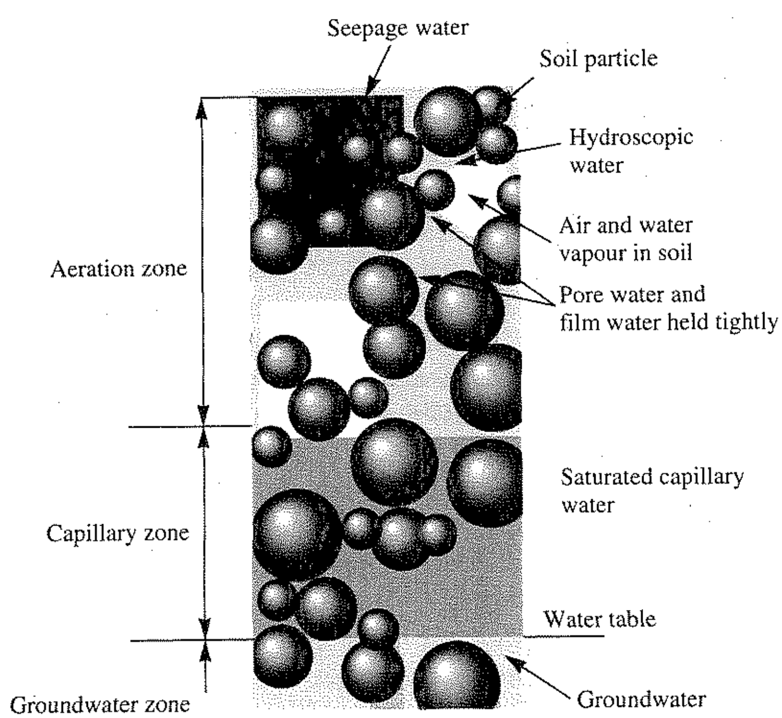

Figure 4.15 Water in the soil (adapted from Weisner, 1970).

Water moves in rivers due to a slope or gradient in its water surface. The steeper the gradient the faster the flow. As lake surfaces have little gradient, water flows slowly. In soils and aquifers, water also moves if it has a gradient, although at several orders of magnitude slower than river flow. This gradient is called the hydraulic gradient. In rivers, water always flows practically horizontally (assuming one-dimensional flow). However, beneath the soil surface, water may flow in either the x, y or z direction. The way water flows in soils is dependent on soil type and its current moisture status. For instance, in summer time, a sandy soil matrix may be dried out, and if rain falls, this rainfall will move vertically down through the soil to help fill the soil pores with water. However, if the soil moisture status is close to field capacity, then the principal direction of movement of water may be in the horizontal direction. This direction is usually along the gradient of the water surface line, which may follow the topographic slope. The rate at which water moves is called its *hydraulic conductivity*. It is easy to evaluate flow behaviour in a saturated porous medium. This is usually the case in aquifers. However, there are times when the soil status is also unsaturated. There may still be water movement in the soil, but it may be restricted due to excessive soil suction.

Darcy's law states:

$$q = -Ki = K\frac{dh}{dz} \tag{4.19}$$

where q = the Darcy flux, $m^3/m^2.s$

i = the hydraulic gradient, $= dh/dz$, m/m

K = the hydraulic conductivity, m/s

Usually h is the height relative to a datum, but for unsaturated flow the total head is

$$h = \Psi + z \tag{4.20}$$

where Ψ = the suction head

The suction head, responsible for holding water to the surface of soil particles in unsaturated flow, becomes significant as soil moisture decreases. The variation in the soil column of hydraulic conductivity and soil suction head is shown for different moisture contents in Fig. 4.17. Soil suction or soil tension is measured using tensiometers in the field.

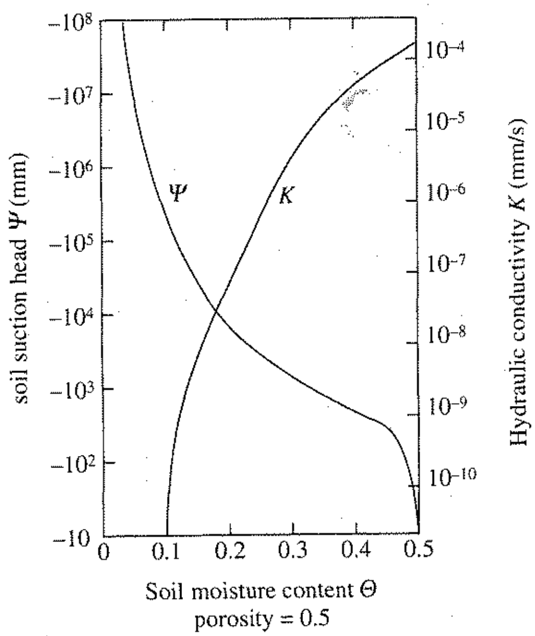

Figure 4.17 Variation of soil suction head ψ and hydraulic conductivity K with moisture content Θ for Yolo light clay (adapted from Raudkivi, 1979).

The pore water is by far the greatest volume of soil water and is the easiest to expel. The hygroscopic water is adsorbed on to the surface of the grain particle and is held there by surface tension forces. The absorbed water (internal to each grain) requires the removal of the pore water and hygroscopic water before it can be dried out. Sandy soils have large pores and thus can be dried out easily. However, clay particles have small pores (but higher porosity than sand) and small particles with intense hygroscopic activity and require high suction forces to break the hygroscopic surface tension forces.

The phenomenon of soil suction is illustrated by placing a drop of water on to a dry soil particle. The water is quickly drawn into the soil until it is saturated and then a thin film attaches to the perimeter of the soil grains. This hygroscopic film is held with intense surface tension forces. These forces are expressed in bars, i.e. 1 bar is the pressure equivalent to 10.23 m height of water column.

Field capacity and *wilting point* are further soil moisture parameters most often used in agriculture soil studies. After the soil has been saturated and the excess water drained away, the soil is then at field capacity. Vegetation extracts moisture from soil until it cannot do so any more. At this point wilting occurs and the moisture content is known as the wilting point. Figure 4.16 shows a general relationship between soil moisture and soil texture.

4.6.4 Movement of Water in Soil and Hydraulic Conductivity

Water movement occurs in soils under three distinct conditions:

- Saturated flow
- Unsaturated flow and
- Vapour phase flow

All water movement beneath the water table is of the saturated flow type. However, a soil may be temporarily saturated above the water table and this occurs if *all* the pores are filled with water. From a two-dimensional perspective, the movement of water may be vertically downwards or laterally as interflow. The rate of movement depends on hydraulic conductivity of the soil. Unsaturated flow takes place in response to gravity or moisture gradient. Once field capacity exists, capillary action draws the water upwards to the roots and vegetation. After wetting of soils, water flows downwards due to gravity. The mechanism of water movement in unsaturated flow is from pore to pore. Water may exist in the vapour phase in the pores of a soil and be drawn upward to evaporate. The rate of movement depends on the temperature gradient, relative humidity, pore size and pore continuity and the amount of available water. So it is important to conceive of evaporation also from the depths of a soil column.

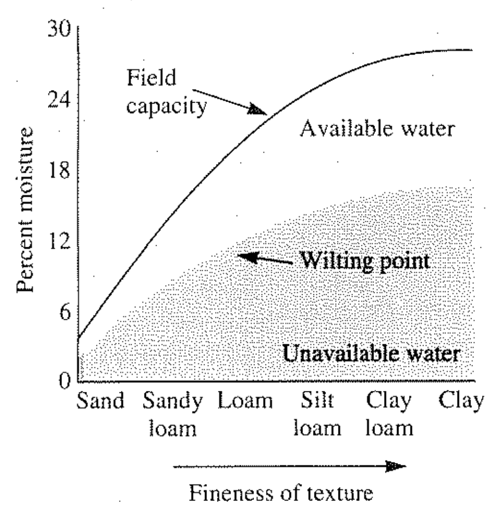

Figure 4.16 General relationship between soil moisture characteristics and soil texture.

4.6.5 Soil Moisture Deficit

Soil moisture deficit (SMD) is a term commonly used in agricultural engineering. When the soil moisture is below field capacity, it is said to have a soil moisture deficit. When the soil is saturated it does not have a soil moisture deficit. SMD is a quantifiable parameter and is related to rainfall magnitude, degree of moisture in soil and evapotranspiration. A catchment loses water at rates greater or less than PE (potential evaporation), depending on whether the soil moisture is above or below field capacity. ET (actual evapotranspiration) is less than PE when the vegetation cannot abstract water from the soil. After a rainfall event (if the soil is saturated), it will hold no more water, thus producing runoff. The soil in this case will continue to 'give up' water to the vegetation until a temporary equilibrium stage is arrived at, when $ET = PE$, i.e. field capacity. At this stage $SMD = 0$. As the soil dries out, SMD increases and ET decreases. The magnitude of SMD and ET varies. As SMD increases further, ET becomes less and at wilting stage SMD is greatest and ET negligible. It is important to note that SMD is a cumulative number, depending on the previous months' SMD.

Figure 4.18 is an idealized and simplified schematic of the time sequence of soil moisture related to rainfall and PE for an annual cycle in the northern temperate zone. Three vegetation types are sketched: grass, shrubs and trees. Each has a different root depth, depicted as three distinct horizontal layers.

In spring when $PE > P$ (precipitation), the soil enters an SMD, first at the surface layers. As the spring goes into summer SMD penetrates deeper until all the root zones (trees included) are in SMD. In the autumn, $P > PE$ and the upper soil layers become replenished with water, while the lower layers are

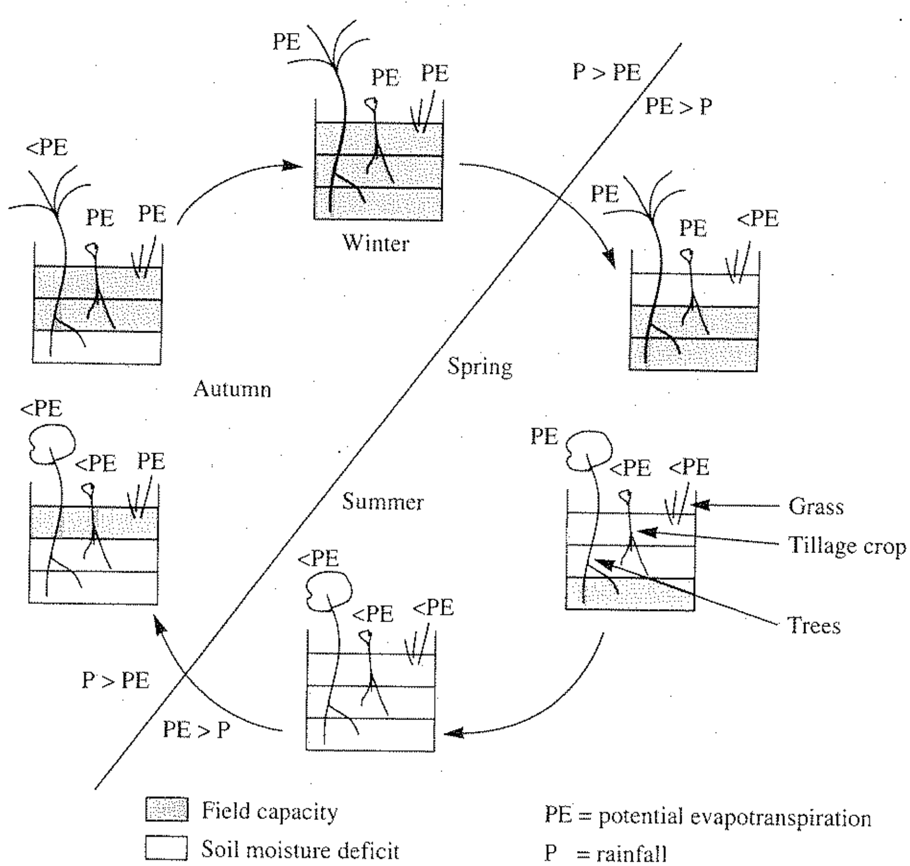

Figure 4.18 Idealized annual moisture cycle for three vegetational types. (Adapted from Bedient and Huber, 1988.)

still in SMD. At this time, the movement of water is vertically downward in the soil column. As autumn moves into winter, the depth of water replenishment deepens until all soil layers are filled with water and there is no SMD at any depth. It can be visualized that in springtime the direction of movement of soil water is downward while the reverse holds in the autumn.

Knowing the soil moisture deficit is important for agriculture and hydrology. During times of high soil moisture deficits catchments tend to be less susceptible to producing flood events. A parameter used in the United Kingdom and Ireland from the *Flood Studies Report* (*FSR*) (NERC, 1975) is the effective mean soil moisture deficit in mm. For instance, some areas in the south-west of Ireland have an EMSMD of 2 mm by comparison with values of 16 mm in East Anglia. The former is susceptible to flooding while the latter is not.

4.6.6 Simple Infiltration Models

Infiltration is the mechanism of water movement into soil under gravity and capillary forces. It was suggested by Horton (1933) that the rate of infiltration of rainfall into soil decreases exponentially with time during a rain event. Some hours into a rain event, the infiltration rate may be close to zero as the soil becomes saturated. The concept of infiltration, as seen by Horton, is shown schematically in Fig. 4.19.

Where $i > f$ at all times, Horton's empirical equation is

$$f = f_c + (f_0 - f_c)e^{-kt} \tag{4.21}$$

where

f_0 = initial infiltration rate

f = infiltration rate at any time, mm/h

f_c = final infiltration rate

k = empirical constant

i = rainfall rate, mm/h

Often f_c is referred to as infiltration potential. In Horton's equation k is a function of surface texture, where k decreases with increasing vegetation. Also, f_c and f_0 are functions of soil type and cover. Figure 4.20 indicates the variation of f with soil cover, rain intensity and topographic slope. Low rain intensity will have a higher proportion of its rainfall infiltrate than a high-intensity event, as shown in Fig. 4.20(*b*).

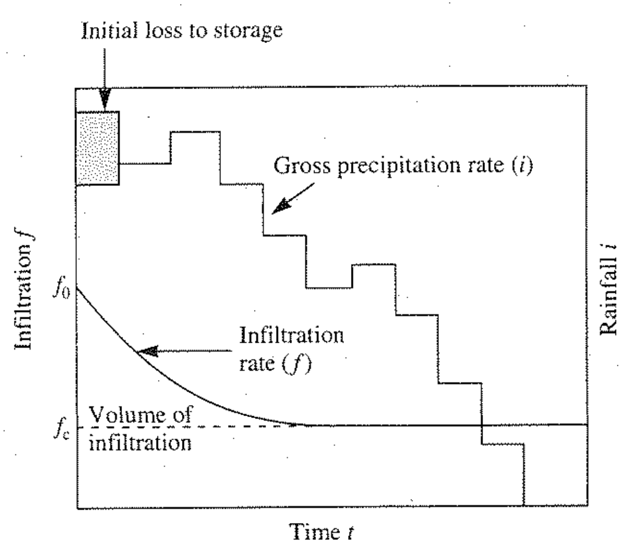

Figure 4.19 Horton's infiltration concept.

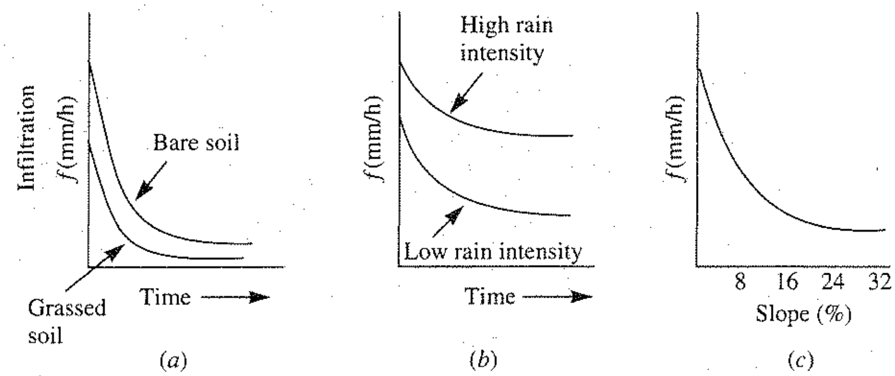

Figure 4.20 Schematic of variation of infiltration capacity.

Example 4.5 Given $f_0 = 100$ mm/h, $k = 0.35$/h and $f_c = 10$ mm/h, find f at $t = 1$, 2 and 6 h and also F_{total} (cumulative infiltration)

Solution From the Horton equation,

$$f = f_c + (f_0 - f_c)e^{-kt}$$
$$f = 10 + 90e^{-0.35t}$$
$$\text{at } 1 \text{ h} \rightarrow f = 73.4 \text{ mm/h}$$
$$2 \text{ h} \rightarrow f = 54.6 \text{ mm/h}$$
$$6 \text{ h} \rightarrow f = 21.0 \text{ mm/h}$$

$$\text{Cumulative infiltration after 6 h} = F_{\text{total}} = \int f\,dt = 285 \text{ mm}$$

The ϕ *index method* of infiltration is sometimes used. This is the simplest method and is measured by finding the loss difference between total precipitation and surface runoff (measured on the stream hydrograph). The infiltration is assumed uniform over the full duration of the rainfall event. It is shown schematically in Fig. 4.21. When considering rainfall events of, say, less than a day, the computation of gross rainfall and effective rainfall will usually ignore evapotranspiration (ET). Longer term events of greater than about two weeks will take ET into account.

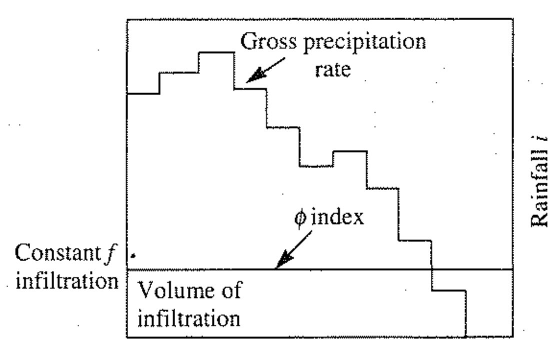

Figure 4.21. The ϕ index infiltration concept.

The reader is referred to Dingman (1994) and Bras (1990) for a more serious mathematical treatment of infiltration.

4.7 EVAPORATION AND EVAPOTRANSPIRATION

Evaporation is the process by which water is returned to the atmosphere, from the liquid or solid state into the vapour state. Transpiration into the atmosphere also occurs through the leafy parts of plants and trees. Because these processes are so interlinked, the 'all in' term used is *evapotranspiration*. In temperate climates, forest land has about twice the evapotranspiration rates of grassland (typically 40 to 70 per cent of total annual rainfall, by comparison with 20 to 40 per cent for grassland, as shown in some British research). This means, of course, that less water infiltrates the soil or becomes part of runoff. About 70 per cent of the mean annual rainfall in the United States is returned to the atmosphere via evapotranspiration, as shown in Table 4.12. In areas of scarce water supplies forest development with higher evapotranspiration losses can reduce the water yield to rivers and lakes.

The global (land + oceans) annual average precipitation of about 1 m is of course equal to the evaporation. As the land surface of the earth evaporates approximately 70 per cent of its rainfall, allowing the remaining 30 per cent to become runoff, then it is clear that on the ocean surface of the earth there is more evaporation than precipitation (Brutsaert, 1982). Figure 4.22 shows the latitudinal distribution of global precipitation and evapotranspiration. Figure 4.23 shows the relationship between evaporation, precipitation and interception for the Amazonian rainforest, after Shuttleworth (1988).

Three types of evaporation/evapotranspiration are:

- Evaporation from a lake surface, E_0
- Actual evapotranspiration, ET
- Potential evapotranspiration, PE

E_0 is the evaporation from a lake or open water body surface. ET is very complex as it includes the evaporation and transpiration from a land surface, vegetated or otherwise. This means that ET for any one surface type will vary depending on its present soil moisture status and is therefore a dynamic parameter. ET will be greater for a soil which is saturated than if it were unsaturated. In an effort to simplify ET the term PE was introduced, which is potential evapotranspiration. This is the evapotranspiration from a soil matrix when the soil moisture is held constant at field capacity. This is achieved by spraying regularly. Meteorological data will normally give values of E_0 and PE but not ET. The latter is usually only determined in catchment research projects, when measurements of the radiation and heat budgets are taken.

Table 4.12 Precipitation–evapotranspiration from continents

Continent	Precipitation (mm/yr)	Evapotranspiration (mm/yr)	Runoff (mm/yr)
Europe	657	375	282
Asia	696	420	276
Africa	695	582	114
Australia	447	420	27
North America	645	403	242
South America	1564	946	618
Antarctica	169	28	141
Total land	746	480	266

Data from Baumgartner and Reichel, 1975

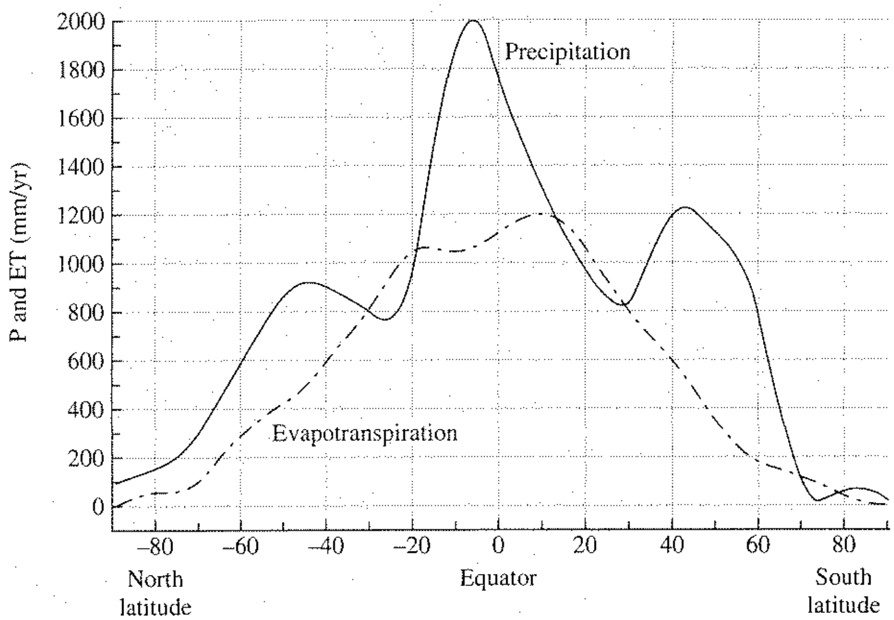

Figure 4.22 Latitudinal distribution of global precipitation and evapotranspiration.

Figure 4.24 shows a comparison of rainfall and potential evaporation at a number of sites in Ireland, averaged over the period 1961 to 1990. The data for this figure is shown in Table 4.13. This is a typical rainfall/potential evaporation plot for the temperate climate. PE exceeds rainfall in summer and as such the soils require artificial watering. In winter, rainfall exceeds PE and this can lead to high runoff, with potential for streamwater pollution from agricultural activities such as slurry spreading.

Two of the factors causing evaporation from any surface are:

- The availability of a supply of heat energy to provide the latent heat of vaporization
- The availability of a transport process to move the water vapour away from the surface, i.e. wind

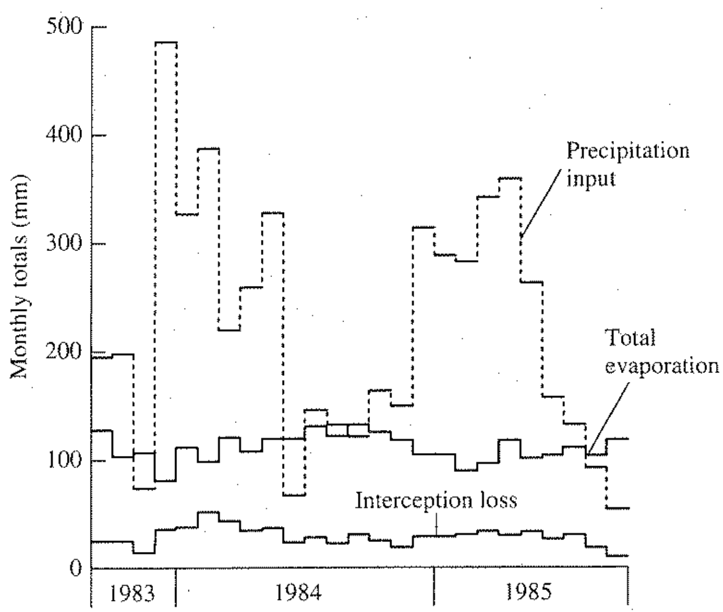

Figure 4.23 Monthly values for precipitation, total evaporation and the interception component for the 25-month period in the Amazonian rainforest (adapted from Shuttleworth, 1988. Reprinted by permission of The Royal Society).

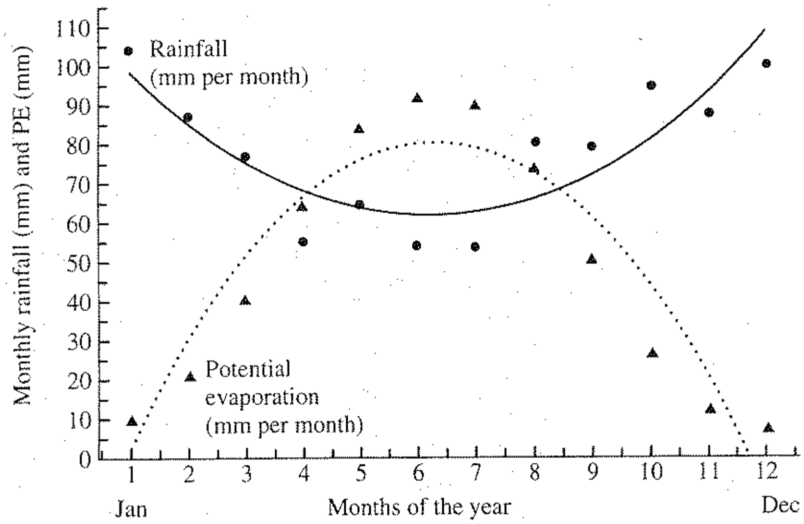

Figure 4.24 Monthly rainfall and potential evaporation trends, Ireland (prepared from Table 4.13).

Solar radiation provides the heat source while wind, along with a vertical humidity gradient, supplies the transport source. These are shown schematically in Fig. 4.25.

Evaporation from a lake surface depends on:

- Available energy as heat
- Solar radiation and more specifically net radiation
- The temperature of the air and water surface
- The wind speed
- Saturation vapour deficit ($e_0 - e_z$)

Table 4.13 Rainfall and potential evaporation at four sites in Ireland from 1961 to 1990

	South Coast*		East Coast†		Midland‡		West§	
Month	Rain	PE	Rain	PE	Rain	PE	Rain	PE
January	104	10	69	9	93	2	121	3
February	87	21	50	21	66	14	83	13
March	77	40	54	39	72	30	96	28
April	55	64	51	61	59	53	62	49
May	64	84	55	83	72	74	78	69
June	54	92	56	94	66	82	71	75
July	53	90	50	91	62	78	64	68
August	80	74	71	73	81	61	97	54
September	79	50	67	50	86	39	104	33
October	95	26	70	25	94	16	124	14
November	88	12	65	10	88	2	118	2
December	100	7	76	5	94	1	124	1
Year average	935	570	732	561	934	446	1143	408

* Cork—Roches Point on South Coast.
† Dublin—Dublin Airport, 3 km from East Coast.
‡ Mullingar—Central Ireland.
§ Claremorris—West of Ireland.
Data from Irish Meteorological Office, 1993

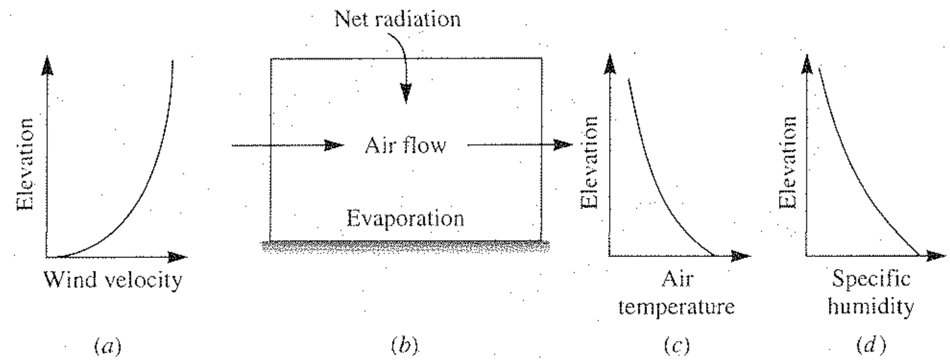

Figure 4.25 Concept of evaporation from an open water surface (adapted from Chow *et al.*, 1988).

Figure 4.25 shows the inputs and outputs to a control volume of 'evaporating air' and those natural processes—radiation, temperature and wind speed—that effect evaporation. The reader is referred to Chow *et al.* (1988), Bras (1990) and Brutsaert (1982) for further details.

At the earth's surface, evaporation is the connecting link between the water budget and the energy budget (Brutsaert, 1982). The most simplified energy budget is represented by

$$R_n = LE + H + G \tag{4.5}$$

where R_n = specific flux of incoming radiation, k cal/m^2 y or W/m^2
L = latent heat of evaporation, J/m^3
E = rate of evaporation, m/y
H = specific flux of sensible heat into the atmosphere, k cal/m^2 y or W/m^2 (i.e. the energy used in heating the 'air')
G = specific flux of heat into the earth, W/m^2

Figure 4.26 shows the diurnal variation of the energy budget over an irrigated wet bare soil in Davis, California, in August 1993. The peak net radiation (post-midday) is about 630 W/m^2. The energy used up in evaporation, LE, peaks at about 400 W/m^2. The sensible heat, H, is then only less than 100 W/m^2. This is to be expected for clear skies over a wet soil. For instance, over dry desert conditions, we might expect about 10 to 30 W/m^2 for LE and 300 to 400 W/m^2 for H. It is important to realize that evaporation

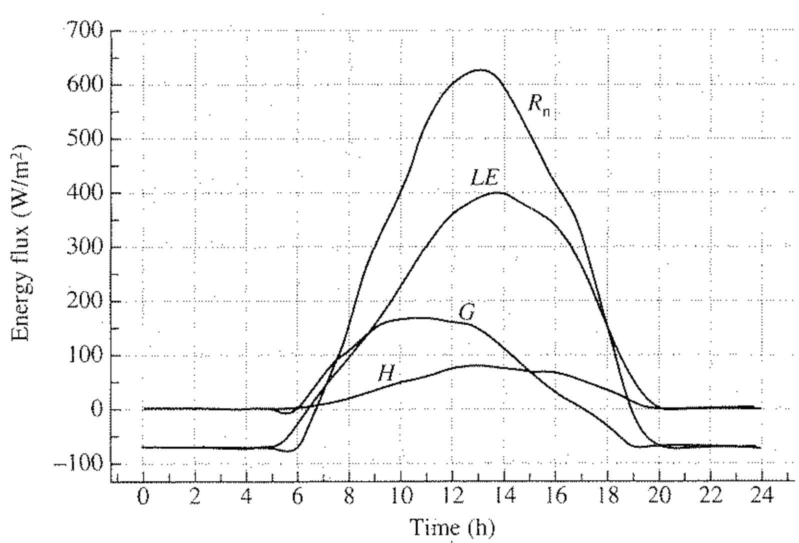

Figure 4.26 Energy balance 23 August 1993, Davis, California, on a wet irrigated bare soil.

can only occur if there is moisture to evaporate. In Fig. 4.26 there was significant water available in the top layers of the soil to produce evaporation with 70 per cent of the net radiation going to evaporation use.

On a global scale, on an annual basis where $G \to 0$, $R_n \approx 70$–80 kcal/cm^2 y. Over land, $LE \approx H$ and over the oceans $LE \approx 90$ per cent of R_n. This suggests that on a global scale $LE \approx 80$ per cent of R_n. While significant spatial and temporal variations of the constituents of the energy budget occur, the above figures emphasize the overwhelming importance of the evapotranspiration process to the overall heat budget and also to the water budget (Brutsaert, 1982). Typically, the annual cyclic behaviour of evaporation parallels that of the cycle of solar radiation and daily air temperatures for land surfaces and shallow water bodies. However, deep waters show peaks in the fall of the year by comparison with peaks in the summer for shallow lakes. Also deep water bodies show minimum evaporation in the spring and while shallow water bodies show minimum in winter (like land). The daily cycle of evaporation follows the cycle of temperature over land and also over water.