Chapter
3

Abstractions From Precipitation

3.1 INTRODUCTION

In Engineering Hydrology runoff due to a storm event is often the major subject of study. All abstractions from precipitation, viz. those due to evaporation, transpiration, infiltration, surface detention and storage, are considered as losses in the production of runoff. Chief components of abstractions from precipitation, knowledge of which are necessary in the analysis of various hydrologic situations, are described in this chapter.

Evaporation from water bodies and soil masses together with transpiration from vegetation is termed as *evapotranspiration*. Various aspects of evaporation from water bodies and evapotranspiration from a basin are discussed in detail in Secs 3.2 through 3.11. Interception and depression storages, which act as 'losses' in the production of runoff, are discussed in Secs 3.12 and 3.13. Infiltration process, which is a major abstraction from precipitation and an important process in groundwater recharge and in increasing soil moisture storage, is described in detail in Secs 3.14 through 3.19.

A: Evaporation

3.2 EVAPORATION PROCESS

Evaporation is the process in which a liquid changes to the gaseous state at the free surface, below the boiling point through the transfer of heat energy. Consider a body of water in a pond. The molecules of water are in constant motion with a wide range of instantaneous velocities. An addition of heat causes this range and average speed to increase. When some molecules possess sufficient kinetic energy, they may cross over the water surface. Similarly, the atmosphere in the immediate neighbourhood of the water surface contains water molecules within the water vapour in motion and some of them may penetrate the water surface. The net escape of water molecules from the liquid state to the gaseous state constitutes evaporation. Evaporation is a cooling process in that the latent heat of vaporization (at about 585 cal/g of evaporated water) must be provided by the water body. The rate of evaporation is dependent on (i) the vapour pressures at the water surface and air above, (ii) air and water temperatures, (iii) wind speed, (iv) atmospheric pressure, (v) quality of water, and (vi) size of the water body.

Vapour Pressure

The rate of evaporation is proportional to the difference between the saturation vapour pressure at the water temperature, e_w and the actual vapour pressure in the air, e_a. Thus

$$E_L = C(e_w - e_a) \tag{3.1}$$

where E_L = rate of evaporation (mm/day) and C = a constant; e_w and e_a are in mm of mercury. Equation (3.1) is known as *Dalton's law of evaporation* after John Dalton (1802) who first recognised this law. Evaporation continues till $e_w = e_a$. If $e_w > e_a$ condensation takes place.

Temperature Other factors remaining the same, the rate of evaporation increases with an increase in the water temperature. Regarding air temperature, although there is a general increase in the evaporation rate with increasing temperature, a high correlation between evaporation rate and air temperature does not exist. Thus for the same mean monthly temperature it is possible to have evaporation to different degrees in a lake in different months.

Wind Wind aids in removing the evaporated water vapour from the zone of evaporation and consequently creates greater scope for evaporation. However, if the wind velocity is large enough to remove all the evaporated water vapour, any further increase in wind velocity does not influence the evaporation. Thus the rate of evaporation increases with the wind speed up to a critical speed beyond which any further increase in the wind speed has no influence on the evaporation rate. This critical wind-speed value is a function of the size of the water surface. For large water bodies high-speed turbulent winds are needed to cause maximum rate of evaporation.

Atmospheric Pressure Other factors remaining same, a decrease in the barometric pressure, as in high altitudes, increases evaporation.

Soluble Salts When a solute is dissolved in water, the vapour pressure of the solution is less than that of pure water and hence causes reduction in the rate of evaporation. The percent reduction in evaporation approximately corresponds to the percentage increase in the specific gravity. Thus, for example, under identical conditions evaporation from sea water is about 2–3% less than that from fresh water.

Heat Storage in Water Bodies Deep water bodies have more heat storage than shallow ones. A deep lake may store radiation energy received in summer and release it in winter causing less evaporation in summer and more evaporation in winter compared to a shallow lake exposed to a similar situation. However, the effect of heat storage is essentially to change the seasonal evaporation rates and the annual evaporation rate is seldom affected.

3.3 EVAPORIMETERS

Estimation of evaporation is of utmost importance in many hydrologic problems associated with planning and operation of reservoirs and irrigation systems. In arid zones, this estimation is particularly important to conserve the scarce water resources. However, the exact measurement of evaporation from a large body of water is indeed one of the most difficult tasks.

The amount of water evaporated from a water surface is estimated by the following methods: (i) using evaporimeter data, (ii) empirical evaporation equations, and (iii) analytical methods.

Types of Evaporimeters

Evaporimeters are water-containing pans which are exposed to the atmosphere and the loss of water by evaporation measured in them at regular intervals. Meteorological data, such as humidity, wind movement, air and water temperatures and precipitation are also noted along with evaporation measurement.

Many types of evaporimeters are in use and a few commonly used pans are described here.

CLASS A EVAPORATION PAN It is a standard pan of 1210 mm diameter and 255 mm depth used by the US Weather Bureau and is known as Class A Land Pan. The depth of water is maintained between 18 cm and 20 cm (Fig. 3.1). The pan is normally made of unpainted galvanised iron sheet. Monel metal is used where corrosion is a problem. The pan is placed on a wooden platform of 15 cm height above the ground to allow free circulation of air below the pan. Evaporation measurements are made by measuring the depth of water with a hook gauge in a stilling well.

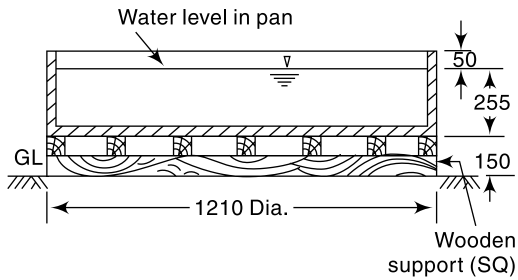


Fig. 3.1 U.S. Class *A* Evaporation Pan

ISI STANDARD PAN This pan evaporimeter specified by IS: 5973–1970, also known as modified Class A Pan, consists of a pan 1220 mm in diameter with 255 mm of depth. The pan is made of copper sheet of 0.9 mm thickness, tinned inside and painted white outside (Fig. 3.2). A fixed point gauge indicates the level of water. A calibrated cylindrical measure is used to add or remove water maintaining the water level in the pan to a fixed mark. The top of the pan is covered fully with a hexagonal wire netting of galvanized iron to protect the water in the pan from birds. Further, the presence of a wire mesh makes the water temperature more uniform during day and night. The evaporation from this pan is found to be less by about 14% compared to that from unscreened pan. The pan is placed over a square wooden platform of 1225 mm width and 100 mm height to enable circulation of air underneath the pan.

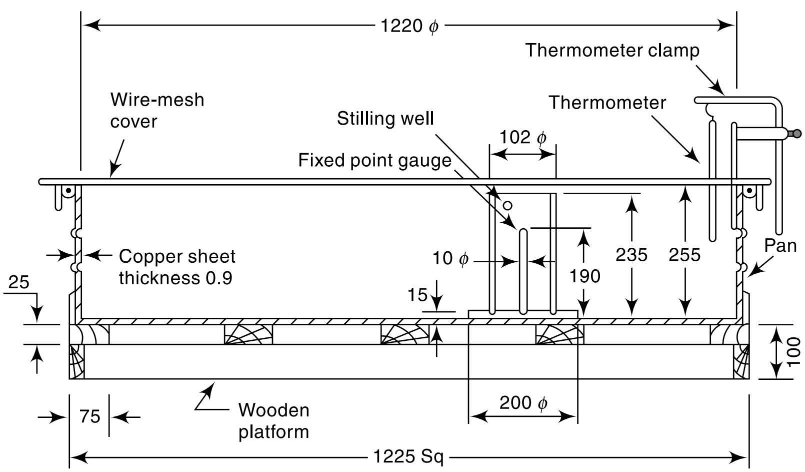


Fig. 3.2 ISI Evaporation Pan

Colorado Sunken Pan This pan, 920 mm square and 460 mm deep is made up of unpainted galvanised iron sheet and buried into the ground within 100 mm of the top (Fig. 3.3). The chief advantage of the sunken pan is that radiation and aerodynamic characteristics are similar to those of a lake. However, it has the following disadvantages: (i) difficult to detect leaks, (ii) extra care is needed to keep the surrounding area free from tall grass, dust, etc., and (iii) expensive to instal.

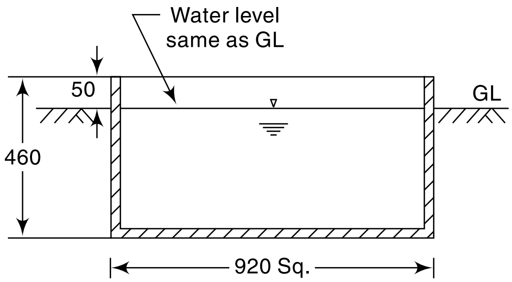


Fig. 3.3 Colorado Sunken Evaporation Pan

US Geological Survey Floating Pan With a view to simulate the characteristics of a large body of water, this square pan (900 mm side and 450 mm depth) supported by drum floats in the middle of a raft (4.25 m × 4.87 m) is set afloat in a lake. The water level in the pan is kept at the same level as the lake leaving a rim of 75 mm. Diagonal baffles provided in the pan reduce the surging in the pan due to wave action. Its high cost of installation and maintenance together with the difficulty involved in performing measurements are its main disadvantages.

Pan Coefficient C_p Evaporation pans are not exact models of large reservoirs and have the following principal drawbacks:

1. They differ in the heat-storing capacity and heat transfer from the sides and bottom. The sunken pan and floating pan aim to reduce this deficiency. As a result of this factor the evaporation from a pan depends to a certain extent on its size. While a pan of 3 m diameter is known to give a value which is about the same as from a neighbouring large lake, a pan of size 1.0 m diameter indicates about 20% excess evaporation than that of the 3 m diameter pan.
2. The height of the rim in an evaporation pan affects the wind action over the surface. Also, it casts a shadow of variable magnitude over the water surface.
3. The heat-transfer characteristics of the pan material is different from that of the reservoir.

In view of the above, the evaporation observed from a pan has to be corrected to get the evaporation from a lake under similar climatic and exposure conditions. Thus a coefficient is introduced as

$$\text{Lake evaporation} = C_p \times \text{pan evaporation}$$

in which C_p = pan coefficient. The values of C_p in use for different pans are given in Table 3.1.

Table 3.1 Values of Pan Coefficient C_p

S.No.	Types of pan	Average value	Range
1.	Class A Land Pan	0.70	0.60–0.80
2.	ISI Pan (modified Class A)	0.80	0.65–1.10
3.	Colorado Sunken Pan	0.78	0.75–0.86
4.	USGS Floating Pan	0.80	0.70–0.82

EVAPORATION STATIONS It is usual to instal evaporation pans in such locations where other meteorological data are also simultaneously collected. The WMO recommends the minimum network of evaporimeter stations as below:

1. Arid zones—One station for every 30,000 km^2,
2. Humid temperate climates—One station for every 50,000 km^2, and
3. Cold regions—One station for every 100,000 km^2.

Currently, about 220 pan-evaporimeter stations are being maintained by India Meteorological Department.

A typical hydrometeorological station contains the following: Ordinary raingauge; Recording raingauge; Stevenson Box with maximum and minimum thermometer and dry and wet bulb thermometers; wind anemometer, wind direction indicator, sunshine recorder, thermohydrograph and pan evaporimeter.

3.4 EMPIRICAL EVAPORATION EQUATIONS

A large number of empirical equations are available to estimate lake evaporation using commonly available meteorological data. Most formulae are based on the Dalton-type equation and can be expressed in the general form

$$E_L = Kf(u)(e_w - e_a) \qquad (3.2)$$

where E_L = lake evaporation in mm/day, e_w = saturated vapour pressure at the water-surface temperature in mm of mercury, e_a = actual vapour pressure of over-lying air at a specified height in mm of mercury, $f(u)$ = wind-speed correction function and K = a coefficient. The term e_a is measured at the same height at which wind speed is measured. Two commonly used empirical evaporation formulae are:

MEYER'S FORMULA (1915)

$$E_L = K_M(e_w - e_a)\left(1 + \frac{u_9}{16}\right) \qquad (3.3)$$

in which E_L, e_w, e_a are as defined in Eq. (3.2), u_9 = monthly mean wind velocity in km/h at about 9 m above ground and K_M = coefficient accounting for various other factors with a value of 0.36 for large deep waters and 0.50 for small, shallow waters.

ROHWER'S FORMULA (1931) Rohwer's formula considers a correction for the effect of pressure in addition to the wind-speed effect and is given by

$$E_L = 0.771(1.465 - 0.000732\, p_a)(0.44 + 0.0733\, u_0)\,(e_w - e_a) \qquad (3.4)$$

in which E_L, e_w, and e_a are as defined earlier in Eq. (3.2),

p_a = mean barometric reading in mm of mercury

u_0 = mean wind velocity in km/h at ground level, which can be taken to be the velocity at 0.6 m height above ground.

These empirical formulae are simple to use and permit the use of standard meteorological data. However, in view of the various limitations of the formulae, they can at best be expected to give an approximate magnitude of the evaporation. References 2 and 3 list several other popular empirical formulae.

In using the empirical equations, the saturated vapour pressure at a given temperature (e_w) is found from a table of e_w *vs* temperature in °C, such as Table 3.3. Often, the wind-velocity data would be available at an elevation other than that needed in the particular equation. However, it is known that in the lower part of the atmosphere, up to a height of about 500 m above the ground level, the wind velocity can be assumed to follow the 1/7 power law as

$$u_h = Ch^{1/7} \tag{3.5}$$

where u_h = wind velocity at a height h above the ground and C = constant. This equation can be used to determine the velocity at any desired level if u_h is known.

Example 3.1

(a) *A reservoir with a surface area of 250 hectares had the following average values of climate parameters during a week: Water temperature = 20°C, Relative humidity = 40%, Wind velocity at 1.0 m above ground surface = 16 km/h. Estimate the average daily evaporation from the lake by using Meyer's formula.*

(b) *An ISI Standard evaporation pan at the site indicated a pan coefficient of 0.80 on the basis of calibration against controlled water budgeting method. If this pan indicated an evaporation of 72 mm in the week under question, (i) estimate the accuracy if Meyer's method relative to the pan evaporation measurements. (ii) Also, estimate the volume of water evaporated from the lake in that week.*

Solution:

(a) From Table 3.3

e_w = 17.54 mm of Hg $\quad$ e_a = 0.4 × 17.54 = 7.02 mm of Hg

u_9 = wind velocity at a height of 9.0 m above ground = $u_1 \times (9)^{1/7}$ = 21.9 km/h

By Meyer's Formula [Eq. (3.3)],

$$E_L = 0.36\,(17.54 - 7.02)\left(1 + \frac{21.9}{16}\right) = 8.97 \text{ mm/day}$$

(b) (i) Daily evaporation as per Pan evaporimeter = $\left(\frac{72.00}{7}\right) \times 0.8 = 8.23$ mm

Error by Meyer's formula = (8.23 – 8.97) = –0.74 mm. Hence, Meyer's method overestimates the evaporation relative to the Pan.

Percentage over estimation by Meyer's formula = (0.74/8.23) × 100 = 9%

(ii) Considering the Pan measurements as the basis, volume of water evaporated from the lake in 7 days = 7 × (8.23/1000) × 250 × 10^4 = 144,025 m^3

[The lake area is assumed to be constant at 250 ha throughout the week.]

3.5 ANALYTICAL METHODS OF EVAPORATION ESTIMATION

The analytical methods for the determination of lake evaporation can be broadly classified into three categories as:

1. Water-budget method,
2. Energy-balance method, and
3. Mass-transfer method.

Water-Budget Method

The water-budget method is the simplest of the three analytical methods and is also the least reliable. It involves writing the hydrological continuity equation for the lake and determining the evaporation from a knowledge or estimation of other variables. Thus considering the daily average values for a lake, the continuity equation is written as

$$P + V_{is} + V_{ig} = V_{os} + V_{og} + E_L + \Delta S + T_L \tag{3.6}$$

where P = daily precipitation

V_{is} = daily surface inflow into the lake
V_{ig} = daily groundwater inflow
V_{os} = daily surface outflow from the lake
V_{og} = daily seepage outflow
E_L = daily lake evaporation
ΔS = increase in lake storage in a day
T_L = daily transpiration loss

All quantities are in units of volume (m^3) or depth (mm) over a reference area. Equation (3.6) can be written as

$$E_L = P + (V_{is} - V_{os}) + (V_{ig} - V_{og}) - T_L - \Delta S \tag{3.7}$$

In this the terms P, V_{is}, V_{os} and ΔS can be measured. However, it is not possible to measure V_{ig}. V_{og} and T_L and therefore these quantities can only be estimated. Transpiration losses can be considered to be insignificant in some reservoirs. If the unit of time is kept large, say weeks or months, better accuracy in the estimate of E_L is possible. In view of the various uncertainties in the estimated values and the possibilities of errors in measured variables, the water-budget method cannot be expected to give very accurate results. However, controlled studies such as at Lake Hefner in USA (1952) have given fairly accurate results by this method.

Energy-Budget Method

The energy-budget method is an application of the law of conservation of energy. The energy available for evaporation is determined by considering the incoming energy, outgoing energy and energy stored in the water body over a known time interval.

Considering the water body as in Fig. 3.4, the energy balance to the evaporating surface in a period of one day is give by

$$H_n = H_a + H_e + H_g + H_s + H_i \tag{3.8}$$

where H_n = net heat energy received by the water surface
$= H_c(1 - r) - H_b$

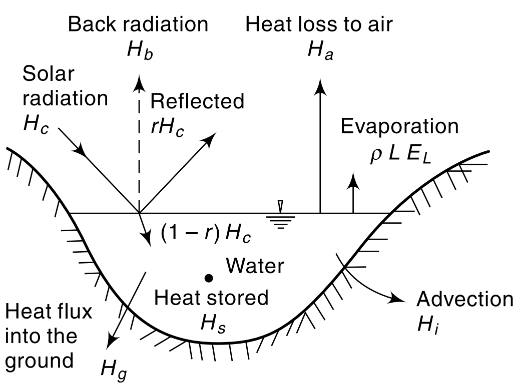


Fig. 3.4 Energy Balance in a Water Body

in which $H_c(1 - r)$ = incoming solar radiation into a surface of reflection coefficient (albedo) r

H_b = back radiation (long wave) from water body
H_a = sensible heat transfer from water surface to air
H_e = heat energy used up in evaporation
= ρLE_L, where ρ = density of water, L = latent heat of evaporation and E_L = evaporation in mm
H_g = heat flux into the ground
H_s = heat stored in water body
H_i = net heat conducted out of the system by water flow (advected energy)

All the energy terms are in calories per square mm per day. If the time periods are short, the terms H_s and H_i can be neglected as negligibly small. All the terms except H_a can either be measured or evaluated indirectly. The sensible heat term H_a which cannot be readily measured is estimated using *Bowen's ratio* β given by the expression

$$\beta = \frac{H_a}{\rho LE_L} = 6.1 \times 10^{-4} \times p_a \frac{T_w - T_a}{e_w - e_a} \tag{3.9}$$

where p_a = atmospheric pressure in mm of mercury, e_w = saturated vapour pressure in mm of mercury, e_a = actual vapour pressure of air in mm of mercury, T_w = temperature of water surface in °C and T_a = temperature of air in °C. From Eqs (3.8) and (3.9) E_L can be evaluated as

$$E_L = \frac{H_n - H_g - H_s - H_i}{\rho L(1+\beta)} \tag{3.10}$$

Estimation of evaporation in a lake by the energy balance method has been found to give satisfactory results, with errors of the order of 5% when applied to periods less than a week. Further details of the energy-budget method are available in Refs 2, 3 and 5.

Mass-Transfer Method

This method is based on theories of turbulent mass transfer in boundary layer to calculate the mass water vapour transfer from the surface to the surrounding atmosphere. However, the details of the method are beyond the scope of this book and can be found in published literature[2, 5]. With the use of quantities measured by sophisticated (and expensive) instrumentation, this method can give satisfactory results.

3.6 RESERVOIR EVAPORATION AND METHODS FOR ITS REDUCTION

Any of the methods mentioned above may be used for the estimation of reservoir evaporation. Although analytical methods provide better results, they involve parameters that are difficult to assess or expensive to obtain. Empirical equations can at best give approximate values of the correct order of magnitude. Therefore, the pan measurements find general acceptance for practical application. Mean monthly and annual evaporation data collected by IMD are very valuable in field estimations. The water volume lost due to evaporation from a reservoir in a month is calculated as

$$V_E = A\, E_{pm}\, C_p \tag{3.11}$$

where V_E = volume of water lost in evaporation in a month (m^3)
A = average reservoir area during the month (m^2)

E_{pm} = pan evaporation loss in metres in a month (m)

= E_L in mm/day × No. of days in the month × 10^{-3}

C_p = relevant pan coefficient

Evaporation from a water surface is a continuous process. Typically under Indian conditions, evaporation loss from a water body is about 160 cm in a year with enhanced values in arid regions. The quantity of stored water lost by evaporation in a year is indeed considerable as the surface area of many natural and man-made lakes in the country are very large. While a small sized tank (lake) may have a surface area of about 20 ha large reservoirs such as Narmada Sagar have surface area of about 90,000 ha. Table 3.2 (a) indicates surface areas and capacities of some large Indian reservoirs.

Table 3.2(a) Surface Areas and Capacities of Some Indian Reservoirs

Sl. No	Reservoir	River	State	Surface area at MRL in km^2	Gross capacity of the reservoir in Mm^3
1.	Narmada Sagar	Narmada	Madhya Pradesh	914	12,230
2.	Nagarjuna Sagar	Krishna	Andhra Pradesh	285	11,315
3.	Sardar Sarovar	Narmada	Gujarat	370	9510
4.	Bhakra	Sutlej	Punjab	169	9868
5.	Hirakud	Mahanadi	Orissa	725	8141
6.	Gandhi Sagar	Chambal	Madhya Pradesh	660	7746
7.	Tungabhadra	Tungabhadra	Karnataka	378	4040
8.	Shivaji Sagar	Koyna	Maharashtra	115	2780
9.	Kadana	Mahi	Gujarat	172	1714
10.	Panchet	Damodar	Jharkhand	153	1497

Using evaporation data from 29 major and medium reservoirs in the country, the National Commission for integrated water resources development (1999)[8] has estimated the national water loss due to evaporation at various time horizons as below:

Table 3.2(b) Water Loss due to Evaporation (Volume in km^3)

Sl. No.	Particular	1997	2010	2025	2050
1.	Live Capacity–Major storage	173.7	211.4	249.2	381.5
2.	Live Capacity–Minor storage	34.7	42.3	49.8	76.3
3.	Evaporation for Major storage Reservoirs @ 15% of live capacity	26.1	31.7	37.4	57.2
4.	Evaporation for Minor storage Reservoirs @ 25% of live capacity	8.7	10.6	12.5	19.1
5.	**Total Evaporation loss**	**35**	**42**	**50**	**76**

Roughly, a quantity equivalent to entire live capacity of minor storages is lost annually by evaporation. As the construction of various reservoirs as a part of water resources developmental effort involve considerable inputs of money, which is a scarce resource, evaporation from such water bodies signifies an economic loss. In semi-arid zones where water is scarce, the importance of conservation of water through reduction of evaporation is obvious.

METHODS TO REDUCE EVAPORATION LOSSES

Various methods available for reduction of evaporation losses can be considered in three categories:

(I) REDUCTION OF SURFACE AREA Since the volume of water lost by evaporation is directly proportional to the surface area of the water body, the reduction of surface area wherever feasible reduces evaporation losses. Measures like having deep reservoirs in place of wider ones and elimination of shallow areas can be considered under this category.

(II) MECHANICAL COVERS Permanent roofs over the reservoir, temporary roofs and floating roofs such as rafts and light-weight floating particles can be adopted wherever feasible. Obviously these measures are limited to very small water bodies such as ponds, etc.

(III) CHEMICAL FILMS This method consists of applying a thin chemical film on the water surface to reduce evaporation. Currently this is the only feasible method available for reduction of evaporation of reservoirs up to moderate size.

Certain chemicals such as *cetyl alcohol* (hexadecanol) and *stearyl alcohol* (octadecanol) form monomolecular layers on a water surface. These layers act as evaporation inhibitors by preventing the water molecules to escape past them. The thin film formed has the following desirable features:

1. The film is strong and flexible and does not break easily due to wave action.
2. If punctured due to the impact of raindrops or by birds, insects, etc., the film closes back soon after.
3. It is pervious to oxygen and carbon dioxide; the water quality is therefore not affected by its presence.
4. It is colourless, odourless and nontoxic.

Cetyl alcohol is found to be the most suitable chemical for use as an evaporation inhibitor. It is a white, waxy, crystalline solid and is available as lumps, flakes or powder. It can be applied to the water surface in the form of powder, emulsion or solution in mineral turpentine. Roughly about 3.5 N/hectare/day of cetyl alcohol is needed for effective action. The chemical is periodically replenished to make up the losses due to oxidation, wind sweep of the layer to the shore and its removal by birds and insects. Evaporation reduction can be achieved to a maximum if a film pressure of 4×10^{-2} N/m is maintained.

Controlled experiments with evaporation pans have indicated an evaporation reduction of about 60% through use of cetyl alcohol. Under field conditions, the reported values of evaporation reduction range from 20 to 50%. It appears that a reduction of 20–30% can be achieved easily in small size lakes (≤1000 hectares) through the use of these monomolecular layers. The adverse effect of heavy wind appears to be the only major impediment affecting the efficiency of these chemical films.

B: EVAPOTRANSPIRATION

3.7 TRANSPIRATION

Transpiration is the process by which water leaves the body of a living plant and reaches the atmosphere as water vapour. The water is taken up by the plant-root system and escapes through the leaves. The important factors affecting transpiration are: atmospheric vapour pressure, temperature, wind, light intensity and characteristics of the plant, such as the root and leaf systems. For a given plant, factors that affect the free-water evaporation also affect transpiration. However, a major difference exists between transpiration and evaporation. Transpiration is essentially confined to daylight hours and the rate of transpiration depends upon the growth periods of the plant. Evaporation, on the other hand, continues all through the day and night although the rates are different.

3.8 EVAPOTRANSPIRATION

While transpiration takes place, the land area in which plants stand also lose moisture by the evaporation of water from soil and water bodies. In hydrology and irrigation practice, it is found that evaporation and transpiration processes can be considered advantageously under one head as evapotranspiration. The term *consumptive use* is also used to denote this loss by evapotranspiration. For a given set of atmospheric conditions, evapotranspiration obviously depends on the availability of water. If sufficient moisture is always available to completely meet the needs of vegetation fully covering the area, the resulting evapotranspiration is called *potential evapotranspiration* (PET). Potential evapotranspiration no longer critically depends on the soil and plant factors but depends essentially on the climatic factors. The real evapotranspiration occurring in a specific situation is called *actual evapotranspiration* (AET).

It is necessary to introduce at this stage two terms: *field capacity* and *permanent wilting point*. Field capacity is the maximum quantity of water that the soil can retain against the force of gravity. Any higher moisture input to a soil at field capacity simply drains away. Permanent wilting point is the moisture content of a soil at which the moisture is no longer available in sufficient quantity to sustain the plants. At this stage, even though the soil contains some moisture, it will be so held by the soil grains that the roots of the plants are not able to extract it in sufficient quantities to sustain the plants and consequently the plants wilt. The field capacity and permanent wilting point depend upon the soil characteristics. The difference between these two moisture contents is called *available water*, the moisture available for plant growth.

If the water supply to the plant is adequate, soil moisture will be at the field capacity and AET will be equal to PET. If the water supply is less than PET, the soil dries out and the ratio AET/PET would then be less than unity. The decrease of the ratio AET/PET with available moisture depends upon the type of soil and rate of drying. Generally, for clayey soils, AET/PET = 1.0 for nearly 50% drop in the available moisture. As can be expected, when the soil moisture reaches the permanent wilting point, the AET reduces to zero (Fig. 3.5). For a catchment in a given period of time, the hydrologic budget can be written as

$$P - R_s - G_o - E_{\text{act}} = \Delta S \tag{3.12}$$

where P = precipitation, R_s = surface runoff, G_o = subsurface outflow, E_{act} = actual evapotranspiration (AET) and ΔS = change in the moisture storage. This water budgeting can be used to calculate E_{act} by knowing or estimating other elements of Eq. (3.12). Generally, the sum of R_s and G_o can be taken as the stream flow at the basin outlet without much error. Method of estimating AET is given in Sec. 3.11.

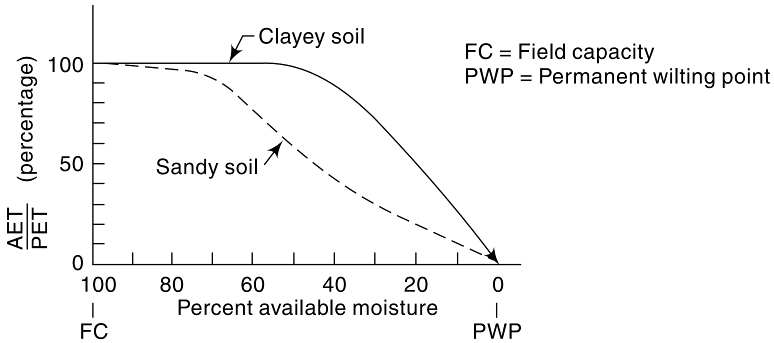


Fig. 3.5 Variation of AET

Except in a few specialized studies, all applied studies in hydrology use PET (not AET) as a basic parameter in various estimations related to water utilizations connected with evapotranspiration process. It is generally agreed that PET is a good approximation for lake evaporation. As such, where pan evaporation data is not available, PET can be used to estimate lake evaporation.

3.9 MEASUREMENT OF EVAPOTRANSPIRATION

The measurement of evapotranspiration for a given vegetation type can be carried out in two ways: either by using lysimeters or by the use of field plots.

Lysimeters

A lysimeter is a special watertight tank containing a block of soil and set in a field of growing plants. The plants grown in the lysimeter are the same as in the surrounding field. Evapotranspiration is estimated in terms of the amount of water required to maintain constant moisture conditions within the tank measured either volumetric ally or gravimetrically through an arrangement made in the lysimeter. Lysimeters should be designed to accurately reproduce the soil conditions, moisture content, type and size of the vegetation of the surrounding area. They should be so buried that the soil is at the same level inside and outside the container. Lysimeter studies are time-consuming and expensive.

Field Plots

In special plots all the elements of the water budget in a known interval of time are measured and the evapotranspiration determined as

$$\text{Evapotranspiration} = [\text{precipitation} + \text{irrigation input} - \text{runoff} \\ - \text{increase in soil storage groundwater loss}]$$

Measurements are usually confined to precipitation, irrigation input, surface runoff and soil moisture. Groundwater loss due to deep percolation is difficult to measure and can be minimised by keeping the moisture condition of the plot at the field capacity. This method provides fairly reliable results.

3.10 EVAPOTRANSPIRATION EQUATIONS

The lack of reliable field data and the difficulties of obtaining reliable evapotranspiration data have given rise to a number of methods to predict PET by using climatological data. Large number of formulae are available: they range from purely empirical ones to those backed by theoretical concepts. Two useful equations are given below.

Penman's Equation

Penman's equation is based on sound theoretical reasoning and is obtained by a combination of the energy-balance and mass-transfer approach. Penman's equation, incorporating some of the modifications suggested by other investigators is

$$\text{PET} = \frac{AH_n + E_a \gamma}{A + \gamma} \tag{3.13}$$

where PET = daily potential evapotranspiration in mm per day

A = slope of the saturation vapour pressure *vs* temperature curve at the mean air temperature, in mm of mercury per °C (Table 3.3)

H_n = net radiation in mm of evaporable water per day

E_a = parameter including wind velocity and saturation deficit

γ = psychrometric constant = 0.49 mm of mercury/°C

The net radiation is the same as used in the energy budget [Eq. (3.8)] and is estimated by the following equation:

$$H_n = H_a(1-r)\left(a + b\frac{n}{N}\right) - \sigma T_a^4 (0.56 - 0.092\sqrt{e_a})\left(0.10 + 0.90\frac{n}{N}\right) \tag{3.14}$$

where H_a = incident solar radiation outside the atmosphere on a horizontal surface, expressed in mm of evaporable water per day (it is a function of the latitude and period of the year as indicated in Table 3.4)

a = a constant depending upon the latitude ϕ and is given by $a = 0.29 \cos \phi$

b = a constant with an average value of 0.52

n = actual duration of bright sunshine in hours

N = maximum possible hours of bright sunshine (it is a function of latitude as indicated in Table 3.5)

r = reflection coefficient (albedo). Usual ranges of values of r are given below.

Surface	**Range of *r* values**
Close ground corps	0.15–0.25
Bare lands	0.05–0.45
Water surface	0.05
Snow	0.45–0.95

σ = Stefan-Boltzman constant = 2.01×10^{-9} mm/day

T_a = mean air temperature in degrees kelvin = 273 + °C

e_a = actual mean vapour pressure in the air in mm of mercury

The parameter E_a is estimated as

$$E_a = 0.35\left(1 + \frac{u_2}{160}\right)(e_w - e_a) \tag{3.15}$$

in which

u_2 = mean wind speed at 2 m above ground in km/day

e_w = saturation vapour pressure at mean air temperature in mm of mercury (Table 3.3)

e_a = actual vapour pressure, defined earlier

For the computation of PET, data on n, e_a, u_2, mean air temperature and nature of surface (i.e. value of r) are needed. These can be obtained from actual observations or through available meteorological data of the region. Equations (3.13), (3.14) and (3.15) together with Tables 3.3, 3.4, and 3.5 enable the daily PET to be calculated. It may be noted that Penman's equation can be used to calculate evaporation from a water surface by using $r = 0.05$. Penman's equation is widely used in India, the UK, Australia and in some parts of USA. Further details about this equation are available elsewhere[2,5,7].

EXAMPLE 3.2 *Calculate the potential evapotranspiration from an area near New Delhi in the month of November by Penman's fomula. The following data are available:*

Latitude : 28°4′N
Elevation : 230 m (above sea level)

Table 3.3 Saturation Vapour Pressure of Water

Temperature (°C)	Saturation vapour pressure e_w (mm of Hg)	A (mm/°C)
0	4.58	0.30
5.0	6.54	0.45
7.5	7.78	0.54
10.0	9.21	0.60
12.5	10.87	0.71
15.0	12.79	0.80
17.5	15.00	0.95
20.0	17.54	1.05
22.5	20.44	1.24
25.0	23.76	1.40
27.5	27.54	1.61
30.0	31.82	1.85
32.5	36.68	2.07
35.0	42.81	2.35
37.5	48.36	2.62
40.0	55.32	2.95
45.0	71.20	3.66

$$e_w = 4.584 \exp\left(\frac{17.27\,t}{237.3 + t}\right) \text{ mm of Hg, where } t = \text{temperature in °C.}$$

Table 3.4 Mean Monthly Solar Radiation at Top of Atmosphere, H_a in mm of Evaporable Water/Day

North latitude	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0°	14.5	15.0	15.2	14.7	13.9	13.4	13.5	14.2	14.9	15.0	14.6	14.3
10°	12.8	13.9	14.8	15.2	15.0	14.8	14.8	15.0	14.9	14.1	13.1	12.4
20°	10.8	12.3	13.9	15.2	15.7	15.8	15.7	15.3	14.4	12.9	11.2	10.3
30°	8.5	10.5	12.7	14.8	16.0	16.5	16.2	15.3	13.5	11.3	9.1	7.9
40°	6.0	8.3	11.0	13.9	15.9	16.7	16.3	14.8	12.2	9.3	6.7	5.4
50°	3.6	5.9	9.1	12.7	15.4	16.7	16.1	13.9	10.5	7.1	4.3	3.0

Table 3.5 Mean Monthly Values of Possible Sunshine Hours, N

North latitude	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0°	12.1	12.1	12.1	12.1	12.1	12.1	12.1	12.1	12.1	12.1	12.1	12.1
10°	11.6	11.8	12.1	12.4	12.6	12.7	12.6	12.4	12.9	11.9	11.7	11.5
20°	11.1	11.5	12.0	12.6	13.1	13.3	13.2	12.8	12.3	11.7	11.2	10.9
30°	10.4	11.1	12.0	12.9	13.7	14.1	13.9	13.2	12.4	11.5	10.6	10.2
40°	9.6	10.7	11.9	13.2	14.4	15.0	14.7	13.8	12.5	11.2	10.0	9.4
50°	8.6	10.1	11.8	13.8	15.4	16.4	16.0	14.5	12.7	10.8	9.1	8.1

Mean monthly temperature	:	*19° C*
Mean relative humidity	:	*75%*
Mean observed sunshine hours	:	*9 h*
Wind velocity at 2 m height	:	*85 km/day*
Nature of surface cover	:	*Close-ground green crop*

SOLUTION: From Table 3.3,

$$A = 1.00 \text{ mm/°C} \qquad e_w = 16.50 \text{ mm of Hg}$$

From Table 3.4

$$H_a = 9.506 \text{ mm of water/day}$$

From Table 3.5

$$N = 10.716 \text{ h} \qquad n/N = 9/10.716 = 0.84$$

From given data

$$e_a = 16.50 \times 0.75 = 12.38 \text{ mm of Hg}$$
$$a = 0.29 \cos 28° \, 4' = 0.2559$$
$$b = 0.52$$
$$\sigma = 2.01 \times 10^{-9} \text{ mm/day}$$
$$T_a = 273 + 19 = 292 \text{ K}$$
$$\sigma T_a^4 = 14.613$$

r = albedo fer close-ground green crop is taken as 0.25

From Eq. (3.14),

$$H_n = 9.506 \times (1 - 0.25) \times (0.2559 + (0.52 \times 0.84)) - 14.613 \times (0.56 - 0.092\sqrt{12.38}) \times (0.10 + (0.9 \times 0.84))$$
$$= 4.936 - 2.946 = 1.990 \text{ mm of water/day}$$

From Eq. (3.15),

$$E_a = 0.35 \times \left(1 + \frac{85}{160}\right) \times (16.50 - 12.38) = 2.208 \text{ mm/day}$$

From Eq. (3.13), noting the value of $\gamma = 0.49$.

$$\text{PET} = \frac{(1 \times 1.990) + (2.208 \times 0.49)}{(1.00 + 0.49)} = 2.06 \text{ mm/day}$$

EXAMPLE 3.3 *Using the data of Example 3.2, estimate the daily evaporation from a lake situated in that place.*

SOLUTION: For estimating the daily evaporation from a lake, Penman's equation is used with the albedo $r = 0.05$.

Hence

$$H_n = 4.936 \times \frac{(1.0 - 0.05)}{(1.0 - 0.25)} - 2.946 = 6.252 - 2.946 = 3.306 \text{ mm of water/day}$$
$$E_a = 2.208 \text{ mm/day}$$

From Eq. (3.13),

$$\text{PET} = \text{Lake evaporation}$$

$$= \frac{(1.0 \times 3.306) + (2.208 \times 0.49)}{(1.0 - 0.49)} = 2.95 \text{ mm/day}$$

Reference Crop Evapotranspiration (ET_O) In irrigation practice, the PET is extensively used in calculation of crop-water requirements. For purposes of standardization, FAO recommends[3] a *reference crop evapotranspiration* or *reference evapotranspiration* denoted as ET_o. The reference surface is a hypothetical grass reference crop with an assumed crop height of 0.12 m, a defined fixed surface resistance of 70 s m^{-1} and an albedo of 0.23. The reference surface closely resembles an extensive surface of green, well-watered grass of uniform height, actively growing and completely shading the ground. The defined fixed surface resistance 70 s m^{-1} implies a moderately dry soil surface resulting from about a weekly irrigation frequency. The FAO recommends a method called *FAO Penman-Monteith method* to estimate ET_o by using radiation, air temperature, air humidity and wind speed data. Details of *FAO Penman-Monteith method* are available in Ref. 3.

The potential evapotranspiration of any other crop (ET) is calculated by multiplying the reference crop evapotranspiration by a coefficient K, the value of which changes with stage of the crop. Thus

$$ET = K(ET_o) \tag{3.16}$$

The value of K varies from 0.5 to 1.3. Table 3.7 gives average values of K for some selected crops.

Empirical Formulae

A large number of empirical formulae are available for estimation of PET based on climatological data. These are not universally applicable to all climatic areas. They should be used with caution in areas different from those for which they were derived.

Blaney-Criddle Formula

This purely empirical formula based on data from arid western United States. This formula assumes that the PET is related to hours of sunshine and temperature, which are taken as measures of solar radiation at an area. The potential evapotranspiration in a crop-growing season is given by

$$E_T = 2.54\, KF$$

and

$$F = \Sigma P_h \bar{T}_f / 100 \tag{3.17}$$

where E_T = PET in a crop season in cm

K = an empirical coefficient, depends on the type of the crop and stage of growth

F = sum of monthly consumptive use factors for the period

P_h = monthly percent of annual day-time hours, depends on the latitude of the place (Table 3.6)

and $\bar{T}_f$ = mean monthly temperature in °F

Values of K depend on the month and locality. Average value for the season for selected crops is given in Table 3.7. The Blaney-Criddle formula is largely used by irrigation engineers to calculate the water requirements of crops, which is taken as the difference between PET and effective precipitation. Blaney-Morin equation is another empirical formula similar to Eq. (3.17) but with an additional correction for humidity.

Table 3.6 Monthly Daytime Hours Percentages, P_h, for use in Blaney-Criddle Formula (Eq. 3.17)

North latitude (deg)	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0	8.50	7.66	8.49	8.21	8.50	8.22	8.50	8.49	8.21	8.50	8.22	8.50
10	8.13	7.47	8.45	8.37	8.81	8.60	8.86	8.71	8.25	8.34	7.91	8.10
15	7.94	7.36	8.43	8.44	8.98	8.80	9.05	8.83	8.28	8.26	7.75	7.88
20	7.74	7.25	8.41	8.52	9.15	9.00	9.25	8.96	8.30	8.18	7.58	7.66
25	7.53	7.14	8.39	8.61	9.33	9.23	9.45	9.09	8.32	8.09	7.40	7.42
30	7.30	7.03	8.38	8.72	9.53	9.49	9.67	9.22	8.33	7.99	7.19	7.15
35	7.05	6.88	8.35	8.83	9.76	9.77	9.93	9.37	8.36	7.87	6.97	6.86
40	6.76	6.72	8.33	8.95	10.02	10.08	10.22	9.54	8.39	7.75	6.72	6.52

Table 3.7 Values of K for Selected Crops

Crop	Average value of K	Range of monthly values
Rice	1.10	0.85–1.30
Wheat	0.65	0.50–0.75
Maize	0.65	0.50–0.80
Sugarcane	0.90	0.75–1.00
Cotton	0.65	0.50–0.90
Potatoes	0.70	0.65–0.75
Natural Vegetation:		
(a) Very dense	1.30	
(b) Dense	1.20	
(c) Medium	1.00	
(d) Light	0.80	

EXAMPLE 3.4 *Estimate the PET of an area for the season November to February in which wheat is grown. The area is in North India at a latitude of 30° N with mean monthly temperatures as below:*

Month	Nov.	Dec.	Jan.	Feb.
Temp. (°C)	16.5	13.0	11.0	14.5

Use the Blaney-Criddle formula.

SOLUTION: From Table 3.7, for wheat $K = 0.65$. Values of P_h for 30° N is read from Table 3.6, the temperatures are converted to Fahrenheit and the calculations are performed in the following table.

Month	$\bar{T}_f$	P_h	$P_h \bar{T}_f / 100$
Nov.	61.7	7.19	4.44
Dec.	55.4	7.15	3.96
Jan.	51.8	7.30	3.78
Feb.	58.1	7.03	4.08
		$\Sigma P_h \bar{T}_f /100 =$	16.26

By Eq. (3.17),

$$E_T = 2.54 \times 16.26 \times 0.65 = 26.85 \text{ cm.}$$

Thornthwaite Formula This formula was developed from data of eastern USA and uses only the mean monthly temperature together with an adjustment for day-lengths. The PET is given by this formula as

$$E_T = 1.6\, L_a \left(\frac{10\bar{T}}{I_t}\right)^a \tag{3.18}$$

where E_T = monthly PET in cm

L_a = adjustment for the number of hours of daylight and days in the month, related to the latitude of the place (Table 3.8)

$\bar{T}$ = mean monthly air temperature °C

I_t = the total of 12 monthly values of heat index = $\sum_{1}^{12} i$, where $i = (\bar{T}/5)^{1.514}$

a = an empirical constant

$= 6.75 \times 10^{-7} I_t^3 - 7.71 \times 10^{-5} I_t^2 + 1.792 \times 10^{-2} I_t + 0.49239$

Table 3.8 Adjustment Factor L_a for Use in Thornthwaite Formula (Eq. 3.18)

North latitude (deg)	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0	1.04	0.94	1.04	1.01	1.04	1.01	1.04	1.04	1.01	1.04	1.01	1.04
10	1.00	0.91	1.03	1.03	1.08	1.06	1.08	1.07	1.02	1.02	0.98	0.99
15	0.97	0.91	1.03	1.04	1.11	1.08	1.12	1.08	1.02	1.01	0.95	0.97
20	0.95	0.90	1.03	1.05	1.13	1.11	1.14	1.11	1.02	1.00	0.93	0.94
25	0.93	0.89	1.03	1.06	1.15	1.14	1.17	1.12	1.02	0.99	0.91	0.91
30	0.90	0.87	1.03	1.08	1.18	1.17	1.20	1.14	1.03	0.98	0.89	0.88
40	0.84	0.83	1.03	1.11	1.24	1.25	1.27	1.18	1.04	0.96	0.83	0.81

3.11 POTENTIAL EVAPOTRANSPIRATION OVER INDIA

Using Penman's equation and the available climatological data, PET estimate[5] for the country has been made. The mean annual PET (in cm) over various parts of the country is shown in the form of *isopleths*—the lines on a map through places having equal depths of evapotranspiration [Fig. 3.6(a)]. It is seen that the annual PET ranges from 140 to 180 cm over most parts of the country. The annual PET is highest at Rajkot, Gujarat with a value of 214.5 cm. Extreme south-east of Tamil Nadu also show high average values greater than 180 cm. The highest PET for southern peninsula is at Tiruchirapalli, Tamil Nadu with a value of 209 cm. The variation of monthly PET at some stations located in different climatic zones in the country is shown in Fig. 3.6(b). Valuable PET data relevant to various parts of the country are available in Refs 4 and 7.

3.12 ACTUAL EVAPOTRANSPIRATION (AET)

AET for hydrological and irrigation applications can be obtained through a process water budgeting and accounting for soil-plant-atmosphere interactions. A simple procedure due to *Doorenbos* and *Pruit* is as follows:

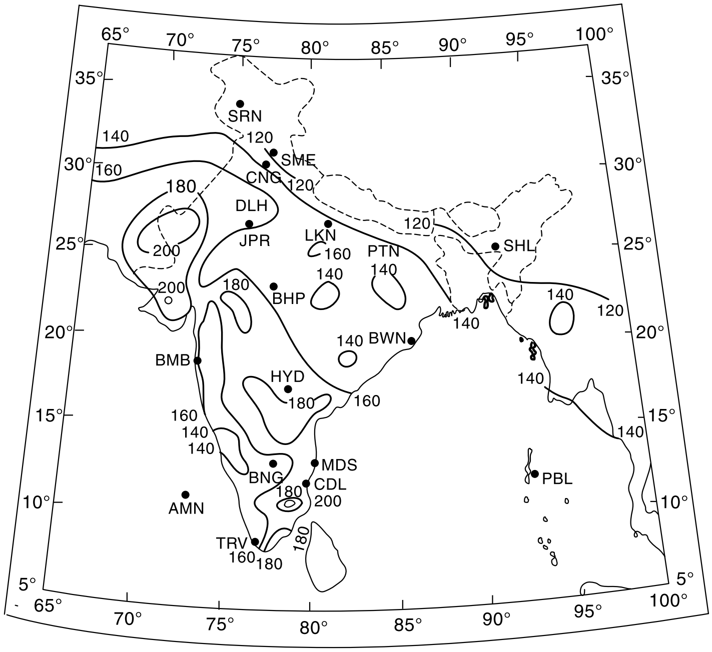


Fig. 3.6(a) Annual PET (cm) over India
(*Source: Scientific: Report No.* 136, IMD, 1971, © Government of India Copyright)

Based upon survey of India map with the permission of the Surveyor General of India, © Government of India Copyright 1984

The territorial waters of India extend into the sea to a distance of 200 nautical miles measured from the appropriate baseline

Responsibility for the correctness of internal details on the map rests with the publisher.

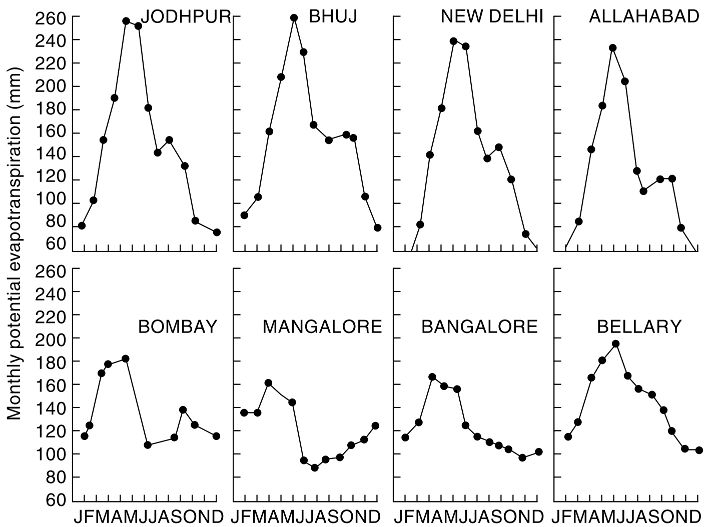


Fig. 3.6(b) Monthly Variation of PET (mm)
(*Source: Scientific Report No.* 136, India Meteorological Department, 1971, © Government of India Copyright)

1. Using available meteorological data the reference crop evapotranspiration (ET_o) is calculated.
2. The crop coefficient K for the given crop (and stage of growth) is obtained from published tables such as Table 3.7. The potential crop evapotrans-piration ET_c is calculated using Eq. 3.16 as $ET_c = K(ET_o)$.
3. The actual evapotranspiration (ET_a) at any time t at the farm having the given crop is calculated as below:
   - If $AASW \geq (1 - p)\, MASW$

     $$ET_a = ET_o \qquad \text{(known as } \textit{potential condition}\text{)} \tag{3.19-a}$$
   - If $AASW < (1 - p)\, MASW$

     $$ET_a = \left[\frac{AASW}{(1-p)\,MASW}\right] ET_c \tag{3.19-b}$$

where $MASW$ = total available soil water over the root depth
$AASW$ = actual available soil-water at time t over the root depth
p = soil-water depletion factor for a given crop and soil complex. (Values of p ranges from about 0.1 for sandy soils to about 0.5 for clayey soils)

[Note the equivalence of terms used earlier as $PET = ET_o$ and $AET = ET_a$]

EXAMPLE 3.5 *A recently irrigated field plot has on Day 1 the total available soil moisture at its maximum value of 10 cm. If the reference crop evapotranspiration is 5.0 mm/day, calculate the actual evapotranspiration on Day 1, Day 6 and Day 7. Assume soil-water depletion factor p = 0.20 and crop factor K = 0.8.*

SOLUTION: Here $ET_o = 5.0$ mm and $MASW = 100$ mm

$(1 - p)\, MASW = (1 - 0.2) \times 100 = 80.0$ and $ET_c = 0.9 \times 5.0 = 4.5$ mm/day

Day 1: $AASW = 100 \text{ mm} > (1 - p)\, MASW$

Hence potential condition exists and $ET_a = ET_c = 4.5$ mm/day

This rate will continue till a depletion of (100 – 80) = 20 mm takes place in the soil. This will take 20/4.5 = 4.44 days. Thus **Day 5** also will have $ET_a = ET_c = 4.5$ mm/day

Day 6: At the beginning of Day 6, $AASW = (100 - 4.5 \times 5) = 77.5$ mm

Since $AASW < (1 - p)\, MASW$,

$$ET_a = \left[\frac{77.5}{80.0}\right] \times 4.5 = 4.36 \text{ mm}$$

Day 7: At the beginning of Day 7, $AASW = (77.5 - 4.36) = 73.14$ mm

Since $AASW < (1 - p)\, MASW$

$$ET_a = \left[\frac{73.14}{80.0}\right] \times 4.5 = 4.11 \text{ mm.}$$

$AASW$ at the end of Day 7 = 73.14 – 4.11 = 69.03 mm.

C: INITIAL LOSS

In the precipitation reaching the surface of a catchment the major abstraction is from the infiltration process. However, two other processes, though small in magnitude, operate to reduce the water volume available for runoff and thus act as abstractions. These are (i) the *interception process*, and (ii) the *depression storage* and together they are called the *initial loss*. This abstraction represents the quantity of storage that must be satisfied before overland runoff begins. The following two sections deal with these two processes briefly.

3.13 INTERCEPTION

When it rains over a catchment, not all the precipitation falls directly onto the ground. Before it reaches the ground, a part of it may be caught by the vegetation and subsequently evaporated. The volume of water so caught is called *interception.* The intercepted precipitation may follow one of the three possible routes:

1. It may be retained by the vegetation as surface storage and returned to the atmosphere by evaporation; a process termed *interception loss;*
2. It can drip off the plant leaves to join the ground surface or the surface flow; this is known as *throughfall*; and
3. The rainwater may run along the leaves and branches and down the stem to reach the ground surface. This part is called *stemflow.*

Interception loss is solely due to evaporation and does not include transpiration, throughfall or stemflow.

The amount of water intercepted in a given area is extremely difficult to measure. It depends on the species composition of vegetation, its density and also on the storm characteristics. It is estimated that of the total rainfall in an area during a plant-growing season the interception loss is about 10 to 20%. Interception is satisfied during the first part of a storm and if an area experiences a large number of small storms, the annual interception loss due to forests in such cases will be high, amounting to greater than 25% of the annual precipitation. Quantitatively, the variation of interception loss with the rainfall magnitude per storm for small storms is as shown in Fig. 3.7. It is seen that the interception loss is large for a small rainfall and levels off to a constant value for larger storms. For a given storm, the interception loss is estimated as

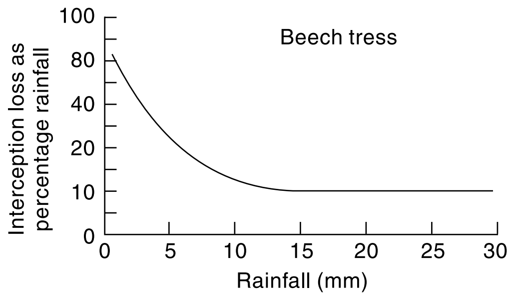


Fig. 3.7 Typical Interception Loss Curve

$$I_i = S_i + K_i Et \tag{3.18}$$

where I_i = interception loss in mm, S_i = interception storage whose value varies from 0.25 to 1.25 mm depending on the nature of vegetation, K_i = ratio of vegetal surface area to its projected area, E = evaporation rate in mm/h during the precipitation and t = duration of rainfall in hours.

It is found that coniferous trees have more interception loss than deciduous ones. Also, dense grasses have nearly same interception losses as full-grown trees and can account for nearly 20% of the total rainfall in the season. Agricultural crops in their growing season also contribute high interception losses. In view of these the interception process has a very significant impact on the ecology of the area related to silvicultural aspects, in *in situ* water harvesting and in the water balance of a region. However, in hydrological studies dealing with floods interception loss is rarely significant and is not separately considered. The common practice is to allow a lump sum value as the initial loss to be deducted from the initial period of the storm.

3.14 DEPRESSION STORAGE

When the precipitation of a storm reaches the ground, it must first fill up all depressions before it can flow over the surface. The volume of water trapped in these depressions is called *depression storage*. This amount is eventually lost to runoff through processes of infiltration and evaporation and thus form a part of the initial loss. Depression storage depends on a vast number of factors the chief of which are: (i) the type of soil, (ii) the condition of the surface reflecting the amount and nature of depression, (iii) the slope of the catchment, and (iv) the antecedent precipitation, as a measure of the soil moisture. Obviously, general expressions for quantitative estimation of this loss are not available. Qualitatively, it has been found that antecedent precipitation has a very pronounced effect on decreasing the loss to runoff in a storm due to depression. Values of 0.50 cm in sand, 0.4 cm in loam and 0.25 cm in clay can be taken as representatives for depression-storage loss during intensive storms.

D: Infiltration

3.15 INFILTRATION

Infiltration is the flow of water into the ground through the soil surface. The distribution of soil moisture within the soil profile during the infiltration process is illustrated in Fig. 3.8. When water is applied at the surface of a soil, four moisture zones in the soil, as indicated in Fig. 3.8 can be identified.

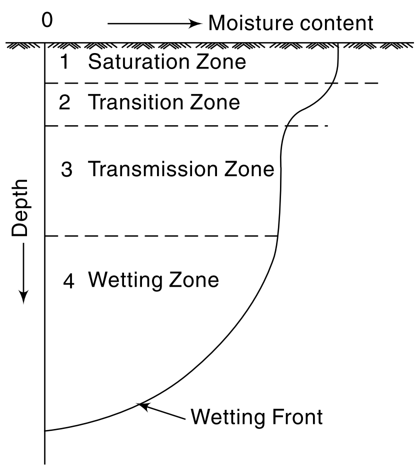


Fig. 3.8 Distribution of Soil Moisture in the Infiltration Process

Zone 1: At the top, a thin layer of *saturated zone* is created.

Zone 2: Beneath zone 1, there is a *transition zone*.

Zone 3: Next lower zone is the *trans-mission zone* where the downward motion of the moisture takes place. The moisture content in this zone is above field capacity but below saturation. Further, it is characterized by unsaturated flow and fairly uniform moisture content.

Zone 4: The last zone is the *wetting zone*. The soil moisture in this zone will be at or near field capacity and the moisture content decreases with the depth. The boundary of the wetting zone is the wetting front where a sharp discontinuity exists between the newly wet soil and original moisture content of the soil. Depending upon the amount of infiltration and physical properties of the soil, the wetting front can extend from a few centimetres to metres.

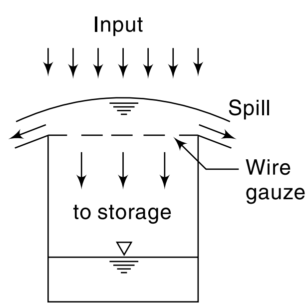


Fig. 3.9 An Analogy for Infiltration

The infiltration process can be easily understood through a simple analogy. Consider a small container covered with wire gauze as in Fig. 3.9. If water is poured into the container a part of it will go into the container and a part overflows. Further, the container can hold only a fixed quantity and when it is full no more flow into the container can take place. While this analogy is highly simplified, it underscores two important aspects; viz. (i) the maximum rate at which the ground can absorb water, the *infiltration capacity* and (ii) the volume of water that the ground can hold, the *field capacity*. Since the infiltered water may contribute to the ground water discharge in addition to increasing the soil moisture, the process can be schematically modelled as in Fig. 3.10(a) and (b) wherein two situations, viz. low intensity rainfall and high intensity rainfall are considered.

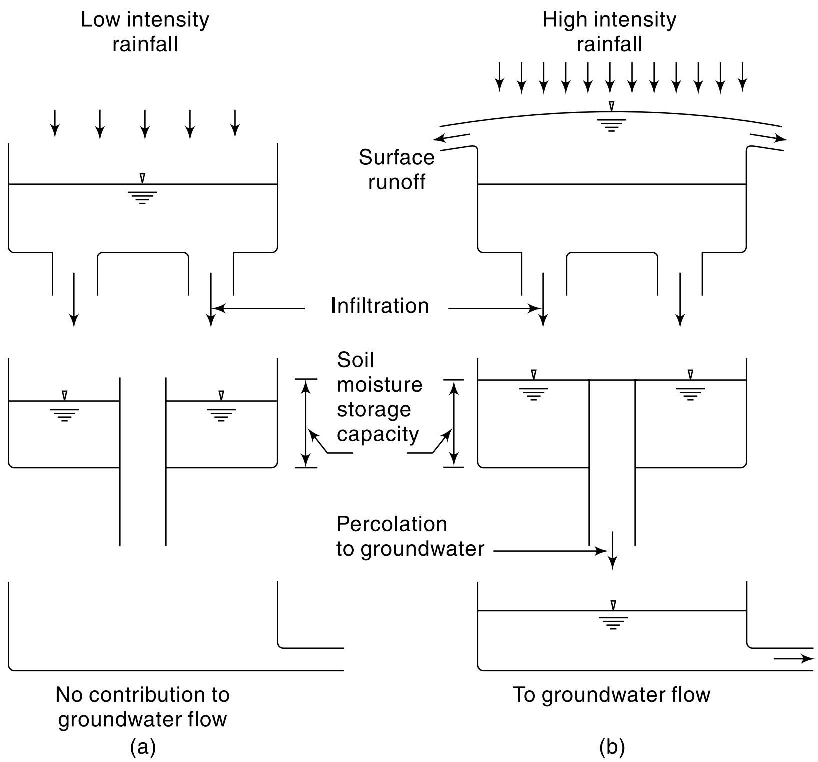


Fig. 3.10 An Infiltration Model

3.16 INFILTRATION CAPACITY

The maximum rate at which a given soil at a given time can absorb water is defined as the *infiltration capacity*. It is designated as f_p and is expressed in units of cm/h. The actual rate of infiltration f can be expressed as

$$f = f_p \text{ when } i \geq f_p$$

and $$f = i \text{ when } i < f_p \tag{3.20}$$

where i = intensity of rainfall. The infiltration capacity of a soil is high at the beginning of a storm and has an exponential decay as the time elapses.

The typical variation of the infiltration capacity f_p of a soil with time is shown in Fig. 3.11. The infiltration capacity of an area is dependent on a large number of factors, chief of them are:

- Characteristics of the soil (Texture, porosity and hydraulic conductivity)
- Condition of the soil surface
- Current moisture content
- Vegetative cover and
- Soil temperature

A few important factors affecting f_p are described below:

Characteristics of Soil The type of soil, viz. sand, silt or clay, its texture, structure, permeability and underdrainage are the important characteristics under this category. A loose, permeable, sandy soil will have a larger infiltration capacity than a tight, clayey soil. A soil with good underdrainage, i.e. the facility to transmit the infiltered water downward to a groundwater storage would obviously have a higher infiltration capacity. When the soils occur in layers, the transmission capacity of the layers determines the overall infiltration rate. Also, a dry soil can absorb more water than one whose pores are already full (Fig. 3.11). The land use has a significant influence on f_p. For example, a forest soil rich in organic matter will have a much higher value of f_p under identical conditions than the same soil in an urban area where it is subjected to compaction.

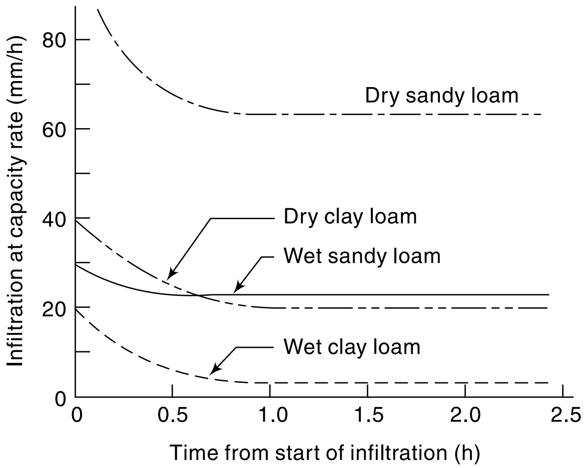


Fig. 3.11 Variation of Infiltration Capacity

Surface of Entry At the soil surface, the impact of raindrops causes the fines in the soil to be displaced and these in turn can clog the pore spaces in the upper layers of the soil. This is an important factor affecting the infiltration capacity. Thus a surface covered with grass and other vegetation which can reduce this process has a pronounced influence on the value of f_p.

Fluid Characteristics Water infiltrating into the soil will have many impurities, both in solution and in suspension. The turbidity of the water, especially the clay and colloid content is an important factor and such suspended particles block the fine pores in the soil and reduce its infiltration capacity. The temperature of the water is a factor in the sense that it affects the viscosity of the water by which in turn affects the infiltration rate. Contamination of the water by dissolved salts can affect the soil structure and in turn affect the infiltration rate.

3.17 MEASUREMENT OF INFILTRATION

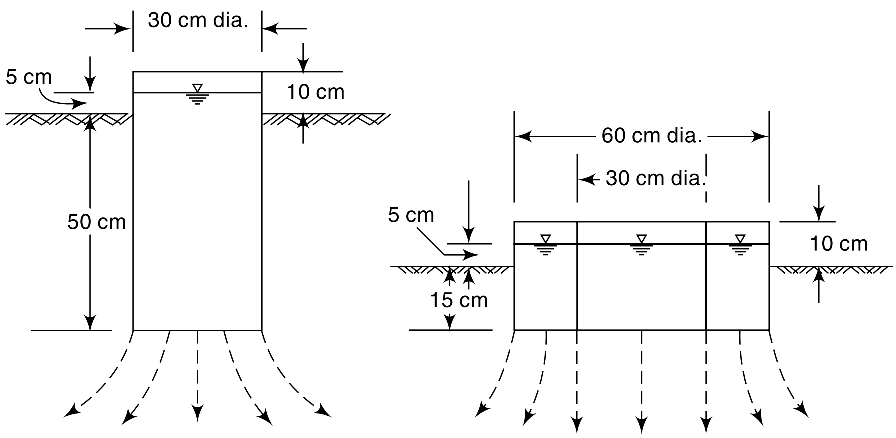
Infiltration characteristics of a soil at a given location can be estimated by

- Using flooding type infiltrometers
- Measurement of subsidence of free water in a large basin or pond
- Rainfall simulator
- Hydrograph analysis

Flooding-Type Infiltrometer

Flooding-type infiltrometers are experimental devices used to obtain data relating to variation of infiltration capacity with time. Two types of flooding type infiltrometers are in common use. They are (a) Tube-type (or Simple) infiltrometer and (b) Double-ring infiltrometer.

Simple (Tube Type) Infiltrometer This is a simple instrument consisting essentially of a metal cylinder, 30 cm diameter and 60 cm long, open at both ends. The cylinder is driven into the ground to a depth of 50 cm (Fig. 3.12(a)). Water is poured into the top part to a depth of 5 cm and a pointer is set to mark the water level. As infiltration proceeds, the volume is made up by adding water from a burette to keep the water level at the tip of the pointer. Knowing the volume of water added during different time intervals, the plot of the infiltration capacity *vs* time is obtained. The experiments are continued till a uniform rate of infiltration is obtained and this may take 2–3 hours. The surface of the soil is usually protected by a perforated disc to prevent formation of turbidity and its settling on the soil surface.



(a) Simple (tube-type) infiltrometer (b) Double-ring infiltrometer

Fig. 3.12 Flooding Type Infiltrometers

A major objection to the simple infiltrometer as above is that the infiltered water spreads at the outlet from the tube (as shown by dotted lines in Fig. 3.12(a)) and as such the tube area is not representative of the area in which infiltration is taking place.

Double-Ring Infiltrometer This most commonly used infiltrometer is designed to overcome the basic objection of the tube infiltrometer, viz. the tube area is not representative of the infiltrating area. In this, two sets of concentrating rings with diameters of 30 cm and 60 cm and of a minimum length of 25 cm, as shown in Fig. 3.12(b), are used. The two rings are inserted into the ground and water is applied into both the rings to maintain a constant depth of about 5.0 cm. The outer ring provides water jacket to the infiltering water from the inner ring and hence prevents the spreading out of the infiltering water of the inner ring. The water depths in the inner and outer rings are kept the same during the observation period. The measurement of the water volume is done on the inner ring only. The experiment is carried out till a constant infiltration rate is obtained. A perforated disc to prevent formation of turbidity and settling of fines on the soil surface is provided on the surface of the soil in the inner ring as well as in the annular space.

As the flooding-type infiltrometer measures the infiltration characteristics at a spot only, a large number of pre-planned experiments are necessary to obtain representative infiltration characteristics for an entire watershed. Some of the chief disadvantages of flooding-type infiltrometers are:

1. the raindrop impact effect is not simulated;
2. the driving of the tube or rings disturbs the soil structure; and
3. the results of the infiltrometers depend to some extent on their size with the larger meters giving less rates than the smaller ones; this is due to the border effect.

Rainfall Simulator

In this a small plot of land, of about 2 m × 4 m size, is provided with a series of nozzles on the longer side with arrangements to collect and measure the surface runoff rate. The specially designed nozzles produce raindrops falling from a height of 2 m and are capable of producing various intensities of rainfall. Experiments are conducted under controlled conditions with various combinations of intensities and durations and the surface runoff rates and volumes are measured in each case. Using the water budget equation involving the volume of rainfall, infiltration and runoff, the infiltration rate and its variation with time are estimated. If the rainfall intensity is higher than the infiltration rate, infiltration capacity values are obtained.

Rainfall simulator type infiltrometers give lower values than flooding type infiltrometers. This is due to effect of the rainfall impact and turbidity of the surface water present in the former.

Hydrograph Analysis

Reasonable estimation of the infiltration capacity of a small watershed can be obtained by analyzing measured runoff hydrographs and corresponding rainfall records. If sufficiently good rainfall records and runoff hydrographs corresponding to isolated storms in a small watershed with fairly homogeneous soils are available, water budget equation can be applied to estimate the abstraction by infiltration. In this the evapotranspiration losses are estimated by knowing the land cover/use of the watershed.

3.18 Modeling Infiltration Capacity

Figure 3.13 shows a typical variation of infiltration capacity f_p with time.

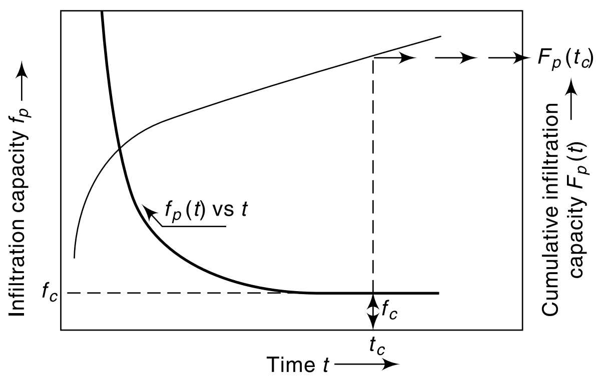


Fig. 3.13 Curves of Infiltration Capacity and Cumulative Infiltration Capacity

Cumulative infiltration capacity $F_p(t)$ is defined as the accumulation of infiltration volume over a time period since the start of the process and is given by

$$F_p = \int_0^t f_p(t)\, dt \tag{3.21-a}$$

Thus the curve $F_p(t)$ vs time in Fig. (3.13) is the mass curve of infiltration. It may be noted that from Eq. (3.21-a) it follow that

$$f_p(t) = \frac{dF_p(t)}{dt} \tag{3.21-b}$$

Many equations have been proposed to express the curves $f_p(t)$ or $F_p(t)$ for use in hydrological analysis. In this section four such equations will be described.

[***Note:*** Practically all the infiltration equations relate infiltration capacity $f_p(t)$ or cumulative infiltration capacity $F_p(t)$ with time and other parameters. As such many authors find it convenient to drop the suffix *p* while denoting capacity. Thus f_p is denoted as *f* and F_p as *F*.]

HORTON'S EQUATION (1933) Horton expressed the decay of infiltration capacity with time as an exponential decay given by

$$f_p = f_c + (f_0 - f_c)\, e^{-K_h t} \qquad \text{for } 0 \geq t \leq t_c \tag{3.22}$$

where f_p = infiltration capacity at any time *t* from the start of the rainfall
f_0 = initial infiltration capacity at $t = 0$
f_c = final steady state infiltration capacity occurring at $t = t_c$. Also, f_c is sometimes known as *constant rate* or *ultimate infiltration capacity.*
K_h = Horton's decay coefficient which depends upon soil characteristics and vegetation cover.

The difficulty of determining the variation of the three parameters f_0, f_c and k_h with soil characteristics and antecedent moisture conditions preclude the general use of Eq. (3.22).

PHILIP'S EQUATION (1957) Philip's two term model relates $F_p(t)$ as

$$F_p = st^{1/2} + Kt \tag{3.23}$$

where s = a function of soil suction potential and called as *sorptivity*
K = Darcy's hydraulic conductivity

By Eq. (3.21-b) infiltration capacity could be expressed as

$$f_p = \frac{1}{2} st^{-1/2} + K \tag{3.24}$$

KOSTIAKOV EQUATION (1932) Kostiakov model expresses cumulative infiltration capacity as

$$F_p = at^b \tag{3.25}$$

where *a* and *b* are local parameters with $a > 0$ and $0 < b < 1$.

The infiltration capacity would now b e expressed by Eq. (3.21-b) as

$$f_p = (ab)\, t^{(b-1)} \tag{3.26}$$

GREEN–AMPT EQUATION (1911) Green and Ampt proposed a model for infiltration capacity based on Darcy's law as

$$f_p = K\left(1 + \frac{\eta S_c}{F_p}\right) \tag{3.27}$$

where η = porosity of the soil
S_c = capillary suction at the wetting front and
K = Darcy's hydraulic conductivity.
Eq. (3.27) could be considered as

$$f_p = m + \frac{n}{F_p} \tag{3.28}$$

where m and n are Green–Ampt parameters of infiltration model.

Estimation of Parameters of Infiltration Models

Data from infiltrometer experiments can be processed to generate data sets f_p and F_p values for various time t values. The following procedures are convenient to evaluate the parameters of the infiltration models.

Horton's Model Value of f_c in a test is obtained from inspection of the data. Equation (3.22) is rearranged to read as

$$(f_p - f_c) = (f_0 - f_c)\, e^{-K_h t} \tag{3.22-a}$$

Taking logarithms $\ln(f_p - f_c) = \ln(f_o - f_c) - K_h t$
Plot $\ln(f_p - f_c)$ against t and fit the best straight line through the plotted points. The intercept gives $\ln(f_0 - f_c)$ and the slope of the straight line is K_h.

Philip's Model Use the expression for f_p as

$$f_p = \frac{1}{2} s t^{-1/2} + K \tag{3.24}$$

Plot the observed values of f_p against $t^{-0.5}$ on an arithmetic graph paper. The best fitting straight line through the plotted points gives K as the intercept and $(s/2)$ as the slope of the line. While fitting Philip's model it is necessary to note that K is positive and to achieve this it may be necessary to neglect a few data points at the initial stages (viz. at small values of t) of the infiltration experiment. K will be of the order of magnitude of the asymptotic value of f_p.

Kostiakov Model Kostiakov model relates F_p to t as

$$F_p = a t^b \tag{3.25}$$

Taking logarithms of both sides of Eq. (3.25),

$$\ln(F_p) = \ln a + b \ln(t)$$

The data is plotted as $\ln(F_p)$ *vs* $\ln(t)$ on an arithmetic graph paper and the best fit straight through the plotted points gives $\ln a$ as intercept and the slope is b. Note that b is a positive quantity such that $0 < b < 1$.

Green–Ampt Model Green–Ampt equation is considered in the form $f_p = m + \frac{n}{F_p}$. Values of f_p are plotted against $(1/F_p)$ on a simple arithmetic graph paper and the best fit straight line is drawn through the plotted points. The intercept and the slope of the line are the coefficients m and n respectively. Sometimes values of f_p and corresponding F_p at very low values of t may have to be omitted to get best fitting straight line with reasonably good correlation coefficient.

[***Note:*** 1. Procedure for calculation of the best fit straight line relating the dependent variable *Y* and independent variable *X* by the least-square error method is described in Section 4.9, Chapter 4.
2. Use of spread sheets (for eg., MS Excel) greatly simplifies these procedures and the best values of parameters can be obtained by fitting regression equations. Further, various plots and the coefficient of correlation, etc. can be calculated with ease.]

EXAMPLE 3.6 *Infiltration capacity data obtained in a flooding type infiltration test is given below:*

Time since start (minutes)	5	10	15	25	45	60	75	90	110	130
Cumulative infiltration depth (cm)	1.75	3.00	3.95	5.50	7.25	8.30	9.30	10.20	11.28	12.36

(a) For this data plot the curves of (i) infiltration capacity vs time, (ii) infiltration capacity vs cumulative infiltration, and (iii) cumulative infiltration vs time.
(b) Obtain the best values of the parameters in Horton's infiltration capacity equation to represent this data set.

SOLUTION: Incremental infiltration values and corresponding infiltration intensities f_p at various data observation times are calculated as shown in the following Table. Also other data required for various plots are calculated as shown in Table 3.9.

Table 3.9 Calculations for Example 3.6

Time in Minutes	Cum. Depth (cm)	Incremental Depth in the interval (cm)	Infiltration rate, f_p(cm/h)	Ln($f_p - f_c$)	Time in hours
0					
5	1.75	1.75	21.00	2.877	0.083
10	3.00	1.25	15.00	2.465	0.167
15	3.95	0.95	11.40	2.099	0.250
25	5.50	1.55	9.30	1.802	0.417
45.	7.25	1.75	5.25	0.698	0.750
60	8.30	1.05	4.20	–0.041	1.000
75	9.30	1.00	4.00	–0.274	1.250
90	10.20	0.90	3.60	–1.022	1.500
110	11.28	1.08	3.24		1.833
130	12.36	1.08	3.24		2.167

(a) Plots of f_p *vs* time and F_p *vs* time are shown in Fig. 3.14. Best fitting curve for plotted points are also shown in the Fig. 3.14-a.
Plot of f_p vs F_p is shown in Fig. 3.14-b.
(b) By observation from Table 3.9, f_c = 3.24 cm/h
Ln($f_p - f_c$) is plotted against time *t* as shown in Fig. 3.14-c. The best fit line through the plotted points is drawn and its equation is obtained as

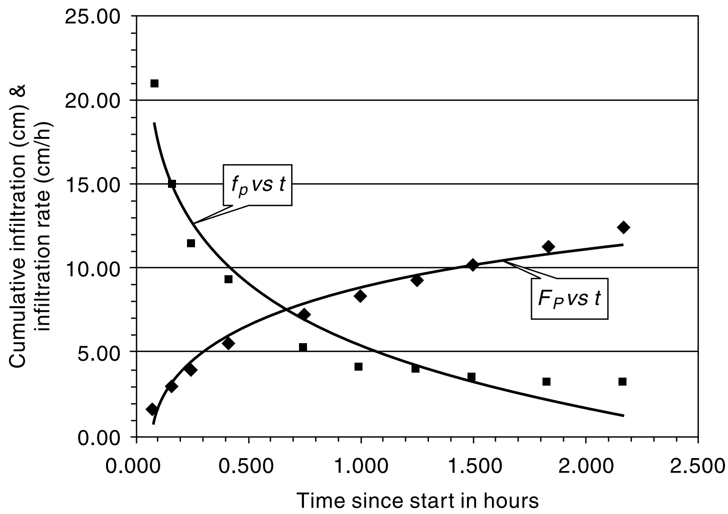


Fig. 3.14 (a) Plot of F_p *vs* Time and f_p *vs* Time

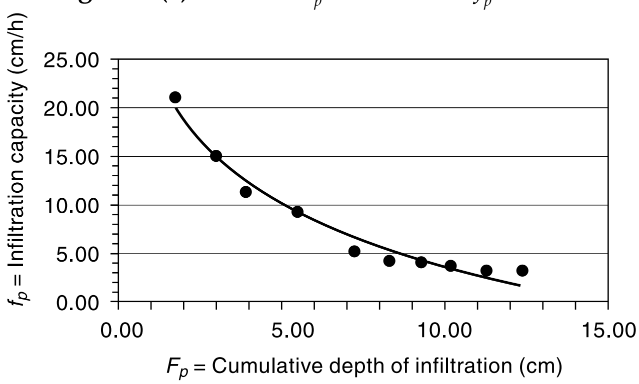


Fig. 3.14 (b) Plot of f_p *vs* F_p

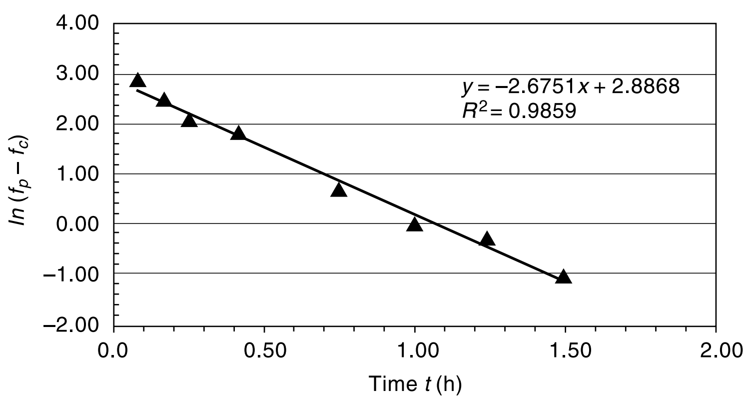


Fig. 3.14 (c) Horton's Equation. Plot of $\ln(f_p - f_c)$ *vs* Time

$\ln(f_p - f_c) = 2.8868 - 2.6751\ t$

$-K_h$ = slope of the best fit line = –2.6751, thus $K_h = 2.6751\ h^{-1}$

$\ln(f_0 - f_c)$ = intercept = 2.8868, thus $f_0 - f_c = 17.94$ and $f_0 = 21.18$ cm/h

EXAMPLE 3.7 *Values of infiltration capacities at various times obtained from an infiltration test are given below. Determine the parameters of (i) Green–Ampt equation, (ii) Philip's equation, and (iii) Kostiakov's equation that best fits this data.*

Time since start (minutes)	5	10	15	20	25	30	60	90	120	150
Cumulative infiltration depth (cm)	1.0	1.8	2.5	3.1	3.6	4.0	6.1	8.1	9.9	11.6

SOLUTION: Incremental infiltration depth values and corresponding infiltration intensities f_p at various data observation times are calculated as shown in Table 3.10. Also, various parameters needed for plotting different infiltration models are calculated as shown in Table 3.10. The units used are f_p in cm/h, F_p in cm and t in hours.

Table 3.10 Calculations Relating to Example 3.7

1	2	3	4	5	6	7	8	9
Time (min)	F_p (cm)	Incremental depth of infiltration (cm)	f_p (cm/h)	t in hours	$t^{-0.5}$	$1/F_p$	Ln F_p	Ln t
5	1.0	1.0	12.0	0.083	3.464	1.000	0.000	–2.485
10	1.8	0.8	9.6	0.167	2.449	0.556	0.588	–1.792
15	2.5	0.7	8.4	0.250	2.000	0.400	0.916	–1.386
20	3.1	0.6	7.2	0.333	1.732	0.323	1.131	–1.099
25	3.6	0.5	6.0	0.417	1.549	0.278	1.281	–0.875
30	4.0	0.4	4.8	0.500	1.414	0.250	1.386	–0.693
60	6.1	2.1	4.2	1.000	1.000	0.164	1.808	0.000
90	8.1	2.0	4.0	1.500	0.816	0.123	2.092	0.405
120	9.9	1.8	3.6	2.000	0.707	0.101	2.293	0.693
150	11.6	1.7	3.4	2.500	0.632	0.086	2.451	0.916

Green–Ampt Equation:

$$f_p = m + \frac{n}{F_p} \qquad (3.28)$$

Values of f_p (col. 4) are plotted against $1/F_p$ (col. 7) on an arithmetic graph paper (Fig. 3.15-a). The best fit straight line through the plotted points is obtained as

$$f_p = 10.042\left(\frac{1}{F_p}\right) + 3.0256.$$

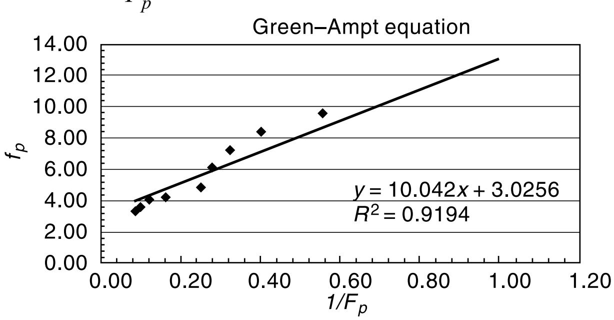


Fig. 3.15 (a) Fitting of Green–Ampt Equation

The coefficients of the Green–Ampt equations are m = 3.0256 and n = 10.042

Philip's Equation: The expression $f_p(t) = \frac{1}{2} s t^{-1/2} + K$ (Eq. 3.24) is used. Values of f_p (Col. 4) are plotted against $t^{-0.5}$ (col. 6) on an arithmetic graph paper (Fig. 15-b). The best fit straight line through the plotted points is obtained as

$f_p = 3.2287\, t^{-0.5} + 1.23$

The coefficients of Philip's equation are $s = 2 \times 3.2287 = 6.4574$ and $K = 1.23$

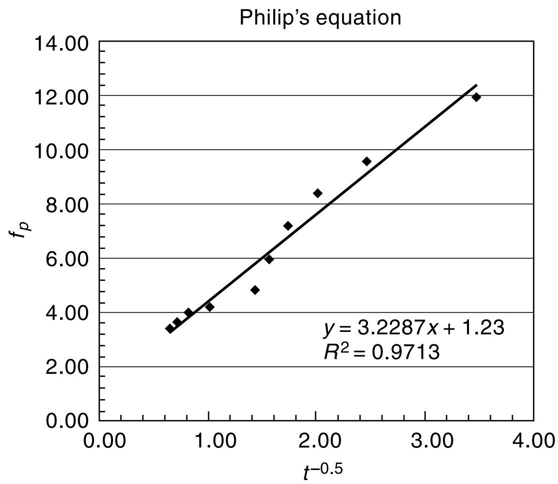


Fig. 3.15 (b) Fitting of Philip's Equation

Kostiakov's Equation:

$F_p(t) = at^b$ Eq. (3.25)

Taking logarithms of both sides of the equation (3.25)

$\ln(F_p) = \ln a + b \ln(t)$.

The data set is plotted as $\ln(F_p)$ *vs* $\ln(t)$ on an arithmetic graph paper (Fig. 3.15-c) and the best fit straight line through the plotted points is obtained as

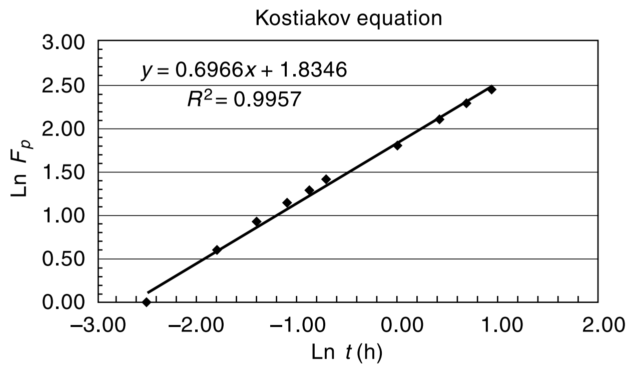


Fig. 3.15 (c) Fitting of Kostiakov Equation

$$\ln(F_p) = 1.8346 + 0.6966 \text{ in}(t).$$

The coefficients of Kostiakov equation are b = 0.6966 and $\ln a$ = 1.8346 and hence a = 6.2626. Best fitting Kostiakov equation for the data is

$$F_p = 6.2626\, t^{0.6966}$$

Example 3.8 *The infiltration capacity in a basin is represented by Horton's equation as*

$$f_p = 3.0 + e^{-2t}$$

where f_p is in cm/h and t is in hours. Assuming the infiltration to take place at capacity rates in a storm of 60 minutes duration, estimate the depth of infiltration in (i) the first 30 minutes and (ii) the second 30 minutes of the storm.

Solution:

$$F_p = \int_0^t f_p\, dt \quad \text{and} \quad f_p = 3.0 + e^{-2t}$$

(i) In the first 0.5 hour

$$F_{p1} = \int_0^{0.5} (3.0 + e^{-2t})\, dt = \left[3.0t - \frac{1}{2}e^{-2t}\right]_0^{0.5}$$

$$= [(3.0 \times 0.5) - (1/2)(e^{-(2 \times 0.5)}] - [-(1/2)] = (1.5 - 0.184) + 0.5$$

$$= 1.816 \text{ cm}$$

(ii) In the second 0.5 hour

$$F_{p2} = \int_{0.50}^{1.0} (3.0 + e^{-2t})\, dt = \left[3.0t - \frac{1}{2}e^{-2t}\right]_{0.5}^{1.0}$$

$$= [(3.0 \times 1.0) - (1/2)(e^{-2})] - [(3.0 \times 0.5) - (1/2)(e^{-(2 \times 0.5)})]$$

$$= (3.0 - 0.0677) - (1.5 - 0.184) = 1.616 \text{ cm}$$

Example 3.9 *The infiltration capacity of soil in a small watershed was found to be 6 cm/h before a rainfall event. It was found to be 1.2 cm/h at the end of 8 hours of storm. If the total infiltration during the 8 hours period of storm was 15 cm, estimate the value of the decay coefficient K_h in Horton's infiltration capacity equation.*

Solution: Horton's equation is $f_p = f_c + (f_0 - f_c)\, e^{-K_h t}$

and
$$F_p = \int_0^t f_p(t)\, dt = f_c t + (f_0 - f_c) \int_0^t e^{-K_h t}\, dt$$

As $t \to \infty$, $\int_0^{\infty} e^{-K_h t}\, dt \to \frac{1}{K_h}$. Hence for large t values

$$F_p = f_c t + \frac{(f_0 - f_c)}{K_h}$$

Here F_p = 15.0 cm, f_0 = 6.0 cm, f_c = 1.2 cm and t = 8 hours.

$$15.0 = (1.2 \times 8) + (6.0 - 1.2)/K_h$$

$$K_h = 4.8/5.4 = 0.888 \text{ h}^{-1}$$

3.19 CLASSIFICATION OF INFILTRATION CAPACITIES

For purposes of runoff volume classification in small watersheds, one of the widely used methods is the *SCS–CN* method described in detail in Chapter 5. In this method, the soils are considered divided into four groups known as hydrologic soil groups. The steady state infiltration capacity, being one of the main parameters in this soil classification, is divided into four infiltration classes as mentioned below.

Table 3.11 Classification of Infiltration Capacities

Infiltration Class	Infiltration Capacity (mm/h)	Remarks
Very Low	< 2.5	Highly clayey soils
Low	2.5 to 25.0	Shallow soils, Clay soils, Soils low in organic matter
Medium	12.5 to 25.0	Sandy loam, Silt
High	>25.0	Deep sands, well drained aggregated soils

3.20 INFILTRATION INDICES

In hydrological calculations involving floods it is found convenient to use a constant value of infiltration rate for the duration of the storm. The defined average infiltration rate is called *infiltration index* and two types of indices are in common use.

φ—Index

The φ-index is the average rainfall above which the rainfall volume is equal to the runoff volume. The φ-index is derived from the rainfall hyetograph with the knowledge of the resulting runoff volume. The initial loss is also considered as infiltration. The φ value is found by treating it as a constant infiltration capacity. If the rainfall intensity is less than φ, then the infiltration rate is equal to the rainfall intensity; however, if the rainfall intensity is larger than φ the difference between the rainfall and infiltration in an interval of time represents the runoff volume as shown in Fig. 3.16. The amount of rainfall in excess of the index is called *rainfall excess*. In connection with runoff and flood studies it is also known as *effective rainfall*, (details in Sec. 6.5, Chapter 6). The φ–index thus accounts for the total abstraction and enables magnitudes to be estimated for a given rainfall hyetograph. A detailed procedure for calculating φ–index for a given storm hyetograph and resulting runoff volume is as follows.

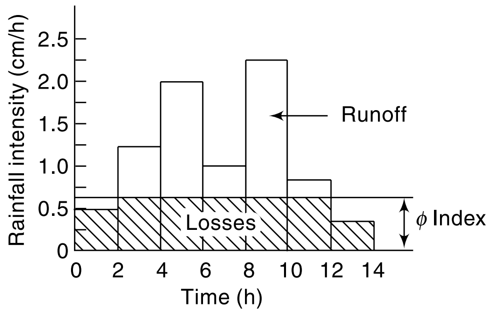


Fig. 3.16 φ-Index

Procedure for Calculation of φ-Index Consider a rainfall hyetograph of event duration D hours and having N pulses of time interval Δt such that

$N \cdot \Delta t = D$. (In Fig. 3.16, $N = 7$)

Let I_i be the intensity of rainfall in ith pulse and R_d = total direct runoff.

$$\text{Total Rainfall } P = \sum_1^N I_i \cdot \Delta t$$

If φ is φ-index, then $P - \varphi \cdot t_e = R_d$

where t_e = duration of rainfall excess.

If the rainfall hyetograph and total runoff depth R_d are given, the φ-index of the storm can be determined by trial and error procedure as given below.

1. Assume that out of given N pulses, M number of pulses have rainfall excess. (Note that $M \le N$). Select M number of pulses in decreasing order of rainfall intensity I_i.
2. Find the value of φ that satisfies the relation
$$R_d = \sum_1^M (I_i - \varphi)\Delta t$$
3. Using the value of φ of Step 2, find the number of pulses (M_c) which give rainfall excess. (Thus M_c = number of pulses with rainfall intensity $I_i \ge \varphi$).
4. If $M_c = M$, then φ of Step 2 is the correct value of φ-index. If not, repeat the procedure Step 1 onwards with new value of M. Result of Step 3 can be used as guidance to the next trial.

Example 3.10 illustrates this procedure in detail.

EXAMPLE 3.10 *A storm with 10 cm of precipitation produced a direct runoff of 5.8 cm. The duration of the rainfall was 16 hours and its time distribution is given below. Estimate the φ-index of the storm.*

Time from start (h)	0	2	4	6	8	10	12	14	16
Cumulative rainfall (cm)	0	0.4	1.3	2.8	5.1	6.9	8.5	9.5	10.0

SOLUTION: Pulses of uniform time duration $\Delta t = 2$ *h* are considered. The pulses are numbered sequentially and intensity of rainfall in each pulse is calculated as shown below.

Table 3.12 Calculations for Example 3.10

Pulse number	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**
Time from start of rain (h)	2	4	6	8	10	12	14	16
Cumulative rainfall (cm)	0.4	1.3	2.8	5.1	6.9	8.5	9.5	10.0
Incremental rain (cm)	0.40	0.90	1.50	2.30	1.80	1.60	1.00	0.50
Intensity of rain (I_i) in cm/h.	0.20	0.45	0.75	1.15	0.90	0.80	0.50	0.25

Here duration of rainfall $D = 16$ h, $\Delta t = 2$ h and $N = 8$.

Trial 1:

Assume $M = 8$, $\Delta t = 2$ h and hence $t_e = M \cdot \Delta t = 16$ hours.

Since $M = N$, all the pulses are included.

$$\text{Runoff } R_d = 5.8 \text{ cm} = \sum_1^8 (I_i - \varphi)\Delta t = \sum_1^8 I_i \times \Delta t - \varphi(8 \times 2)$$

$$5.8 = \{(0.20 \times 2) + (0.45 \times 2) + (0.75 \times 2) + (1.15 \times 2) + (0.90 \times 2) + (0.80 \times 2) + (0.50 \times 2) + (0.25 \times 2)\} - 16\,\varphi = 10.0 - 14\,\varphi$$

$$\varphi = 4.2/14 = 0.263 \text{ cm/h}$$

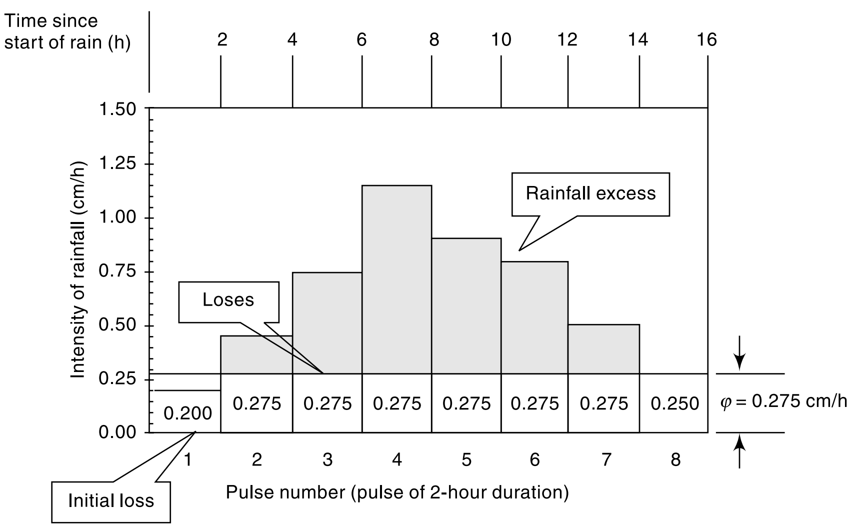


Fig. 3.17 Hyetograph and Rainfall Excess of the Storm – Example 3.10

By inspection of row 5 of Table 3.12, M_c = number of pulses having $I_i \geq \varphi$, that is $I_i \geq 0.263$ cm/h is 6.

Thus $M_c = 6 \neq M$. Hence assumed M is **not** correct. Try a new value of $M < 8$ in the next trial.

Trial 2:

Assume $M = 7$, $\Delta t = 2$ h and hence $t_e = M \cdot \Delta t = 14$ hours.

Select these 7 pulses in decreasing order of I_i. Pulse 1 is omitted.

$$\text{Runoff } R_d = 5.8 \text{ cm} = \sum_1^7 (I_i - \varphi)\,\Delta t = \sum_1^7 (I_i \cdot \Delta t - \varphi(7 \times 2)$$

$$5.8 = \{(0.45 \times 2) + (0.75 \times 2) + (1.15 \times 2) + (0.90 \times 2) + (0.80 \times 2) + (0.50 \times 2) + (0.25 \times 2)\} - 14\,\varphi = 9.6 - 14\,\varphi$$

$$\varphi = 3.8/14 = 0.271 \text{ cm/h}$$

By inspection of row 5 of Table 3.12, M_c = number of pulses having $I_i \geq \varphi$, that is $I_i \geq 0.271$ cm/h is 6.

Thus $M_c = 6 \neq M$. Hence assumed M is **not** O.K. Try a new value of $M < 7$ in the next trial.

Trial 3:

Assume $M = 6 \cdot \Delta t = 2$ h and hence $t_e = M \cdot \Delta t = 12$ hours.

Select these 6 pulses in decreasing order of I_i. Pulse 1 and 8 are omitted.

Runoff $R_d = 5.8$ cm = a

$$5.8 = \{(0.45 \times 2) + (0.75 \times 2) + (1.15 \times 2) + (0.90 \times 2) + (0.80 \times 2) + (0.50 \times 2)\} - 12\,\varphi = 9.1 - 12\,\varphi$$

$$\varphi = 3.3/12 = 0.275 \text{ cm/h}$$

By inspection of row 5 of Table 3. 12, M_c = number of pulses having $I_i \geq \varphi$, that is $I_i \geq 0.275$ cm/h is 6.

Thus $M_c = 6 = M$. Hence assumed M is OK.

The φ-index of the storm is 0.275 cm/h and the duration of rainfall excess = t_e = 12 hours. The hyetograph of the storm, losses, the rainfall excess and the duration of rainfall excess are shown in Fig. 3.17.

W-Index

In an attempt to refine the φ-index the initial losses are separated from the total abstractions and an average value of infiltration rate, called W-index, is defined as

$$W = \frac{P - R - I_a}{t_e} \tag{3.29}$$

where P = total storm precipitation (cm)

R = total storm runoff (cm)

I_a = Initial losses (cm)

t_e = duration of the rainfall excess, i.e. the total time in which the rainfall intensity is greater than W (in hours) and

W = defined average rate of infiltration (cm).

Since I_a rates are difficult to obtain, the accurate estimation of W-index is rather difficult.

The minimum value of the W-index obtained under very wet soil conditions, representing the constant minimum rate of infiltration of the catchment, is known as $W_{\min}$. It is to be noted that both the φ-index and W-index vary from storm to storm.

Computation of W-index To compute W-index from a given storm hyetograph with known values of I_a and runoff R, the following procedure is followed:

(i) Deduct the initial loss I_a from the storm hyetograph pulses starting from the first pulse

(ii) Use the resulting hyetograph pulse diagram and follow the procedure indicated in Sec. 3.19.1.

Thus the procedure is exactly same as in the determination of φ-index except for the fact that the storm hyetograph is appropriately modified by deducting I_a.

φ-INDEX FOR PRACTICAL USE The φ-index for a catchment, during a storm, depends in general upon the soil type, vegetal cover, initial moisture condition, storm duration and intensity. To obtain complete information on the interrelationship between these factors, a detailed expensive study of the catchment is necessary. As such, for practical use in the estimation of flood magnitudes due to critical storms a simplified relationship for φ-index is adopted. As the maximum flood peaks are invariably produced due to long storms and usually in the wet season, the initial losses are assumed to be negligibly small. Further, only the soil type and rainfall are found to be critical in the estimate of the φ-index for maximum flood producing storms.

On the basis of rainfall and runoff correlations, CWC[1] has found the following relationships for the estimation of φ-index for flood producing storms and soil conditions prevalent in India

$$R = \alpha I^{1.2} \tag{3.30}$$

$$\varphi = \frac{I - R}{24} \tag{3.31}$$

where R = runoff in cm from a 24-h rainfall of intensity I cm/h and α = a coefficient which depends upon the soil type as indicated in Table 3.13. In estimating the maximum floods for design purposes, in the absence of any other data, a φ-index value of 0.10 cm/h can be assumed.

Table 3.13 Variation of Coefficient α in Eq. 3.30

Sl. No.	Type of Soil	Coefficient α
1.	Sandy soils and sandy loam	0.17 to 0.25
2.	Coastal alluvium and silty loam	0.25 to 0.34
3.	Red soils, clayey loam, grey and brown alluvium	0.42
4.	Black–cotton and clayey soils	0.42 to 0.46
5.	Hilly soils	0.46 to 0.50

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Revision Questions

3.1 Discuss briefly the various abstractions from precipitation.
3.2 Explain briefly the evaporation process.
3.3 Discuss the factors that affect the evaporation from a water body.
3.4 Describe a commonly used evaporimeter.
3.5 Explain the energy budget method of estimating evaporation from a lake.
3.6 Discuss the importance of evaporation control of reservoirs and possible methods of achieving the same.
3.7 Describe the factors affecting evapotranspiration process.
3.8 List the various data needed to use Penman's equation for estimating the potential evapotranspiration from a given area.
3.9 Describe briefly (a) Reference crop evapotranspiration and (b) Actual evapotrans-piration.
3.10 Explain briefly the infiltration process and the resulting soil moisture zones in the soil.
3.11 Discuss the factors affecting the infiltration capacity of an area.
3.12 Describe the commonly used procedures for determining the infiltration characteristics of a plot of land. Explain clearly the relative advantages and disadvantages of the enumerated methods.
3.13 Describe various models adopted to represent the variation of infiltration capacity with time.
3.14 Explain a procedure for fitting Horton's infiltration equation for experimental data from a given plot.
3.15 Distinguish between
(a) Infiltration capacity and infiltration rate (b) Actual and potential evapotranspiration
(c) Field capacity and permanent wilting point (d) Depression storage and interception

Problems

3.1 Calculate the evaporation rate from an open water source, if the net radiation is 300 W/m^2 and the air temperature is 30° C. Assume value of zero for sensible heat, ground heat flux, heat stored in water body and advected energy. The density of water at 30° C = 996 kg/m^3.
[**Hint:** Calculate latent heat of vapourisation L by the formula:
$L = 2501 - 2.37\ T$ (kJ/kg), where T = temperature in °C.]

3.2 A class A pan was set up adjacent to a lake. The depth of water in the pan at the beginning of a certain week was 195 mm. In that week there was a rainfall of 45 mm and 15 mm of water was removed from the pan to keep the water level within the specified depth range. If the depth of the water in the pan at the end of the week was 190 mm calculate the pan evaporation. Using a suitable pan coefficient estimate the lake evaporation in that week.

3.3 A reservoir has an average area of 50 km^2 over an year. The normal annual rainfall at the place is 120 cm and the class A pan evaporation is 240 cm. Assuming the land flooded by the reservoir has a runoff coefficient of 0.4, estimate the net annual increase or decrease in the streamflow as a result of the reservoir.

3.4 At a reservoir in the neighbourhood of Delhi the following climatic data were observed. Estimate the mean monthly and annual evaporation from the reservoir using Meyer's formula.

Month	Temp. (°C)	Relative humidity (%)	Wind velocity at 2 m above GL (km/h)
Jan	12.5	85	4.0
Feb	15.8	82	5.0
Mar	20.7	71	5.0

(*Contd.*)

(*Contd.*)

Apr	27.0	48	5.0
May	31.0	41	7.8
Jun	33.5	52	10.0
Jul	30.6	78	8.0
Aug	29.0	86	5.5
Sep	28.2	82	5.0
Oct	28.8	75	4.0
Nov	18.9	77	3.6
Dec	13.7	73	4.0

3.5 For the lake in Prob. 3.4, estimate the evaporation in the month of June by (a) Penman formula and (b) Thornthwaite equation by assuming that the lake evaporation is the same as PET, given latitude = 28° N and elevation = 230 m above MSL. Mean observed sunshine = 9 h/day.

3.6 A reservoir had an average surface area of 20 km^2 during June 1982. In that month the mean rate of inflow = 10 m^3/s, outflow = 15 m^3/s, monthly rainfall = 10 cm and change in storage = 16 million m^3. Assuming the seepage losses to be 1.8 cm, estimate the evaporation in that month.

3.7 For an area in South India (latitude = 12° N), the mean monthly temperatures are given.

Month	**June**	**July**	**Aug**	**Sep**	**Oct**
Temp (°C)	31.5	31.0	30.0	29.0	28.0

Calculate the seasonal consumptive use of water for the rice crop in the season June 16 to October 15, by using the Blaney–Criddle formula.

3.8 A catchment area near Mysore is at latitude 12° 18′ N and at an elevation of 770 m. The mean monthly temperatures are given below.

Month	**Jan**	**Feb**	**Mar**	**Apr**	**May**	**Jun**	**Jul**	**Aug**	**Sep**	**Oct**	**Nov**	**Dec**
Mean monthly temp. (°C)	22.5	24.5	27.0	28.0	27.0	25.0	23.5	24.0	24.0	24.5	23.0	22.5

Calculate the monthly and annual PET for this catchment using the Thornthwaite formula.

3.9 A wheat field has maximum available moisture of 12 cm. If the reference evapotranspiration is 6.0 mm/day, estimate the actual evapotranspiration on Day 2, Day 7 and Day 9 after irrigation. Assume soil-water depletion factor $p = 0.20$ and crop factor $K = 0.65$.

3.10 Results of an infiltrometer test on a soil are given below. Determine the Horton's infiltration capacity equation for this soil.

Time since start in (h)	0.25	0.50	0.75	1.00	1.25	1.50	1.75	2.0
Infiltration capacity in cm/h	5.6	3.20	2.10	1.50	1.20	1.10	1.0	1.0

3.11 Results of an infiltrometer test on a soil are given below. Determine the best values of the parameters of Horton's infiltration capacity equation for this soil.

Time since start in minutes	5	10	15	20	30	40	60	80	100
Cumulative infiltration in mm	21.5	37.7	52.2	65.8	78.4	89.5	101.8	112.6	123.3

3.12 Results of an infiltrometer test on a soil are as follows:

Time since start in minutes	5	10	15	20	30	40	60	120	150
Cumulative infiltration in mm	1.00	1.80	2.50	3.10	4.20	5.10	6.60	11.00	12.90

Determine the parameters of (i) Kostiakov's equation, (ii) Green–Ampt equation, and (iii) Philips equation

3.13 Determine the best values of the parameters of Horton's infiltration capacity equation for the following data pertaining to infiltration tests on a soil using double ring infiltrometer.

Time since start in minutes	5	10	15	25	40	60	75	90	110	130
Cumulative infiltration in mm	21.0	36.0	47.6	56.9	63.8	69.8	74.8	79.3	87.0	92.0

3.14 For the infiltration data set given below, establish (a) Kostiakov's equation, (b) Philips equation, and (c) Green–Ampt equation.

Time since start in minutes	10	20	30	50	80	120	160	200	280	360
Cumulative Infiltration in mm	9.8	18.0	25.0	38.0	55.0	76.0	94.0	110.0	137.0	163.0

3.15 Following table gives the values of a field study of infiltration using flooding type infiltrometer. (a) For this data plot the curves of (i) infiltration capacity f_p (mm/h) *vs* time (h) on a log–log paper and obtain the equation of the best fit line, and (ii) Cumulative infiltration (mm) F_p *vs* time (h) on a semi-log paper and obtain the equation of the best fit line. (b) Establish Horton's infiltration capacity equation for this soil.

Time since start in minutes	2	10	30	60	90	120	240	360
Cumulative Infiltration in cm	7.0	20.0	33.5	37.8	39.5	41.0	43.0	45.0

3.16 The infiltration capacity of a catchment is represented by Horton's equation as

$$f_p = 0.5 + 1.2e^{-0.5t}$$

where f_p is in cm/h and t is in hours. Assuming the infiltration to take place at capacity rates in a storm of 4 hours duration, estimate the average rate of infiltration for the duration of the storm.

3.17 The infiltration process at capacity rates in a soil is described by Kostiakov's equation as $F_p = 3.0\,t^{0.7}$ where F_p is cumulative infiltration in cm and t is time in hours. Estimate the infiltration capacity at (i) 2.0 h and (ii) 3.0 h from the start of infiltration.

3.18 The mass curve of an isolated storm in a 500 ha watershed is as follows:

Time from start (h)	0	2	4	6	8	10	12	14	16	18
Cumulative rainfall (cm)	0	0.8	2.6	2.8	4.1	7.3	10.8	11.8	12.4	12.6

If the direct runoff produced by the storm is measured at the outlet of the watershed as 0.340 Mm^3, estimate the φ-index of the storm and duration of rainfall excess.

3.19 The mass curve of an isolated storm over a watershed is given below.

Time from start (h)	0	0.5	1.0	1.5	2.0	2.5	3.0	3.5	4.0	4.5	5.0
Cummulative rainfall (cm)	0	0.25	0.50	1.10	1.60	2.60	3.50	5.70	6.50	7.30	7.70

If the storm produced a direct runoff of 3.5 cm at the outlet of the watershed, estimate the φ-index of the storm and duration of rainfall excess.

3.20 In a 140-min storm the following rates of rainfall were observed in successive 20-min intervals: 6.0, 6.0, 18.0, 13.0, 2.0, 2.0 and 12.0 mm/h. Assuming the φ-index value as 3.0 mm/h and an initial loss of 0.8 mm, determine the total rainfall, net runoff and W-index for the storm.

3.21 The mass curve of rainfall of duration 100 min is given below. If the catchment had an initial loss of 0.6 cm and a φ-index of 0.6 cm/h, calculate the total surface runoff from the catchment.

Time from start of rainfall (min)	0	20	40	60	80	100
Cummulative rainfall (cm)	0	0.5	1.2	2.6	3.3	3.5

3.22 An isolated 3-h storm occurred over a basin in the following fashion:

% of catchment area	φ–index (cm/h)	Rainfall (cm)		
		1st hour	2nd hour	3rd hour
20	1.00	0.8	2.3	1.5
30	0.75	0.7	2.1	1.0
50	0.50	1.0	2.5	0.8

Estimate the runoff from the catchment due to the storm.

Objective Questions

3.1 If e_w and e_a are the saturated vapour pressures of the water surface and air respectively, the Dalton's law for evaporation E_L in unit time is given by E_L =
(a) $(e_w - e_a)$ (b) $K e_w e_a$ (c) $K(e_w - e_a)$ (d) $K(e_w + e_a)$

3.2 The average pan coefficient for the standard US Weather Bureau class A pan is
(a) 0.85 (b) 0.70 (c) 0.90 (d) 0.20

3.3 A canal is 80 km long and has an average surface width of 15 m. If the evaporation measured in a class A pan is 0.5 cm/day, the volume of water evaporated in a month of 30 days is (in m^3)
(a) 12600 (b) 18000 (c) 180000 (d) 126000

3.4 The ISI standard pan evaporimeter is the
(a) same as the US class A pan
(b) has an average pan coefficient value of 0.60

(c) has less evaporation than a US class A pan
(d) has more evaporation than a US class A pan.

3.5 The chemical that is found to be most suitable as water evaporation inhibitor is
(a) ethyl alcohol (b) methyl alcohol
(c) cetyl alcohol (d) butyl alcohol.

3.6 Wind speed is measured with
(a) a wind vane (b) a heliometer
(c) Stevenson box (d) anemometer

3.7 If the wind velocity at a height of 2 m above ground is 5.0 kmph, its value at a height of 9 m above ground can be expected to be in km/h about
(a) 9.0 (b) 6.2 (c) 2.3 (d) 10.6

3.8 Evapotranspiration is confined
(a) to daylight hours (b) night-time only
(c) land surfaces only (d) none of these.

3.9 Lysimeter is used to measure
(a) infiltration (b) evaporation (c) evapotranspiration (d) vapour pressure.

3.10 The highest value of annual evapotranspiration in India is at Rajkot, Gujarat. Here the annual PET is about
(a) 150 cm (b) 150 mm (c) 210 cm (d) 310 cm.

3.11 Interception losses
(a) include evaporation, through flow and stemflow
(b) consists of only evaporation loss
(c) includes evaporation and transpiration losses
(d) consists of only stemflow.

3.12 The infiltration capacity of a soil was measured under fairly identical general conditions by a flooding type infiltrometer as f_f and by a rainfall simulator as f_r. One can expect
(a) $f_f = f_r$ (b) $f_f > f_r$ (c) $f_f < f_r$ (d) no fixed pattern.

3.13 A watershed 600 ha in area experienced a rainfall of uniform intensity 2.0 cm/h for duration of 8 hours. If the resulting surface runoff is measured as 0.6 Mm^3, the average infiltration capacity during the storm is
(a) 1.5 cm/h (b) 0.75 cm/h (c) 1.0 cm/h (d) 2.0 cm/h

3.14 In a small catchment the infiltration rate was observed to be 10 cm/h at the beginning of the rain and it decreased exponentially to an equilibrium value of 1.0 cm/h at the end of 9 hours of rain. If a total of 18 cm of water infiltered during 9 hours interval, the value of the decay constant K_h in Horton's infiltration equation in (h^{-1}) units is
(a) 0.1 (b) 0.5 (c) 1.0 (d) 2.0

3.15 In Horton's infiltration equation fitted to data from a soil, the initial infiltration capacity is 10 mm/h, final infiltration capacity is 5 mm/h and the exponential decay constant is 0.5 h^{-1}. Assuming the infiltration takes place at capacity rates, the total infiltration depth for a uniform storm of duration 8 hours is
(a) 40 mm (b) 60 mm (c) 80 mm (d) 90 mm

3.16 The rainfall on five successive days on a catchment was 2, 6, 9, 5 and 3 cm. If the φ-index for the storm can be assumed to be 3 cm/day, the total direct runoff from the catchment is
(a) 20 cm (b) 11 cm (c) 10 cm (d) 22 cm

3.17 A 6-h storm had 6 cm of rainfall and the resulting runoff was 3.0 cm. If the φ-index remains at the same value the runoff due to 12 cm of rainfall in 9 h in the catchment is
(a) 9.0 cm (b) 4.5 cm (c) 6.0 cm (d) 7.5 cm

3.18 For a basin, in a given period Δt, there is no change in the groundwater and soil water status. If P = precipitation, R = total runoff, E = Evapotranspiration and ΔS = increase in the surface water storage in the basin, the hydrological water budget equation states
(a) $P = R - E \pm \Delta S$ (b) $R = P + E - \Delta S$ (c) $P = R + E + \Delta S$ (d) None of these

Chapter
4

STREAMFLOW MEASUREMENT

4.1 INTRODUCTION

Streamflow representing the runoff phase of the hydrologic cycle is the most important basic data for hydrologic studies. It was seen in the previous chapters that precipitation, evaporation and evapotranspiration are all difficult to measure exactly and the presently adopted methods have severe limitations. In contrast the measurement of streamflow is amenable to fairly accurate assessment. Interestingly, streamflow is the only part of the hydrologic cycle that can be measured accurately.

A stream can be defined as a flow channel into which the surface runoff from a specified basin drains. Generally, there is considerable exchange of water between a stream and the underground water. Streamflow is measured in units of discharge (m^3/s) occurring at a specified time and constitutes historical data. The measurement of discharge in a stream forms an important branch of *Hydrometry*, the science and practice of water measurement. This chapter deals with only the salient streamflow measurement techniques to provide an appreciation of this important aspect of engineering hydrology. Excellent treatises[1, 2, 4, 5] and a bibliography[6] are available on the theory and practice of streamflow measurement and these are recommended for further details.

Streamflow measurement techniques can be broadly classified into two cate-gories as (i) direct determination and (ii) indirect determination. Under each category there are a host of methods, the important ones are listed below:

1. Direct determination of stream discharge:
   - (a) Area-velocity methods,
   - (b) Dilution techniques,
   - (c) Electromagnetic method, and
   - (d) Ultrasonic method.
2. Indirect determination of streamflow:
   - (a) Hydraulic structures, such as weirs, flumes and gated structures, and
   - (b) Slope-area method.

Barring a few exceptional cases, continuous measurement of stream discharge is very difficult. As a rule, direct measurement of discharge is a very time-consuming and costly procedure. Hence, a two step procedure is followed. First, the discharge in a given stream is related to the elevation of the water surface (Stage) through a series of careful measurements. In the next step the stage of the stream is observed routinely in a relatively inexpensive manner and the discharge is estimated by using the previously determined stage–discharge relationship. The observation of the stage is easy, inexpensive, and if desired, continuous readings can also be obtained. This method of discharge determination of streams is adopted universally.