Water Losses

The various water losses that occur in nature are:

(i) Interception loss-due to surface vegetation, i.e., held by plant leaves.

(ii) Evaporation:

(a) from water surface, i.e., reservoirs, lakes, ponds, river channels, etc.

(b) from soil surface, appreciably when the ground water table is very near the soil surface.

(iii) Transpiration—from plant leaves.

(iv) Evapotranspiration for consumptive use—from irrigated or cropped land.

(v) Infiltration—into the soil at the ground surface.

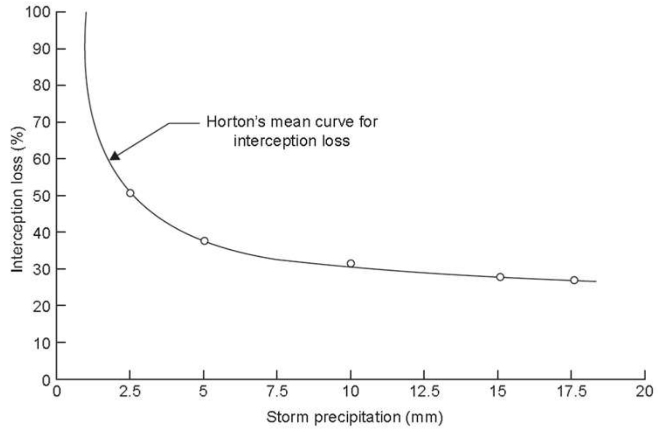
(vi) Watershed leakage—ground water movement from one basin to another or into the sea.

The various water losses are discussed below:

Interception loss:

The precipitation intercepted by plant leaves and buildings and returned to atmosphere (by evaporation) without reaching the ground surface is called interception loss. Interception loss is high in the beginning of storms and gradually decreases; the loss is of the order of 0.5 to 2 mm per shower and it is greater in the case of light showers than when rain is continuous. Figure below shows the *Horton's* mean curve of interception loss for different showers.

Effective rain = Rainfall – Interception loss



Interception loss (Horton)

Evaporation:

Evaporation is the rate of liquid water transformation to vapor from open water, bare soil, or covered soil with vegetation. Evaporation usually stated by millimeters of evaporated water per day.

Evaporation from free water surfaces and soil are of great importance in water resources studies. It affects the yield of river basins, the necessary capacity of reservoirs, the size of pumping plant, the consumptive use of water by crops …etc.

Potential evaporation is defined as the quantity of water evaporated per unit area, per unit time from an idealized, extensive free water surface under exiting atmospheric conditions.

Factors affecting evaporation are:

Solar radiation. The change of state of water from liquid to gas requires an energy input (known as the latent heat of vaporization), this energy comes from the sun.

Wind. As the water vaporize into the boundary layer between earth and air becomes saturated and it must be removed and continually replaced by dryer air if evaporation to be proceed. This movement of air in the boundary layer depends on wind and so the wind speed is important.

Relative humidity. As the air humidity rises, its ability to absorb more water vapor decreases and the rate of evaporation slows.

Temperature. An energy input is necessary for evaporation to proceed. If the temperature of air and ground is high, evaporation will proceed more rapidly than if they are cool, also the air capacity to absorb water vapor increases as its temperature rises.

Transpiration and evapo-transpiration.

Growing vegetation of all kind needs water to sustain life, only a small fraction of the water needed by plant is retained in the plant structure. Most of it passes through the roots to the stem and is transpired into the atmosphere through the leafy part of the plant.

In field condition it is practically impossible to differentiate between evaporation and transpiration if the ground is covered with vegetation. The two processes are commonly linked together and referred to as *evapo-transpiration* (E_{pt}) .

Evapotranspiration (E_t) or consumptive use is the total water lost from a cropped (or irrigated) land due to evaporation from the soil and transpiration by the plants or used by the plants in building up of plant tissue.

Potential Evapotranspiration is the evapotranspiration that would occur from a well vegetated surface when moisture supply is not limited, and this is calculated in a way similar to that for open water evaporation. Actual Evapotranspiration drops below its potential level as the soil dries out.

Evaporation from water surfaces (Lake Evaporation)

The factors affecting evaporation are air and water temperature, relative humidity, wind velocity, surface area (exposed), barometric pressure and salinity of the water, the last two having a minor effect. The rate of evaporation is a function of the differences in vapour pressure at the water surface and in the atmosphere, and the Dalton's law of evaporation is given by (mass transfer)

$$E = K\,(e_s - e_a)$$

Where: E = daily evaporation
e_s = saturated vapour pressure at the temperature of water
e_a = vapour pressure of the air (about 2 m above)
K = a constant.

the Dalton's law states that the evaporation is proportional to the difference in vapour pressures e_s and e_a. A more general form of the above Eq. is given by

$$E = K\,(e_s - e_a)(a + bV)$$

Where: K', a, b = constants and V = wind velocity.

Higher the temperature and wind velocity, greater is the evaporation, while greater the humidity and dissolved salts, smaller is the evaporation.

Definitions:

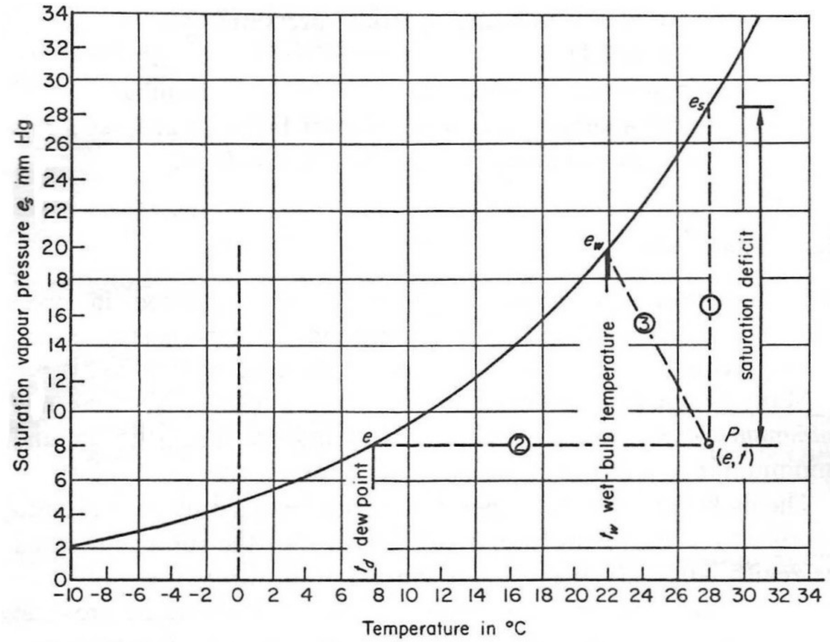
Air easily absorbs moisture in the form of water vapor. The amount of water vapour absorbed by air depends on the temperature of the air and of the water. The greater the temperature the more vapour the air can contain. The water vapor exerts a *potential pressure* usually measured in bars (1bar=100000N/m2) or mm height of a mercury column (Hg), (1mm Hg=1.36 mbar).

Suppose an evaporating surface of water is in a closed system and enveloped in air. If a source of heat energy is available to the system, evaporation of the water into the air will take place until a state of equilibrium is reached when the air is saturated with vapor and can absorbed no more. The molecules of water vapour will then exert a pressure which is known as *saturation vapour pressure (e_s),* for the particular temperature of the system.

The value of e_s changes with temperature as indicated in figure below. Referring to this figure, consider what can happen to a mass of atmospheric air *P*, whose temp is *t* and whose vapor pressure is e_a. Since *P* is lies below the saturation curve, it is clear that the air mass could absorb more water vapor and that is it did so while its temperature remained constant, then the position of *P* would move vertically up dashed line 1 until the air was saturated. The corresponding vapor pressure of *P* in the new position would be e_s. The increase (e_s- e_a) is known as the *saturation deficit*.

Alternatively, if no change were to take place in the humidity of the air while it was cooled, then P would move horizontally to the left along line 2 until the saturation line was intersected again. At this point *P* would be saturated, at a new temperature t_d, the *dewpoint*. Cooling of the air beyond this point would result in condensation or mist being formed.

If the water is allowed to evaporate freely into the air mass, neither of the above two possibilities occurs. This is because the evaporation requires heat which is withdrawn from the air itself. This heat called the latent heat of evaporation, h_r, and given by:



Saturation vapour pressure of water in air

$$h_r = 606.5 - 0.695t \qquad cal/g$$

⇨ the latent heat is the amount of heat absorbed by a unit mass of water, without changing in temperature , while passing from the liquid to the vapor state.

So, as the humidity and vapor pressure rise, the temperature of the air falls and the point *P* moves diagonally along line 3 until saturation vapour pressure is reached at the point defined by e_w and t_w. This temperature t_w is called the *wet bulb temperature* and is the temperature to which the original air can be cooled by evaporating water into it. This is the temperature found by a wet bulb thermometer.

The relative humidity is now given as:

$$h = \frac{e_a}{e_s} \text{ , or as percentage, } h = \frac{e_a}{e_s} \times 100$$

Example:
An air mass is at temperature of 28 ^{0}C with relative humidity of 70%. Determine:
a. Saturation vapour pressure.
b. Saturation deficit.
c. Actual vapor pressure.
d. Dewpoint.
e. Wet bulb temperature.

Solution:
T= 28 ^{0}C ⇨ from the vapour pressure curve ⇨ es =28.3 mmHg.

H=70% ⇨ ea =28.3x0.7 = 19.81 mmHg.

$\therefore e_s - e_a$=28.3-19.81=8.50 mmHg

t_d ⇨ from same curve ⇨ t_d=22 ^{0}C

From the same curve ⇨ wet bulb temp=24.7^{0}C

Methods of Estimating Lake Evaporation

Evaporation from water surfaces can be determined from the following methods :

(i) The storage equation (water budget equation)

$$P + I \pm G = E + O \pm \Delta S$$

Where:
P = Precipitation
I = surface inflow
G = subsurface inflow or outflow
E = evaporation
O = surface outflow
ΔS = change in surface water storage

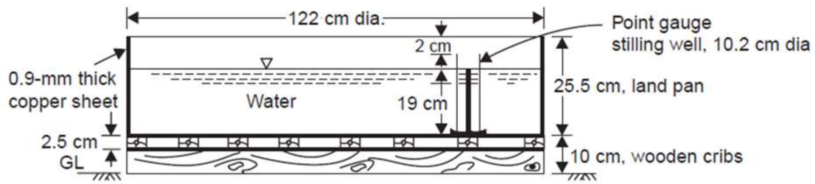
(ii) Auxiliary pans like land pans, floating pans, Colorado sunken pans, etc.
(iii) Evaporation formula like that of Dalton's law.
(iv) Humidity and wind velocity gradients.
(v) The energy budget, this method involves too many hydro-meteorological factors (variables) with too much sophisticated instrumentation and hence it is a specialist approach
(vi) Combination of aerodynamic and energy balance equations, Penman's equation (involves too many variables)

EVAPORATION PANS:

(i) *Floating pans* (made of GI) of 90 cm square and 45 cm deep are mounted on a raft floating in water. The volume of water lost due to evaporation in the pan is determined by knowing the volume of water required to bring the level of water up to the original mark daily and after making allowance for rainfall, if there has been any.

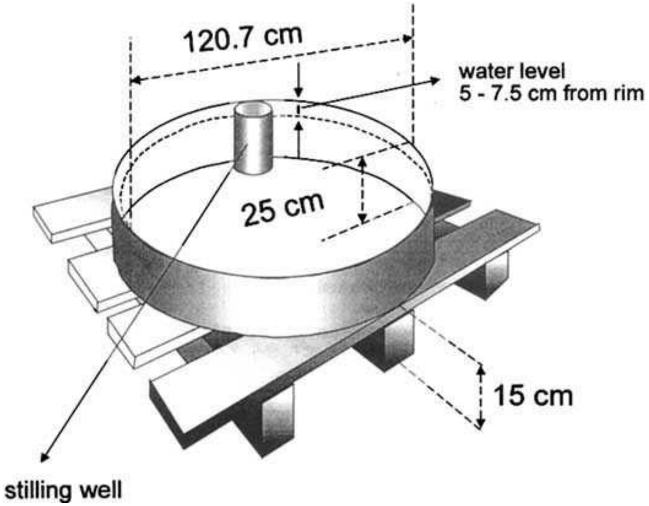
(ii) *Colorado sunken pan*. This is 92 cm square and 42-92 cm deep and is sunk in the ground such that only 5-15 cm depth projects above the ground surface and thus the water level is maintained almost at the ground level. The evaporation is measured by a point gauge

(iii) *Land pan*. Evaporation pans are installed in the vicinity of the reservoir or lake to determine the lake evaporation. The IMD Land pan shown in Figure below is 122 cm diameter and 25.5 cm deep made of unpainted GI; and set on wood grillage 10 cm above ground to permit circulation of air under the pan. The pan has a stilling well, Vernier point gauge, a thermometer with clip and may be covered with a wire screen. The amount of water lost by evaporation from the pan can be directly measured by the point gauge. The air temperature is determined by reading a dry bulb thermometer. An anemometer is normally mounted at the level of the instrument to provide the wind speed information required. Allowance has to be made for rainfall, if there has been any. Water is added to the pan from a graduated cylinder to bring the water level to the original mark (5 cm below the top of the pan).



IMD land pan

In the USA the standard or class A pan is (very similar to the IMD land pan) circular 1.22m in diameter and 254mm deep, filled to a depth of 180mm, set on a timber grillage with the pan bottom 150mm above ground level.



Pan coefficient: Evaporation pan data cannot be applied to free water surfaces directly but must be adjusted for the differences in physical and climatological factors. For example, a lake is larger and deeper and may be exposed to different wind speed, as compared to a pan. The small volume of water in the metallic pan is greatly affected by temperature fluctuations in the air or by solar radiations in contrast with large bodies of water (in the reservoir) with little temperature fluctuations. Thus the pan evaporation data have to be corrected to obtain the actual evaporation from water surfaces of lakes and reservoirs, i.e., by multiplying by a coefficient called pan coefficient and is defined as:

$$pan\ coefficient = \frac{Lake\ Evaporation}{Pan\ Evaporation}$$

The experimental values for pan coefficients range from 0.67 to 0.82 with an average of 0.7.

<u>Example:</u>

Compute the daily evaporation from a Class A pan if the amounts of water added to bring the level to the fixed point are as follows:

Day:	1	2	3	4	5	6	7
Rainfall (mm):	14	6	12	8	0	5	6
Water (added or removed):	-5	3	0	0	7	4	3

What is the evaporation loss of water in this week from a lake (surface area = 640 ha) in the vicinity, assuming a pan coefficient of 0.75?

<u>Solution:</u>

Pan evaporation, Ep, mm = Rainfall + water added or water removed

Day:	1	2	3	4	5	6	7
Ep(mm):	14-5=9	6+3=9	12	8	7	5+4=9	6+3=9

⇨ pan evaporation in the week = $\sum_1^7 Ep$ = 63 mm

Pan coefficient $=\frac{E_{Lake}}{E_{pan}}=0.75$

⇨ Lake evaporation in the week= 63x0.75= 47.25 mm

Water lost from the lake = 640 x (47.25/1000) =30.24 ha.m

$\simeq 0.3\ Mm^3$

Penman Equation:

Penman derived the following eq.

$$E = \frac{\Delta}{\Delta + \gamma} Q_n + \frac{\gamma}{\Delta + \gamma} E_a$$

Where:

Δ is slope of the saturation-vapor-pressure versus temperature curve at the air temperature T_a,

E_a is the evaporation given by Dalton assuming the water surface T_o=T_a, $E_a = (e_s - e_a)(a + bv)$.

Q_n is the net radiation absorbed by water body expressed in the same unit as E.

$$\gamma = 0.00066\, p$$

$$\Delta = (0.00815 T_a + 0.8912)^7$$

p is the atmospheric pressure, p=1013 mbar.

Ta is the air temperature ^{0}C.

Knowing that : $\frac{\Delta}{\Delta+\gamma} + \frac{\gamma}{\Delta+\gamma} = 1$

$$Q_n = 7.14 \times 10^{-3} Q_s + 5.26 \times 10^{-6} Q_s (T_a + 17.8)^{1.87} + 3.94 \times 10^{-6} Q_s^{\ 2} - 2.39 \times 10^{-9} Q_s^{\ 2} (T_a - 7.2)^2 - 1.02$$

Q_n is the net radiation, expressed in equivalent millimeters of evaporation per day.

Q_s is the daily solar radiation in calories per square centimeter per day.

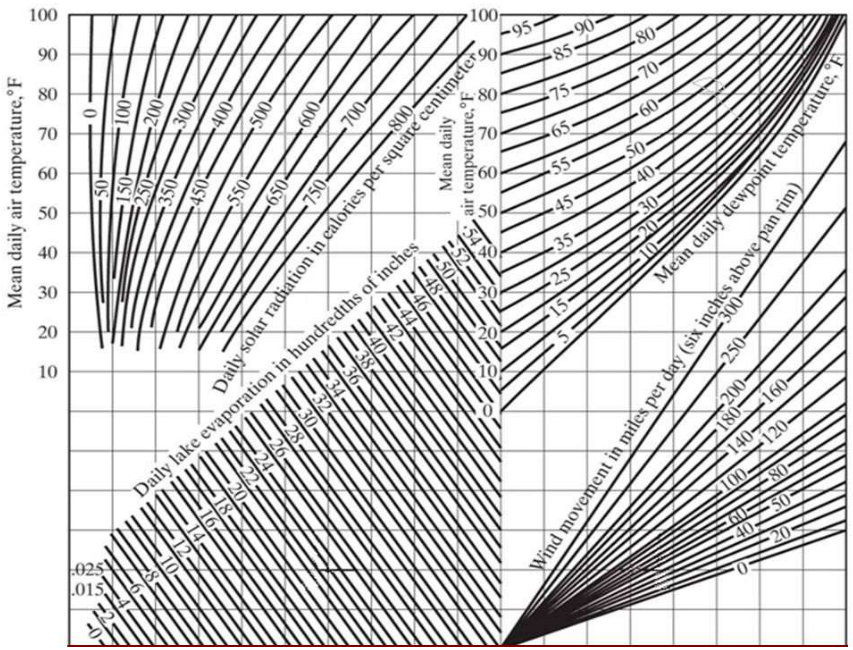
$$e_s - e_a = 33.86[(0.00738 T_a + 0.8072)^8 - (0.00738 T_d + 0.8072)^8]$$

Where; T_d is dewpoint, ^{0}C.

$$E_a = (e_s - e_a)^{0.88} (0.42 + 0.0029 v_p)$$

v_p is the wind movement 150mm above the pan rim (km/day)

Figure below is a graphical solution of the Penman equation for the estimation of lake evaporation as a function of solar radiation, air temperature, dewpoint and wind movement.



Shallow-lake evaporation as a function of solar radiation, air temperature, dewpoint, and wind speed according to Penman's Eq.

Example:

Calculate lake evaporation for the following data using Penman's equation.

v_p=130 km/day, T_a=18 ^{0}C, T_d=8 ^{0}C, Qs=450 cal./cm^2/day.

Solution:

$$e_s - e_a = 33.86[(0.00738T_a + 0.8072)^8 - (0.00738T_d + 0.8072)^8]$$
$$= 33.86[(0.00738 \times 18 + 0.8072)^8 - (0.00738 \times 8 + 0.8072)^8]$$
$$= 9.91$$

$$E_a = (e_s - e_a)^{0.88}(0.42 + 0.0029v_p)$$
$$\Rightarrow E_a = (9.91)^{0.88}(0.42 + 0.0029 \times 130) = 6.00$$

$$\gamma = 0.00066\,p = 0.00066 \times 1013 = 0.668$$
$$\Delta = (0.00815T_a + 0.8912)^7 = (0.00815 \times 18 + 0.8912)^7 = 1.038$$

$$\Rightarrow \frac{\Delta}{\Delta+\gamma} = 0.608 \quad \Rightarrow \frac{\gamma}{\Delta+\gamma} = 0.392$$

$$Q_n = 4.84 \text{ equivalent mm/day}$$

⇨ E = 0.608x4.84+0.392x6.00= 5.30 mm/day =0.208″/day

Estimation of Evapotranspiration

The following are some of the methods of estimating evapotranspiration:

(i) Tanks and lysimeter experiments.
(ii) Installation of sunken (Colorado) tanks.
(iii) Evapotranspiration equations as developed by Lowry-Johnson, Penman, Thornthwaite, Blaney Criddle, etc.
(iv) Evaporation index method, i.e., from pan evaporation data as developed by Hargreaves and Christiansen.

Thornthwaite's Formula:

Thornthwaite carried out many experiments using lysimeters. He suggested a method for the estimation of evapo-transpiration in the latitudes of USA.

If t_n=average monthly temperature of the consecutive months of the year in ^{0}C (n=1, 2, 3, ….,12), and j=monthly heat index.

Then $j = 0.09{t_n}^{1.5}$

$J = \sum_1^{12} j$ (for 12 months)

The potential evapotranspiration for any month with average temperature $t\,^0$C is given, as P_{ex} , by:

$$PE_x = 16\left(\frac{10\,t}{J}\right)^a$$

Where : $a = 0.016J + 0.5$

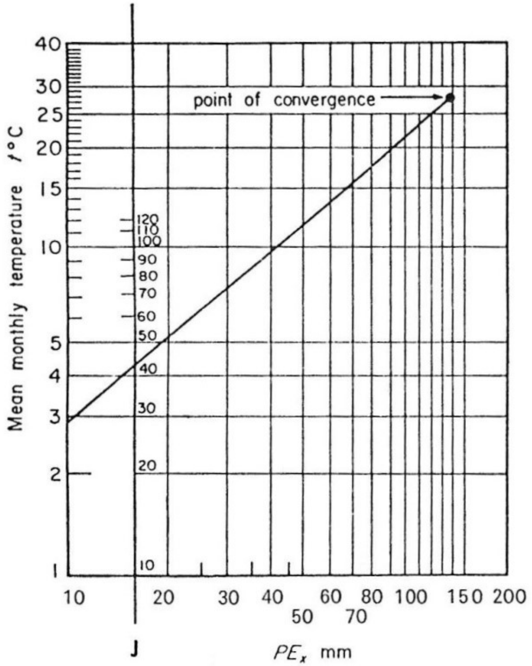
PE_x is a theoretical standard monthly value based on 30 days and 12hr sunrise per day. The actual PE for a certain month is given by:

$$PE = PE_x\frac{DT}{360}$$

Where: D= number of days in the month.

T= average number of hours between sunrise and sunset in the month.

Thornthwaite published a nomogram and table for the solution of his formula, the nomogram shown in figure below.



t °C	PEx
26·5	135·0
27·0	139·5
27·5	143·7
28·0	147·8
28·5	151·7
29·0	155·4
29·5	158·9
30·0	162·1
30·5	165·2
31·0	168·0
31·5	170·7
32·0	173·1
32·5	175·3
33·0	177·2
33·5	179·0
34·0	180·5
34·5	181·8
35·0	182·9
35·5	183·7
36·0	184·3
36·5	184·7
37·0	184·9
37·5	185·0
38·0	185·0

Nomogram and table for finding potential evapo-transpiration

Example:

Compute the potential evapo-transpiration for April and November according to Thornthwaite for the following data:

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
tn °C	-5	0	5	9	13	17	19	17	13	9	5	0
t °C	--	--	--	10	--	--	--	--	--	--	3	--
Day hours	--	--	--	13	--	--	--	--	--	--	9	--

Solution:

Month	tn °C	$j = 0.09t_n{}^{1.5}$	PEx mm/month	PE mm/month
Jan	-5	--	--	--
Feb	0	0.00	--	--
Mar	5	1.01	--	--
Apr	9	2.43	48.43	52.47
May	13	4.22	--	--
Jun	17	6.31	--	--
Jul	19	7.45	--	--
Aug	17	6.31	--	--
Sep	13	4.22	--	--
Oct	9	2.43	--	--
Nov	5	1.01	13.42	10.1
Dec	0	0.00	--	--
	Σ	35.38		

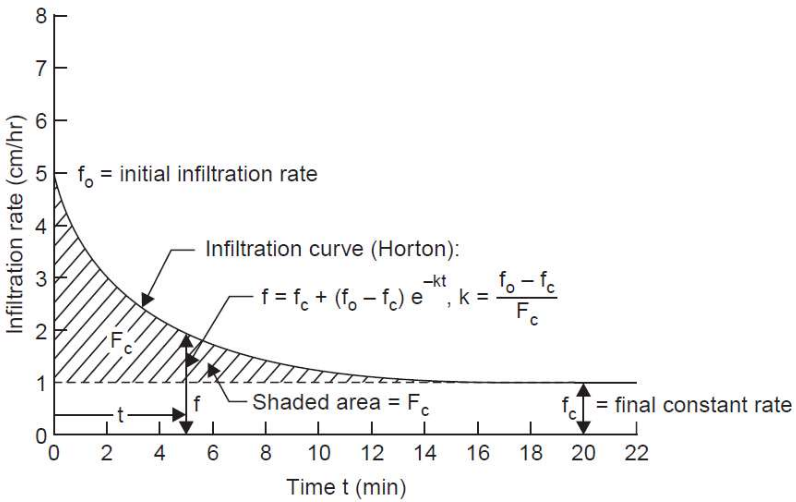
$$a = 0.016J + 0.5 = 0.016 \times 35.38 + 0.5 = 1.066$$

Infiltration:

Water entering the soil at the ground surface is called infiltration. It complements the soil moisture deficiency and the excess moves downward by the force of gravity called deep seepage or percolation and builds up the ground water table. The maximum rate at which the soil in any given condition is capable of absorbing water is called its infiltration capacity *(f_p)*. Infiltration rate *(f)* often begins at a high rate and decreases to a fairly steady state rate *(f_c)* as the rain continues, figure below. The infiltration rate *(f)* at any time t is given by Horton's equation.

$$f = f_c + (f_o - f_c)e^{-kt}$$

f_0 = initial rate of infiltration capacity
f_c = final constant rate of infiltration at saturation
k = a constant depending primarily upon soil and vegetation
F_c = shaded area in Fig.
t = time from beginning of the storm



Infiltration Curve (Horton)

The infiltration takes place at capacity rates only when the intensity of rainfall equals or exceeds fp;
$f = fp$ when $i \geq fp$
But $f = i$ when $i < fp$

The infiltration depends upon the intensity and duration of rainfall, weather (temperature), soil characteristics, vegetal cover, land use, initial soil moisture content (initial wetness), entrapped air and depth of the ground water table. The vegetal cover provides protection against rain drop impact and helps to increase infiltration.

Methods of Determining Infiltration

The methods of determining infiltration are:
(i) Infiltrometers.
(ii) Observation in pits and ponds.
(iii) Placing a catch basin below a laboratory sample.
(iv) Artificial rain simulators.
(v) Hydrograph analysis.

Double-ring infiltrometer. A double ring infiltrometer is shown in Figure (A) below. The two rings (22.5 to 90 cm diameter) are driven into the ground by a driving plate and hammer, to penetrate into the soil uniformly without tilt to a depth of 15 cm. After driving is over, any disturbed soil adjacent to the sides tamped with a metal tamper. Point gauges are fixed in the center of the ring. Water is poured into the rings to maintain the desired depth (2.5 to 15 cm with a minimum of 5 mm) and the water added to maintain the original constant depth at regular time intervals (after the starting of the experiment) of 5, 10, 15, 20, 30, 40, 60 min, etc. up to a period of at least 6 hours is noted and the results are plotted as infiltration rate in cm/hr versus time in minutes as shown in Figure (B) below. The purpose of the outer tube is to eliminate to some extent the edge effect of the surrounding drier soil and to prevent the water within the inner space from spreading over a larger area after penetrating below the bottom of the ring.

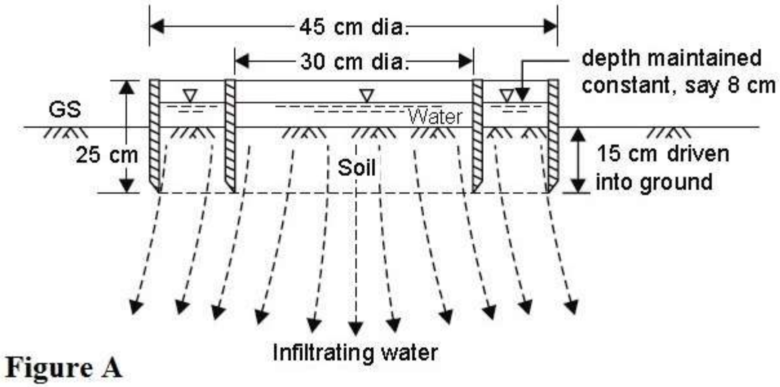


Figure A

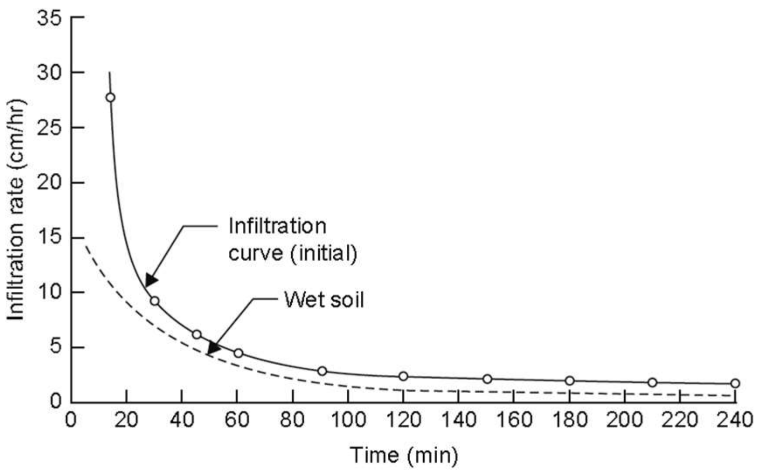


Figure B Typical infiltration curve

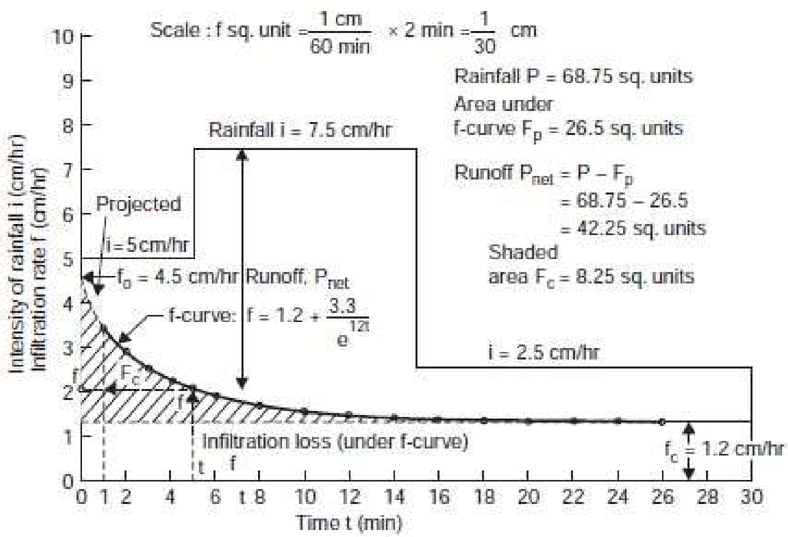
<u>Example:</u>

For a given basin, the following are the infiltration capacity rates at various time intervals after the beginning of the storm. Make a plot of the f-curve and establish an equation of the form developed by Horton. Also determine the total rain and the excess rain (runoff).

Time (min)	Rainfall (cm/hr)	f (cm/hr)
1	5.0	3.90
2	5.0	3.40
3	5.0	3.10
4	5.0	2.70
5	5.0	2.50
6	7.5	2.30
8	7.5	2.00
10	7.5	1.80
12	7.5	1.54
14	7.5	1.43
16	2.5	136
18	2.5	131
20	2.5	1.28
22	2.5	1.25
24	2.5	1.23
26	2.5	1.22
28	2.5	1.20
30	2.5	1.20

<u>Solution:</u>

The precipitation and infiltration rates versus time are plotted as shown in Figure.



From figure above, shaded area

$$F_c = 4.25\ sq\ units = 8.25\left(\frac{1cm}{60min} \times 2min\right) = 8.25 \times \frac{1}{30}$$

$$= 0.275\ cm$$

$$k = \frac{(4.5-1.2)cm/hr}{0.275\ cm} = 12\ hr^{-1}$$

The Horton's equation is:

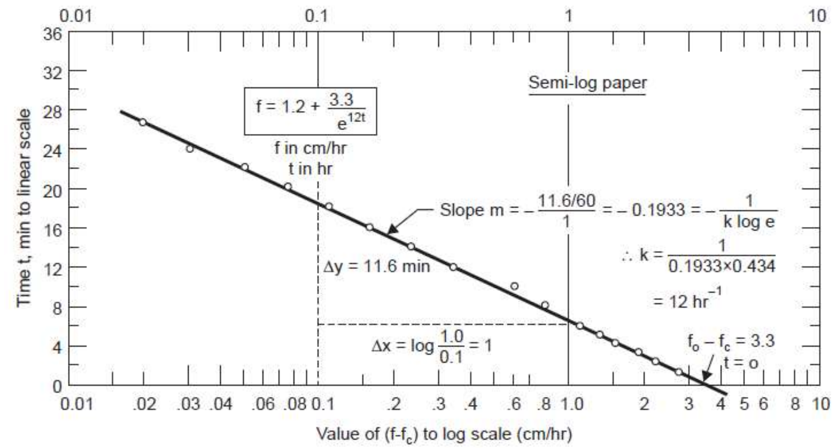
$$f = f_c + (f_o - f_c)e^{-kt} = 1.2 + (4.5-1.2)e^{-12t} = 1.2 + \frac{3.3}{e^{-12t}}$$

The total infiltration loss F_p can also be determined by integrating the Horton's equation for the duration of the storm.

$$F_P = \int_0^t f dt = \int_0^{30/60}\left(1.2 + \frac{3.3}{e^{-12t}}\right)dt = 1.2t + \frac{3.3}{-12e^{-12t}}\Big]_0^{30/60} = 0.88\ cm$$

To determine the Horton's constant by drawing a semi-log plot of t vs. $(f - f_c)$:

The Horton's equation is



Semi-log plot for infiltration constants

From the graph, when $t=0$,

$f - f_c = 3.3 = f_o - f_c$, (Since $f = f_0$ $when$ $t = 0$)

$\therefore f_o = 3.3+1.2 = 4.5$ cm/hr

Hence the Horton's equation is of the form

$$f = 1.2 + (4.5-1.2)e^{-12t} = 1.2 + \frac{3.3}{e^{-12t}}$$

Total rain $P = 5 \times \frac{5}{60} + 7.5 \times \frac{10}{60} + 2.5 \times \frac{15}{60} = 2.29\ cm$

Infiltration loss $F_P = 0.88$ cm

$\therefore P_{net} = excess\ rain = Runoff = P - F_P = 2.29 - 0.88 = 1.41cm$

Example:

In a double ring infiltrometer test, a constant depth of 100 mm was restored at every time interval the level dropped as given below:

Time (min): 0 5 10 15 25 45 60 75 90 110 130

Depth of water (mm): 100 83 87 90 85 78 85 85 85 80 80

Establish the infiltration equation of the form developed by Horton.

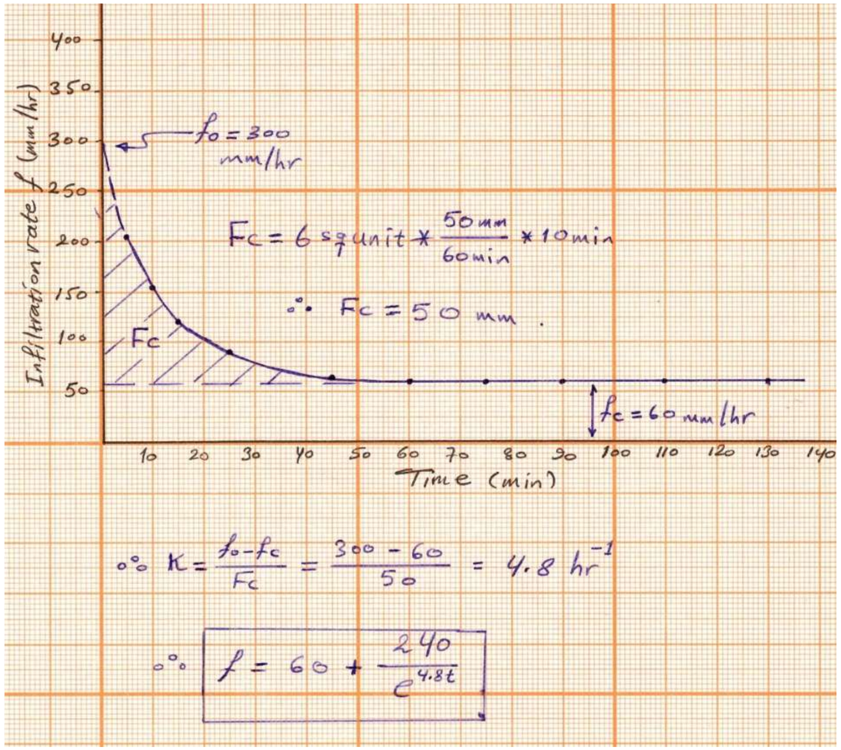
Solution:

Time (min)	Depth to water Surface (mm) Before filling	After filling	Depth of infiltration (mm)	Infiltration rate $f = \frac{d}{\Delta t} \times 60$ (mm/hr)	$f - f_c$ (mm/hr) $f_c = 60$ (mm/hr)
0	100	--	0	--	--
5	83	100	17	204	144
10	87	100	13	156	96
15	90	100	10	120	60
25	85	100	15	90	30
45	78	100	22	66	6
60	85	100	15	60	0
75	85	100	15	60	0
90	85	100	15	60	0
110	80	100	20	60	0
130	80	100	20	60	0

Now plot on millimetric paper, t vs f, figure below.

Solution:

The precipitation and infiltration rates versus time are plotted as shown in Figure.



INFILTRATION INDICES

The infiltration curve expresses the rate of infiltration (cm/hr) as a function of time. The area between the rainfall graph and the infiltration curve represents the rainfall excess, while the area under the infiltration curve gives the loss of rainfall due to infiltration. The rate of loss is greatest in the early part of the storm, but it may be rather uniform particularly with wet soil conditions from rainfall.

Estimates of runoff volume from large areas are sometimes made by the use of infiltration indices, which assume a constant average infiltration rate during a storm, although in actual practice the infiltration will be varying with time. This is also due to different states of wetness of the soil after the start of the rainfall. There are three types of infiltration indices:

(i) ϕ-index (ii) *W*-index (iii) f_{ave}-index.

(i) ϕ-index : The ϕ-index is defined as that rate of rainfall above which the rainfall volume equals the runoff volume. The ϕ -index is relatively simple and all losses due to infiltration, interception and depression storage (storage in pits and ponds) are accounted for; hence,

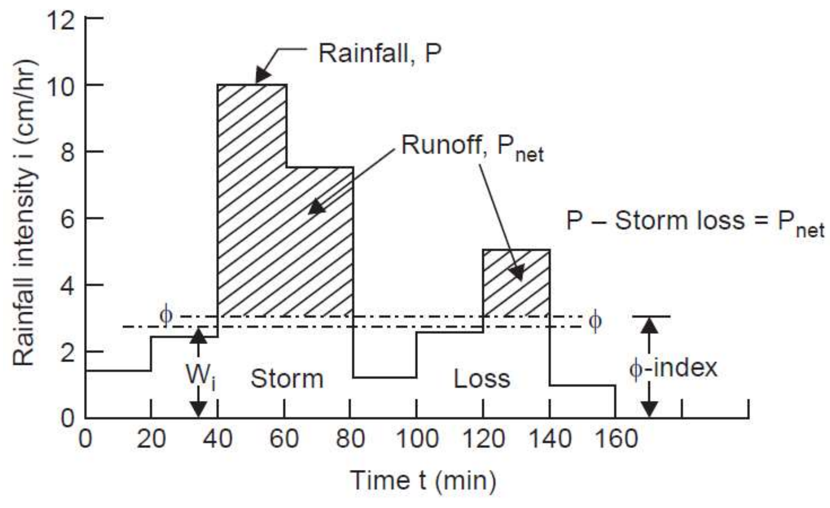
$$\emptyset = \frac{F_P}{t_R} = \frac{P-R}{t_R}$$

Provided $i > \phi$ throughout the storm. The bar graph showing the time distribution of rainfall, storm loss and rainfall excess (net rain or storm runoff) is called a hyetograph, Figure below. Thus, the ϕ-index divides the rainfall into net rain and storm loss.

(ii) W-index: The W-index is the average infiltration rate during the time rainfall intensity exceeds the infiltration capacity rate, i.e.,

$$W = \frac{P-R-S}{t_R} = \emptyset - \frac{S}{t_R}$$

Where P = total rainfall
R = surface runoff
S = effective surface retention
t_R = duration of storm during which $i > fp$
F_p = total infiltration



Infiltration loss by ϕ-index

The W-index attempts to allow for depression storage, short rainless periods during a storm and eliminates all rain periods during which $i < f$. Thus, the W-index is essentially equal to the ϕ-index minus the average rate of retention by interception and depression storage, i.e., $W < \phi$.

Information on infiltration can be used to estimate the runoff coefficient C in computing the surface runoff as a percentage of rainfall i.e.,

$$R = C \times P$$

$$C = \frac{i - W}{i}$$

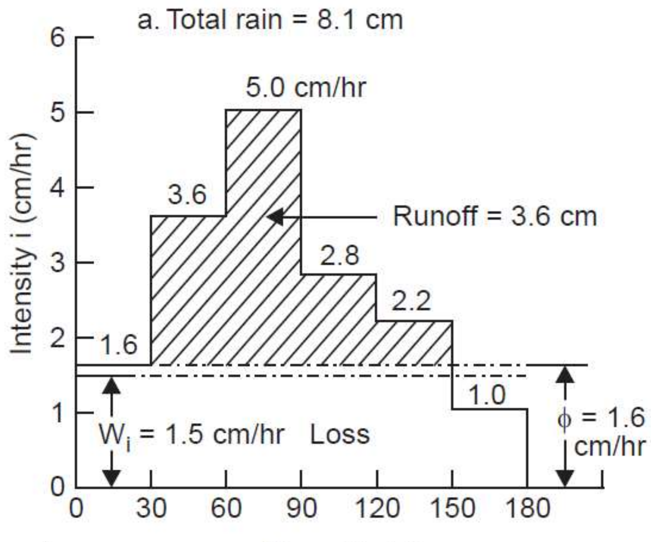
(iii) *fave*-index: In this method, an average infiltration loss is assumed throughout the storm, for the period $i > f$.

<u>Example</u>

The rates of rainfall for the successive 30 min period of a 3-hour storm are: 1.6, 3.6, 5.0, 2.8, 2.2, 1.0 cm/hr.
The corresponding surface runoff is estimated to be 3.6 cm.
Establish the ϕ-index. Also determine the W-index.

<u>Solution</u>

Construct the hyetograph as shown in Figure A.



A

$\Sigma(i - \phi)t = P_{net}$, and thus it follows

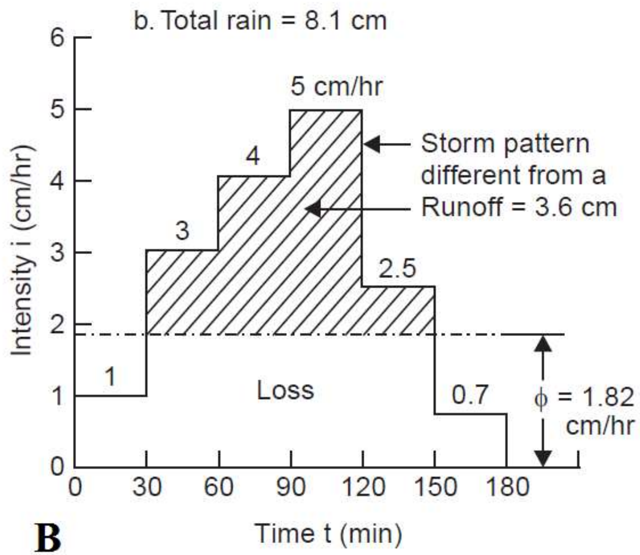
$$[(3.6 - \emptyset) + (5.0 - \emptyset) + (2.8 - \emptyset) + (2.2 - \emptyset)] \times \frac{30}{60} = 3.6$$

$$\Rightarrow \emptyset = 1.6\ cm/hr$$

$$P = (1.6 + 3.6 + 5.0 + 2.8 + 2.2 + 1.0)\frac{30}{60} = 8.1 \text{ cm}$$

$$\Rightarrow W = \frac{P-R}{t_R} = \frac{8.1-3.6}{3} = 1.5\ cm/hr$$

Suppose the same 3-hour storm had a different pattern as shown in Figure B producing the same total rainfall of 8.1 cm. To obtain the same runoff of 3.6 cm (shaded area), the ϕ -index can be worked out as 1.82 cm/hr. Hence, it may be seen that a single determination of ϕ -index is of limited value and many such determinations have to be made and averaged, before the index is used. The determination of ϕ -index for a catchment is a trial and error procedure.



B