Processes in hydrologic cycle

Evaporation

Water evaporates from the oceans and the land surface due to the **heat energy** provided by the **solar radiation** to become part of the atmosphere.

Evapotranspiration

Evaporation from the land surface is accompanied by transpiration by plants. Transpiration is the evaporation of water from aerial parts and of plants, especially leaves but also stems, flowers and fruits. Transpiration is a side effect of the plant needing to open its **stomata** in order to obtain carbon dioxide gas from the air for photosynthesis.

Precipitation

Water vapor is transported and lifted in the atmosphere until it condenses and precipitates on the land or the oceans as **rain, snow, hail, sleet** etc. Some precipitation falls as snow and can accumulate as ice caps and glaciers, which can store frozen water for thousands of years.

Interception by vegetation and depression storage

A part of the precipitated water may be **intercepted by vegetation** or temporarily retained in the soil in **surface depressions (depression storage)** near where it falls and is ultimately returned to the atmosphere by evaporation and transpiration by plants.

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Snowmelt

Snow packs in warmer climates often melt when spring arrives, and the melted water flows overland as snowmelt.

Estimated world water quantities

Table 2.1: Estimated world water quantities

Item	Area (10^6 km^2)	Volume (km^3)	Percent of total water	Percent of fresh water
Oceans	361.3	1,338,000,000	96.5	
Groundwater				
Fresh	134.8	10,530,000	0.76	30.1
Saline	134.8	12,870,000	0.93	
Soil Moisture	82.0	16,500	0.0012	0.05
Polar ice	16.0	24,023,500	1.7	68.6
Other ice and snow	0.3	340,600	0.025	1.0
Lakes				
Fresh	1.2	91,000	0.007	0.26
Saline	0.8	85,400	0.006	
Marshes	2.7	11,470	0.0008	0.03
Rivers	148.8	2,120	0.0002	0.006
Biological water	510.0	1,120	0.0001	0.003
Atmospheric water	510.0	12,900	0.001	0.04
Total water	510.0	1,385,984,610	100	
Fresh water	148.8	35,029,210	2.5	100

Table from World Water Balance and Water Resources of the Earth, Copyright, UNESCO, 1978.

Ocean (saline) 96.5%
Other (saline) 1%
Fresh water 2.5%
Total 100%

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Residence time

The residence time T_r is the average duration for a water molecule to pass through a subsystem of the hydrologic cycle. It is calculated by dividing the volume of water S in storage by the flow rate Q (i.e. $T_r = S/Q$).

The volume of **atmospheric moisture** (Table 2.1) is 12,900 km^3. The flow rate of moisture from the atmosphere as precipitation (Table 2.2) is 458,000 + 119,000 = 577,000 km^3/yr (or the flow rate of moisture to the atmosphere as evaporation is 505,000 + 72,000 = 577,000 km^3/yr), so the average residence time for moisture in the atmosphere is T_r = 12,900/577,000 = 0.0224yr = 8.2 days. The very short residence time for moisture in the atmosphere is one reason why weather cannot be forecast accurately more than a few days ahead.

Similarly, the volume of **water in the rivers** (Table 2.1) is 2,120 km^3. The average flow rate of water in global rivers (Table 2.2) is 44,700 km^3/yr, so the residence time for global rivers is T_r = 2,120/44,700 = 0.0474 yr = 17.3 days. The global residence time for **groundwater** is T_r = (10,530,000+12,870,000)/2,200 = 10,636.36 yrs. The very long residence time for groundwater is the reason why it takes long time to clean groundwater if it is contaminated.

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Lecture Note – 3

Precipitation (Background)

Water Vapor

Atmospheric water mostly exists as a gas, or vapor, but briefly and locally it becomes a liquid in rainfall and in water droplets in clouds, or it becomes a solid in snowfall, in hail, or in ice crystals in clouds.

Specific humidity (q_v)

Specific humidity is defined as the mass of water vapor per unit mass of moist air. As mass per unit volume equals density, therefore specific humidity is also defined as the ratio of density of water vapor to density of moist air.

Mathematically, $q_v = \frac{\rho_v}{\rho_a}$

$q_v = 0.622\frac{e}{p}$

Where,
ρ_v = Density of water vapor
ρ_a = Density of moist air

q_v (unit) = Kg of water / Kg of moist air
(Kg/Kg)

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Vapor pressure

Vapor pressure is the partial pressure of water vapor in the atmosphere.

Mathematically, $e = \rho_v R_v T$ $PV = nRT$

Where,

e = Vapor pressure of water vapor (Pascal) → $N m^{-2}$

ρ_v = Density of water vapor (kg/m^3)

R_v = Gas constant for water vapor (J/kg-K)

T = Absolute temperature (K)

$P_d = \rho_d R_d T$ (dry air)

$P = \rho_a R_a T$ (Total)

$P = P_d + e$

Saturation vapor pressure (e_s)

For a given air temperature, there is a maximum moisture content the air can hold; the corresponding vapor pressure is called the saturation vapor pressure.

capacity এর corresponding V.P.

Mathematically, $e_s = 611 \exp\left(\frac{17.27T}{237.3 + T}\right)$ [Raudkivi, 1979]

Where,

e_s = Saturation vapor pressure (Pascal) → N/m^2

T = Given air temperature (°C)

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Relative humidity (R_h)

It is the ratio of actual vapor pressure to its saturation value at a given air temperature.

Mathematically, $R_h = \frac{e}{e_s}$ (actual) / (maximum)

Dew-point temperature (T_d) = Wet-bulb temperature

The temperature at which air would just become saturated at a given specific humidity (q_v = Constant) is its dew-point temperature.

Q. Specific ও Relative humidity এর মধ্যে Relative টা কেন বেশি use করা হয়?

(A)

1) sp. humidity এর mass measure করা difficult
2) Relative humidity সহজে বুঝা যায়। (easy to underс

- e_s কমা মানে R_h বেড়ে যাবে, যদিও moisture content same. (e)
- Td এর নিচে সবসময় Tp তে Rh 100%, যদিও moisture content lower.

Q.

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and the specific humidity is given by (3.2.6) with $p = 100$ kPa $= 100 \times 10^3$ Pa

$$q_v = 0.622\frac{e}{p}$$

$$= 0.622\left(\frac{1819}{100 \times 10^3}\right)$$

$$= 0.0113 \text{ kg water/kg moist air}$$

The air density is calculated from the ideal gas law (3.2.7). The gas constant R_a is given by (3.2.8) with $q_v = 0.0113$ kg/kg as $R_a = 287(1 + 0.608q_v) = 287(1 + 0.608 \times 0.0113) = 289$ J/kg·K, and $T = 20°C = (20 + 273)$ K $= 293$ K, so that

$$\rho_a = \frac{p}{R_a T}$$

$$= \frac{100 \times 10^3}{289 \times 293}$$

$$= 1.18 \text{ kg/m}^3$$

Precipitation

Precipitation denotes all forms of water that reach the earth from the atmosphere.

Formation of precipitation

- Lifting of air mass in the atmosphere so that it cools and some of its moisture condenses. There are three main mechanisms of air mass lifting:

  A. Frontal lifting: Lifting of warm air on one side of a frontal surface over colder, denser air on the other side.

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B. Orographic lifting: Air mass rises to pass over a mountain range.

C. Convective lifting: Rising of warmer, lighter air in colder, denser surroundings.

for snow → freezing nuclei

- Condensation requires a seed called condensation ← for rainfall nuclei on which the droplets form. Condensation nuclei (0.1 to 10 μm) usually consist of products of combustion, oxides of nitrogen and salt particles.

$= 10^{-6}$ m

- The tiny droplets grow by condensation and impact with their neighbors as they are carried by turbulent air motion, until they become large enough so that the force of gravity overcomes that of friction.

- As they begin to fall, further increase in size occurs as they hit other droplets in the fall path.

- However, as the drop falls, water evaporates from its surface and the drop size diminishes, so the drop

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may be reduced to the size of an aerosol and be carried upwards in the cloud through turbulent action.

Aerosol: Remains airborne indefinitely except for precipitation. (size < 3 μm).

V=gt
S= ½gt²
এ hit করে মাটিতে আ

- The cycle of condensation, falling, evaporation, and rising occurs on average about 10 times before the drop reaches a critical size of about 0.1 mm, which is large enough to fall through the bottom of the cloud.
- Up to 1 mm in diameter, the droplets remain spherical in shape, but beyond this size they begin to flatten out on the bottom until they are no longer stable falling through the air and break up into small raindrops and droplets. Normal raindrops falling through the cloud base are 0.1 to 3 mm in diameter.

Cloud seeding

It is a process of artificially nucleating clouds to induce precipitation. Silver iodide is a common nucleating agent and is spread from aircraft in which a silver iodide solution is evaporated with a propane flame to produce particles.

Forms of Precipitation

Rain: Consists of liquid water drops mostly larger than 0.5 mm in diameter. On the basis of intensity, rainfall is classified as,

Light: Trace to 2.5 mm/h

Moderate: >2.5 mm/h to 7.6 mm/h

Heavy: >7.6 mm/h

Snow: Snow consists of ice crystals. The average density is 0.1 gm/cm^3.

Drizzle: (Sometimes called **mist**) consists of tiny liquid water droplets, usually with diameters between 0.1 and 0.5 mm. Drizzle usually falls from low stratus and rarely exceeds 1 mm/hr.

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Lecture Note – 6

Infiltration

The vertical movement of water through the soil surface is known as infiltration.

Importance of infiltration

1. It affects the timing, distribution, and magnitude of surface runoff.
2. It is the primary step in the natural groundwater recharge.

Infiltration capacity (f_c)

The maximum rate at which the ground can absorb water at a given time as its infiltration capacity.

If f is the actual rate of infiltration then,

$f = f_c$ when $i > f_c$, and
$f = i$ when $i < f_c$

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, and finally settles down to a constant value.

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Infiltration capacity equation

$$f_{ct} = f_{cf} + \left(f_{co} - f_{cf}\right)e^{-K_h t} \; ; \; 0 \leq t \leq t_d \rightarrow$$ Horton's equation (1930)

Where,

f_{ct} = Infiltration capacity at any time t from the start of rainfall

f_{co} = Infiltration capacity at t = 0

f_{cf} = Final steady state value

t_d = Duration of rainfall

K_h = Constant depending on soil characteristics and vegetative cover

Typically,

For a bare sandy soil, f_c = 1.2 cm/hr , with vegetative cove, f_c = 12 cm/hr

For a bare clay soil, f_c = 0.15 cm/hr f_c = 1.5 cm/hr

A good grass cover or vegetation increases this value by as much as 10 times.

Infiltration index

In hydrological calculations, it is convenient to use a constant value of infiltration rate for the duration of the storm. The average infiltration rate is called infiltration index. **

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φ-index

It 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 resulting runoff volume. In this case, the initial loss is also considered as infiltration.

$$\varphi - index = \frac{P - R}{t_e}$$

Where,
P = Total storm precipitation
R = Total storm runoff
t_e = Duration of rainfall excess, i.e. the total time in which the rainfall intensity is greater than φ-index

Now,

if i < φ-index, then infiltration rate (f) = i (No runoff case)
if i > φ-index, then infiltration rate (f) = φ-index, and the difference between rainfall and infiltration in an interval of time represents the runoff volume in that time.

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rainfall excess
(effective rainfall)

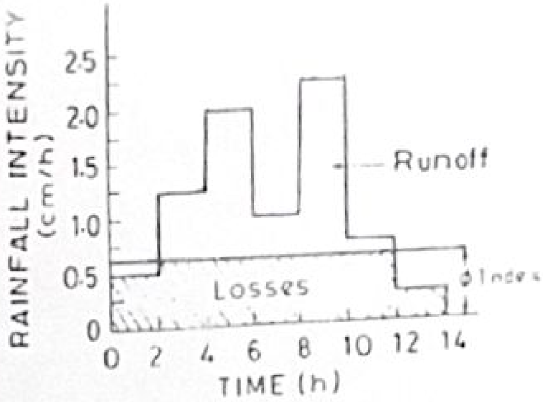


Fig. 3.13 φ Index

এই math টা 30min ও 15 min interval এর জন্য solve করবা।

EXAMPLE 3.5: *A storm with* *10.0 cm* *precipitation produced a direct runoff of 5.8 cm. Given the time distribution of the storm as below, estimate the φ index of the storm.*

exam এ hour এ না দিয়ে min এ দিতে পারে।

		0.5	1	1.5	2	2.5	3	3.5	4
Time from start (h)	0	1	2	3	4	5	6	7	8
Incremental rainfall in each hour (cm)	0	0.4	0.9	1.5	2.3	1.8	1.6	1.0	0.5

← 30 min interval এ time

not cumilitive
cumilitive হলে বিয়োগ করে নিব

সব add করলে 10c.m পাব
exam এ 10c.m নাও দিতে পারে।

Total infiltration $= 10.0 - 5.8 = 4.2$ cm

Assume t_e = time of rainfall excess $= 8$ h for the first trial.

8 hour এই runoff আছে (ধরে নিয়েছি) তাই i should be > φ

Then

For 30min,
$\phi = \frac{4.2 \text{ cm}}{4 \text{ hour}} = 1.05$ cm/hr $= 0.525$ cm/30min

$$\phi = \frac{4.2}{8} = 0.525 \text{ cm/h}$$

এটাকে compare করে Table rainfall থেকে index করব। এর value গুলোর আগে

But this value of φ makes the rainfalls of the first hour and eighth hour ineffective as their magnitude is less than 0.525 cm/h. The value of t_e is therefore modified.

(runoff হয়নি)
এই ২টা hour invalid হওয়াতে 8 hour থেকে 2 hour বাদ দিয়ে $t_e = 6$ hr. ধরেছি।

Assume $t_e = 6$ h for the second trial

In this period,

r = i   r = i

Infiltration $= (10.0 - 0.4 - 0.5 - 5.8)$

$= 3.3$ cm

$$\phi = \frac{3.3}{6} = 0.55 \text{ cm/h}$$

r = rainfal
i = infiltration

- 30 min এ অল্পই rainfall, তাই Infiltration rate বেড়ে যায়
- Finally, $\phi = 0.55 \times 2$ (for 30min) cm/hr
  $\phi = 0.55 \times 4$ (for 15min) cm/hr

for 30min, $\phi = 0.55$ cm/30min
$= 0.55 \times 2$ cm/1 hr

Sabbin sir (Hydrograph)
@ WRE-451
CE (4-I) (A-B-C) ১৬/

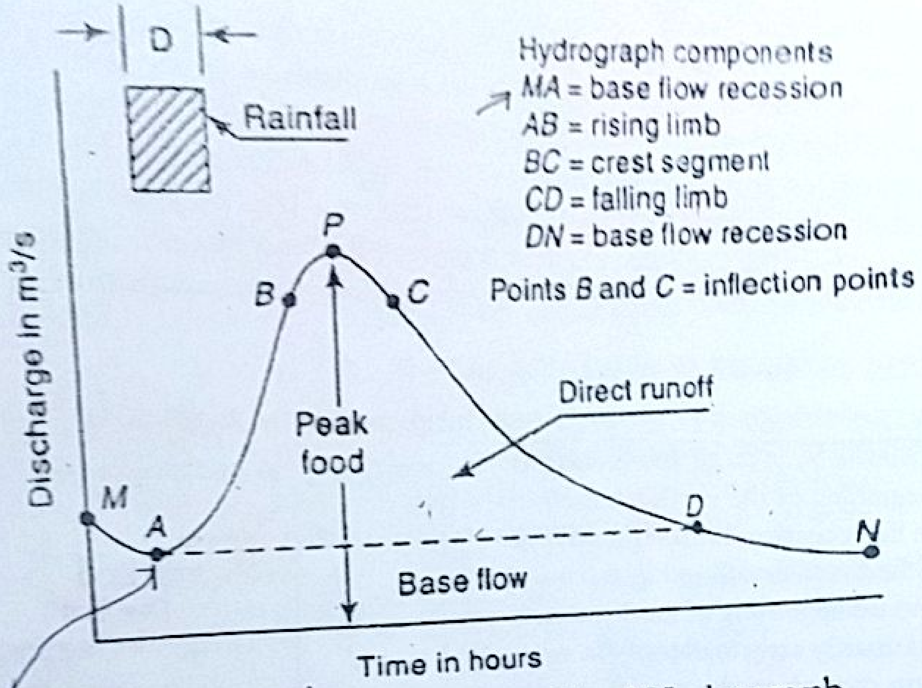


Fig. 6.1 Elements of a Flood Hydrograph

Q. কমছে কেন discharge?

যখন বৃষ্টি হয় না, তখন GWT নিচে নেমে যায়, ফলে base flow কমে যায়
So, discharge Q কমে যায়।

- বৃষ্টি হলে
- BC তে ২টা point of inflection আছে, curvature change.
- আগে higher rate এ Q বাড়ে, পরে lower rate এ Q বাড়ে।
- একই ভাবে, higher rate এ আগে

6.4 BASE FLOW SEPARATION

In many hydrograph analyses a relationship between the surface-flow hydrograph and the effective rainfall (i.e. rainfall minus losses) is sought to be established. The surface-flow hydrograph is obtained from the total storm hydrograph by separating the quick-response flow from the slow'response runoff. It is usual to consider the interflow as a part of the surface flow in view of its quick response. Thus only the base flow is to be deducted from the total storm hydrograph to obtain the surface flow hydrograph.

There are three methods of base-flow separation that are in common use.

METHODS OF BASE-FLOW SEPARATION

METHOD I—STRAIGHT-LINE METHOD In this method the separation of the base flow is achieved by joining with a straight line the beginning of the surface runoff to a point on the recession limb representing the end of the direct runoff. In Fig. 6.5 point A represents the beginning of the direct runoff and it is usually easy to identify in view of the sharp change in the runoff rate at that point.

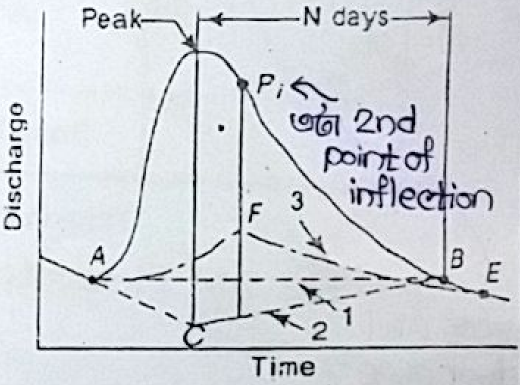


Fig. 6.5 Base Flow Seperation Methods

Point B, marking the end of the direct runoff is rather difficult to locate exactly. An empirical equation for the time interval N (days) from the peak to the point B is

$$N = 0.83A^{0.2} \tag{6.4}$$

where A = drainage area in km^2 and N is in days. Points A and B are joined by a straight line to demarcate to the base flow and surface runoff. It should be realised that the value of N obtained as above is only approximate and the position of B should be decided by considering a number of hydrographs for the catchment. This method of base-flow separation is the simplest of all the three methods.

METHOD 2 In this method the base flow curve existing prior to the commencement of the surface runoff is extended till it intersects the ordinate drawn at the peak (point C in Fig. 6.5). This point is joined to point B by a straight line. Segment AC and CB demarcate the base flow and surface runoff. This is probably the most widely used base-flow separation procedure.

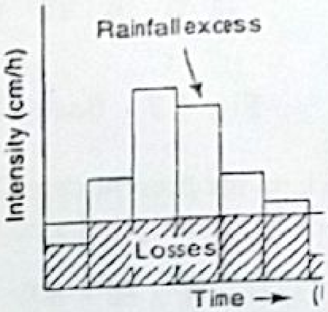
METHOD 3 In this method the base flow recession curve after the depletion of flood water is extended backwards till it intersects the ordinate at the point of inf tion (line *EF* in Fig. 6.5). Points *A* and *F* are joined by an arbitrary smooth curve. method of base-flow separation is realistic in situations where the groundwater co butions are significant and reach the stream quickly.

It is seen that all the three methods of base-flow separation are rather arbitrary. selection of anyone of them depends upon the local practice and successful pre tions achieved in the past. The surface runoff hydrograph obtained after the base- separation is also known as *direct runoff hydrograph* (DRH).

6.5 EFFECTIVE RAINFALL (ER) = Direct runoff

Effective rainfall (also known as *Excess rainfall*) (ER) is that part of the rainfall becomes direct runoff at the outlet of the watershed. It is thus the total rainfall given duration from which abstractions such as infiltration and initial losses are tracted. As such, ER could be defined as that rainfall that is neither retained o land surface nor infiltrated into the soil.

For purposes of correlating DRH with the rainfall which produced the flow, the hyetograph of the rainfall is also pruned by deducting the losses. Figure 6.6 shows the hyetograph of a storm. The initial loss and infiltration losses are subtracted from it. The resulting hyetograph is known as *effective rainfall hyetograph* (ERH). It is also known as *excess rainfall hyetograph*.



ϕ index line
এই line এর উপরের rainfall হল effective rainfall

Fig. 6.6 Effective Rai Hyetograph (E

Both DRH and ERH represent the same total quantity but in different units. Since ERH is usually in cm/h plotted against time, the area of ERH multiplied by the catchment area gives the total volume of direct runoff which same as the area of DRH. The initial loss and infiltration losses are estimated ba the available data of the catchment.

1st 4 hour 2nd 4 hour

EXAMPLE 6.2 *Rainfall of magnitude 3.8 cm and 2.8 cm occurring on two cons 4-h durations on a catchment of area 27 km² produced the following hydrograph at the outlet of the catchment. Estimate the rainfall excess and ϕ index.*

Total Hydrograph বা storm Hydrograph.
48 এর পর st. line এ constant হয়ে গেছে.
আগে A ও E point identify করব।

		← rainfall start										
Time from start of rainfall (h)	−6	0	6	12	18	24	30	36	42	48	54	60
Observed flow (m^3/s)	6	5	13	26	21	16	12	9	7	5	5	4.5

SOLUTION: The hydrograph is plotted to scale (Fig. 6.7). It is seen that the hydrograph has a base-flow component. For using the simple straight-line method flow separation, by eq. (6.4)

$$N = 0.83 \times (27)^{0.2} = 1.6 \text{ days} = 38.5 \text{ h}$$

However, by inspection, DRH starts at $t = 0$, has the peak at $t = 12$ h and ends at (which gives a value of $N = 48 - 12 = 36$ h). As $N = 36$ h appears to be more sati

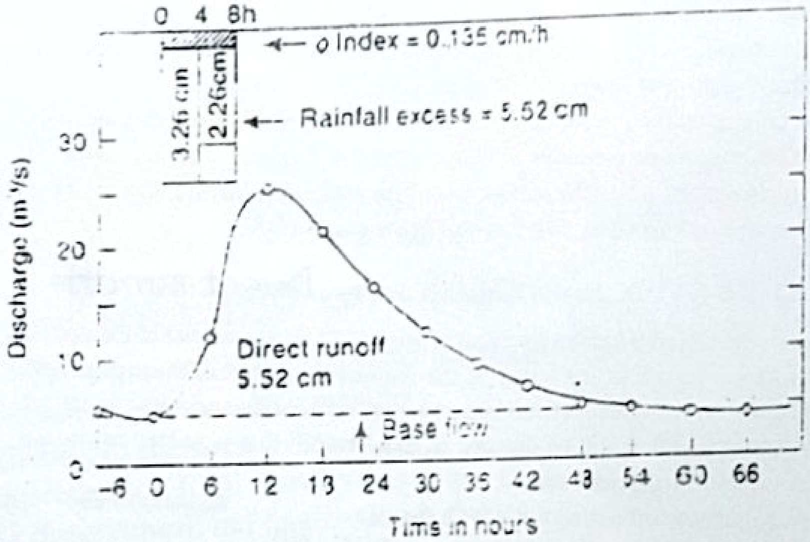


Fig. 6.7 Base Flow Separation—Example 6.2

than N = 38.5 h. in the present case DRH is assumed to exist from t = 0 to 48 h. A straight line base flow separation gives a constant value of 5 m³/s for the base flow.

$$\text{Area of DRH} = (6 \times 60 \times 60)\left[\frac{1}{2}(8) + \frac{1}{2}(8+21) + \frac{1}{2}(21+16) + \frac{1}{2}(16+11) + \frac{1}{2}(11+7) + \frac{1}{2}(7+4) + \frac{1}{2}(4+2) + \frac{1}{2}(2)\right]$$

$$= 3600 \times 6 \times (8 + 21 + 16 + 11 + 7 + 4 + 2) = 1.4904 \times 10^6 \text{ m}^3$$

= Total direct runoff due to storm

$$\text{Runoff depth} = \frac{\text{runoff volume}}{\text{catchment area}} = \frac{1.4904 \times 10^6}{27 \times 10^6} = 0.0552 \text{ m}$$

= 5.52 cm = rainfall excess

Total rainfall = 3.8 + 2.8 = 6.6 cm

Duration = 8 h

$$\phi \text{ index} = \frac{(6.6 - 5.52)}{8} = 0.135 \text{ cm/h} \quad = \frac{\text{Total loss}}{\text{Total time}}$$

rainfall এর timeline use করবে

$$\left(100\% \times \frac{5.52}{6.6}\right) \leftarrow \text{Runoff co-efficient}$$

4

EXAMPLE 6.4 *Given below are the ordinates of a 6-h unit hydrograph for a catchment. Calculate the ordinates of the DRH due to a rainfall excess of 3.5 cm occurring in 6 h.*

Time (h)	0	3	6	9	12	15	18	24	30	36	42	48	54	60	69
UH ordinate (m^3/s)	0	25	50	85	125	160	185	160	110	60	36	25	16	8	0

SOLUTION: The desired ordinates of the DRH are obtained by multiplying the ordinates of the unit hydrograph by a factor of 3.5 as in Table 6.3. The resulting DRH as also the unit hydrograph are shown in Fig. 6.10 (a). Note that the time base of DRH is not changed and remains the same as that of the unit hydrograph. The intervals of coordinates of the unit hydrograph (shown in column 1) are not in any way related to the duration of the rainfall excess and can be any convenient value.

Table 6.3 Calculation of DRH Due to 3.5 ER – Example 6.4

Time (h)	Ordinate of 6-h unit hydrograph (m^3/s)	Ordinate of 3.5 cm DRH (m^3/s)
1	2	3
0	0	0
3	25	87.5 = 25x3.5
6	50	175.0
9	85	297.5
12	125	437.5
15	160	560.0
18	185	647.5
24	160	560.0
30	110	385.0
36	60	210.0
42	36	126.0 = 36x3.5
48	25	87.5
54	16	56.0
60	8	28.0
69	0	0

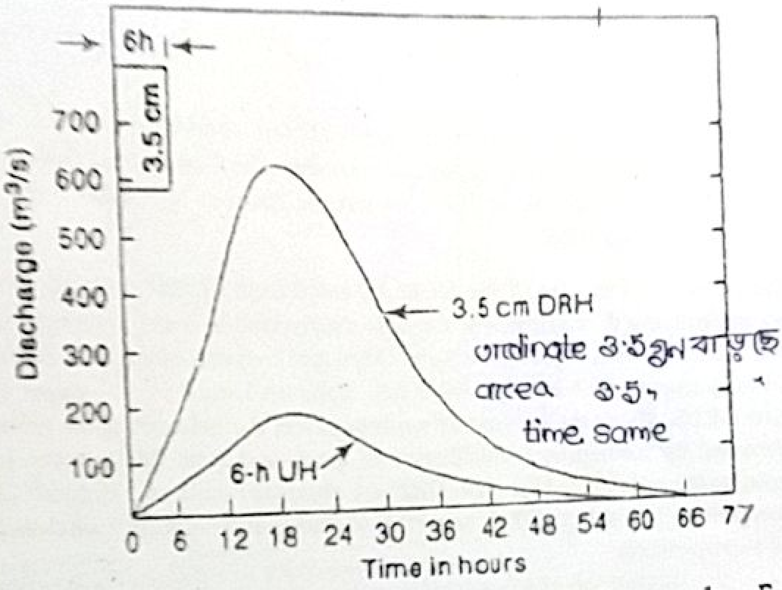


Fig. 6.10(a) 3.5 cm DRH derived from 6-h Unit Hydrograph—Example 6.4

... of the catchment area. This indicates that larger durations are admissible for larger catchments. By experience it is found that the duration of the unit hydrograph should not exceed 1/5 to 1/3 basin lag. For catchments of sizes larger than 250 km² the duration of 6 h is generally satisfactory.

EXAMPLE 6.6 *Following are the ordinates of a storm hydrograph of a river draining a catchment area of 423 km² due to a 6-h isolated storm. Derive the ordinates of a 6-h unit hydrograph for the catchment.*

Time from start of storm (h)	–6	0	6	12	18	24	30	36	42	48
Discharge (m^3/s)	10	10	30	87.5	111.5	102.5	85.0	71.0	59.0	47.5
Time from start of storm (h)	54	60	66	72	78	84	90	96	102	
Discharge (m^3/s)	39.0	31.5	26.0	21.5	17.5	15.0	12.5	12.0	12.0	

* যদি 90 তে Q = 10 হত, এবং পরের 96 ও 102 hr এও Q = 10 হত, তবে, পুরোটা baseflow 10, কারণও 10, so constant base flow 10 হত। তখন SH থেকে 10 বাদ দিলে DRH পাওয়া যেত।

The storm hydrograph is plotted to scale (Fig. 6.12). Denoting the time from beginning of storm as *t*, by inspection of Fig. 6.12,

A = beginning of DRH $t = 0$

B = end of DRH $t = 90$ h

P_m = peak $t = 20$ h

Hence

$$N = (90 - 20) = 70 \text{ h} = 2.91 \text{ days}$$

By Eq. (6.4),

$$N = 0.83\,(423)^{0.2} = 2.78 \text{ days}$$

However, $N = 2.91$ days is adopted for convenience. A straight line joining A and B is taken as the divide line for base-flow separation. The ordinates of DRH are obtained by subtracting the base flow from the ordinates of the storm hydrograph. The calculations are shown in Table 6.5.

* এই মানে ৩, 0 তে 10 m³/s, 90 পর্যন্ত বেড়ে 12.5 m³/s হয়েছে। 90 এর পর const. হয়ে গেছে।

so প্রত্যেক hr এর base flow value interpolate করে বের করব। যেমন 6 hr এ → $= 10 + \frac{2.5}{90} \times 6$

• base flow value কে whole no. round off করলে calculation easy হয়।

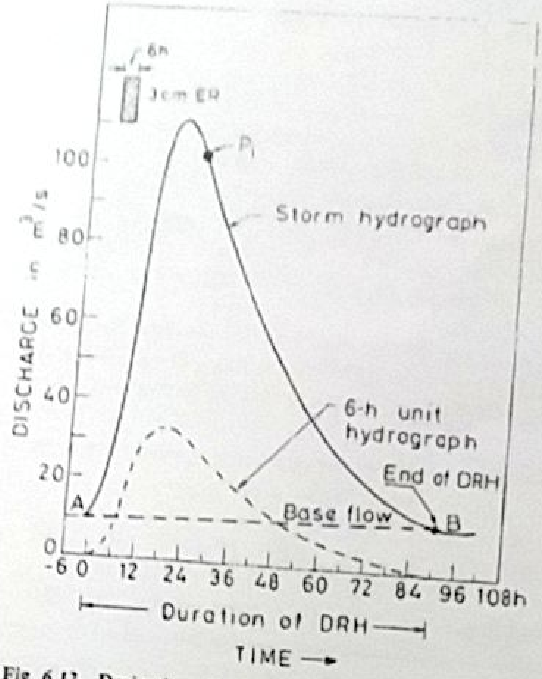


Fig. 6.12 Derivation of unit hydrograph from a storm hydrograph

Volume of DRH $= 60 \times 60 \times 6 \times$ (sum of DRH ordinates)

$= 60 \times 60 \times 6 \times 587 =$ **12.68 Mm³** Million or Mega=M

Drainage area $= 423\ km^2 = 423\ Mm^2$

Runoff depth = ER depth $= \frac{12.68}{423} = 0.03\ m = 3\ cm.$

The ordinates of DRH (col. 4) are divided by 3 to obtain the ordinates of the 6-h unit hydrograph, see Table 6.5.

✱ চাকরী পরীক্ষায় এবার আসে

EXAMPLE 6.7 *The peak of a flood hydrograph due to a 3-h effective storm is 270 m³/s. The mean depth of rainfall is 5.9 cm. Assuming an average infiltration loss of 0.3 cm/h and a constant base flow of 20 m³/s, estimate the peak of the 3-h unit hydrograph.*

Duration of rainfall excess = 3 h

Total depth of rainfall = 5.9 cm

Loss @ 0.3 cm/h for 3 h = 0.9 cm

Rainfall excess = 5.9-0.9 = 5.0 cm

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TABLE 6.5 CALCULATION OF THE ORDINATES OF A 6-h UNIT HYDROGRAPH—EXAMPLE 6.6

Time from beginning of storm (h)	Ordinate of storm hydrograph (m^3/s)	Base flow (m^3/s)	Ordinate of DRH (m^3/s)	Ordinate of 6-h unit hydrograph (Col. 4 ÷ 3)
1	2	3	4 = ২-৩	5
−6	10.0	10.0	0	0
0	10.0	10.0	0	0
6	30.0	10.0	20.0	6.7
12	87.5	10.5	77.0	25.7
18	111.5	10.5	101.0	33.7
24	102.5	10.5	92.0	30.7
30	85.0	11.0	74.0	24.7
36	71.0	11.0	60.0	20.0
42	59.0	11.0	48.0	16.0
48	47.5	11.5	36.0	12.0
54	39.0	11.5	27.5	9.2
60	31.5	11.5	20.0	6.6
66	26.0	12.0	14.0	4.6
72	21.5	12.0	9.5	3.2
78	17.5	12.0	5.5	1.8
84	15.0	12.5	2.5	0.8
90	12.5	12.5	0	0
96	12.0	12.0	0	0
102	12.0	12.0	0	0
			Sum = 587.0	195.7

এখানে base flow 0.5 এ roundoff করা আছে।

= Σ(ordinate) × এর time laps
এখানে time lap হল 6hr
= (6 × 3600) sec

Example 6.7 (*Contd.*)

Peak flow:

Peak of flood hydrograph = 270 m^3/s

Base flow = 20 m^3/s

Peak of DRH = 250 m^3/s

$$\text{Peak of 3-h unit hydrograph} = \frac{\text{peak of DRH}}{\text{rainfall excess}} = \frac{250}{5.0} = 50 \text{ m}^3/\text{s}$$

Unit Hydrograph from a Complex Storm

When suitable simple isolated storms are not available, data from complex storms of long duration will have to be used in unit-hydrograph derivation. The problem is to decompose a measured composite flood hydrograph into its component DRHs and base flow. A common unit hydrograph of appropriate duration is assumed to exist. This problem is thus the