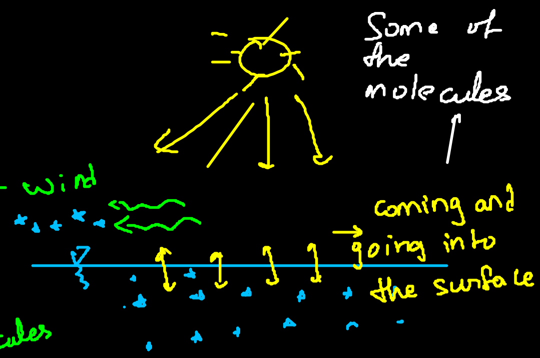
Evaporation

- Water is below boiling point
- Heat is supplied to the system.



Carries the vapor molecules away from the water body.

★ If the kinetic energy of the molecules goes above a threshold level, the water molecules escape the surface. → **Evaporation**: Net escape of water molecules from the liquid state to gaseous.

⊛ Latent Heat :-

Is the amount of heat absorbed by a unit mass of water without change in temperature while passing from liquid to gaseous state (Latent Heat of Vaporisation)

$$\boxed{l_v = 2.501 \times 10^6 - 2370T \quad T = {}^\circ C}$$

$\boxed{l_v = J/kg}$

⊛ Factors affecting Evaporation

- Amount of Incoming solar Radiation
- Temperature
- Wind conditions.
- Vapour Pressure at the water surface.
- Water Quality (Sea water Evap is less than freshwater)
- Altitude
- Seasonal.
- Size → Surface Area.
- Depth ⇒ More evap in shallow lake.

* **Transpiration**:- Is a process of loss of water through plant leaves in which water is extracted from the plants roots, transported upwards, diffused into atmosphere from pores.

* Evaporation + Transpiration = **Evapotranspiration** (ET)

PET ⇒ Maximum amount of ET that can occur when water is freely available at a region.

AET ⇒ Actual ET at any given time. | AET < PET

$$\boxed{\text{Aridity Index} = \frac{AET}{PET}}$$

AI = 0 at PWP
= 1 at FC.

Soil moisture

Field capacity = max. soil moisture soil can hold

PWP = Permanent wilting point, below which plants can no longer extract water.

* Methods of Evaporation Estimation
  - Experimented
  - Emperical
  - Analytical methods.

* Experimental:- Lake Evaporation = C_p × Pan Evaporation

C_p → Pan coefficient ≈ 0.7 to 0.8

* Avg. water consumption in India: $\boxed{165\ LPCD}$

→ Hirakud dam ≈ $725\ km^2$

Odisha popl^n = 25 millions

Evap ≈ 160 cm/yr. ; Vol. of water lost = $1.60\ \frac{m}{yr} \times 725 \times 10^6\ m^2$

$= 1160\ Mm^3/yr$

⇒ $1160 \times 10^9\ L/yr$;

→ If this available for water supply, $25 \times 10^6 \times 165\ L/d$.

$$\Rightarrow \frac{1160 \times 10^9\ L/yr}{25 \times 10^6 \times 165 \times 365\ L} = 0.77\ yrs.$$

$$\approx \boxed{280\ days}$$

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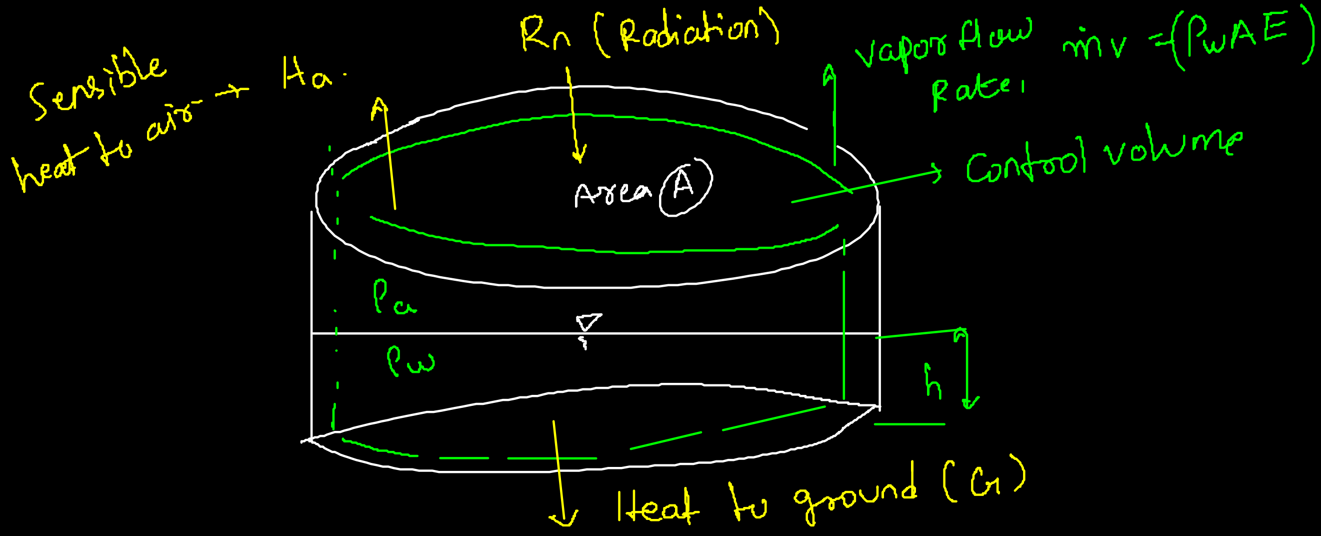
* Analytical methods → Energy Balance

→ Aerodynamic

* Energy Balance Method

→ we apply both Continuity Eq^n and Energy Eq^n to estimate the evaporation.



* $\dot{m}_v$ = Vapour flow Rate = $\rho_w A E$ (E = Evaporation Rate)

↳ $\frac{\partial}{\partial t}$ is (−ve) for liquid, +ve for vapour phase.

and $\boxed{E = -\frac{dh}{dt}}$ (h decreases with time 't')

* Here B = Mass of Water in liquid phase

$\boxed{\beta = \frac{dB}{dm} = 1;}$ $\boxed{\frac{dB}{dt} = (-\dot{m}_v)}$

* Continuity for water phase

$$-\dot{m}_v = \frac{d}{dt}\iiint_{C.V.} \rho_w \, d\forall + \boxed{\iint_{C.S} \rho_w \cdot \vec{V} \cdot d\vec{A}}$$

↳ No flow across water. the control surface = 0

$$\Rightarrow \boxed{\dot{m}_v = -\rho_w A \cdot \frac{dh}{dt}}$$

⊛ Continuity for vapour phase

$$\frac{dB}{dt} = \dot{m}_v \quad ; \quad B = \text{water vapor mass} \Rightarrow \boxed{\beta = q_v}$$

$$\Rightarrow \quad \dot{m}_v = \frac{d}{dt}\iiint_{C.V} q_v \cdot \rho_a \, d\forall + \iint_{C.S} q_v \cdot \rho_a \, \vec{V} \cdot d\vec{A}$$

STEADY FLOW of Vapour

$$\Rightarrow \quad \dot{m}_v = \iint_{C.S} q_v \cdot \rho_a \, \vec{V} \cdot d\vec{A}$$

⇒ From liquid phase continuity

$$\rho_w A \frac{dh}{dt} = \iint_{C.S} q_v \rho_a \vec{V} \cdot d\vec{A}$$

$$\Rightarrow \quad \boxed{E = \frac{1}{\rho_w A} \iint_{C.S} q_v \rho_a \cdot \vec{V} \cdot d\vec{A}}$$

⊛ Now apply the Energy Equation.

B = Total energy of the fluid in the CV.

$$\beta = (dB/dm)$$

$$\Rightarrow \frac{dB}{dt} = \frac{dH}{dt} - \frac{dW}{dt}$$

$\frac{dW}{dt}$ → work done by system or surroundings

$\frac{dH}{dt}$ → Change in Sensible heat

and $B = \frac{dB}{dm} = \left(e_u + \frac{V^2}{2} + gz\right)$

$$\Rightarrow \frac{dH}{dt} - \frac{dW}{dt} = \frac{d}{dt}\iiint_{C.V}\left(e_u + \frac{V^2}{2} + gz\right)\rho_w \, d\forall + \iint_{C.S}\left(e_u + \frac{V^2}{2} + gz\right)\rho_w \vec{V}\cdot d\vec{A}$$

Here $\boxed{\frac{dW}{dt} = 0;}$ $\boxed{V = 0}$ and $\frac{dz}{dt} \approx$ Small.

For water

$$\frac{d}{dt}\left(\frac{V^2}{2}\rho_w \, d\forall\right) \approx 0 \qquad \frac{d}{dt}\left(gz\rho_w \, d\forall\right) \approx 0$$

$$\Rightarrow \frac{dH}{dt} = \frac{d}{dt}\iint_{C.V} e_u \rho_w \, d\forall + \boxed{\iint_{C.S}\left(e_u + \frac{V^2}{2} + gz\right)\rho_w \vec{V}\cdot d\vec{A}}$$

(A) Net outflow of heat energy carried across the control surface

↳ The only exchange is R_n, G and H_a. So, the vapour does not carry any energy to the atmosphere across the C/S

= 0

$$\Rightarrow \boxed{\frac{dH}{dt} = \frac{d}{dt}\iiint_{C.V} e_u \rho_w \, d\forall}$$ $\Rightarrow$ change in internal energy of the water in the pan.

$$\boxed{\frac{dH}{dt} = R_n - H_s - G}$$

* If we assume 'T' within the CV is constant, then the only change in Heat energy stored equals to . . . . . .

(A) change in heat stored in CV

$= $ change in internal energy

$$= \boxed{l_v \cdot \dot{m}_v}$$

$$\Rightarrow \boxed{R_n - H_s - G = l_v \cdot \dot{m}_v}$$

Here $\dot{m}_v = \rho_w A E$ (Assume unit Area)

$$\Rightarrow R_n - H_s - G = l_v \cdot \rho_w E$$

$$\Rightarrow \boxed{E = \frac{R_n - H_s - G}{\rho_w l_v}}$$

(A) Practically H_s and $G << R_n$

$$\Rightarrow \boxed{E = \frac{R_n}{\rho_w l_v}}$$

$R_n = W/m^2 = (J/s/m^2)$

$\rho_w = kg/m^3$

$l_v = J/kg$.

Aerodynamic Method

- Humidity Deficit
- Wind velocity.

→ Before that, we look at some emperical Eq^n s.

① **Meyers**

$$\boxed{E_L = K_m (e_s - e_a)\left(1 + \frac{u_9}{16}\right)}$$

u_9 = Monthly avg. wind @ 9m above GL (Kmph)

K_m = coefficient ≈ 0.36 (large, deep lake)

≈ 0.5 (Small, shallow)

Rohwers Formula.

$$E_L = 0.771\,(1.465 \cdot P_a)\, f(u_0)\,(e_{sat} - e_a)$$

$f(u_0)$ → wind velocity at GL

P_a = Pressure Reading

AERODYNAMIC Method

$f \propto (dq_v/dz)$ → Humidity gradient

$\tau \propto \left(\frac{du}{dz}\right)$ ← wind velocity gradient.

① $\boxed{\dot{m}_v = -\rho_a K_w \left(\frac{dq_v}{dz}\right)}$ — ①

→ Convection

* Mass flux of vapour is proportional to q_v gradient

K_w = Vapour eddy diffusivity

② Momentum flux; $\tau \propto (du/dz)$

$$\Rightarrow \boxed{\tau = \rho_a K_m \frac{du}{dz}}$$

K_m = Momentum Diffusivity. (L^2/T)

$$\Rightarrow \dot{m}_v = -\rho_a K_w \cdot \frac{(q_{v2} - q_{v1})}{(z_2 - z_1)}$$

$$\tau = \rho_a K_m \frac{(u_2 - u_1)}{(z_2 - z_1)}$$

Ⓐ $$\frac{\dot{m}_v}{\tau} = \frac{-K_w}{K_m} \cdot \left(\frac{q_{v2} - q_{v1}}{u_2 - u_1}\right)$$

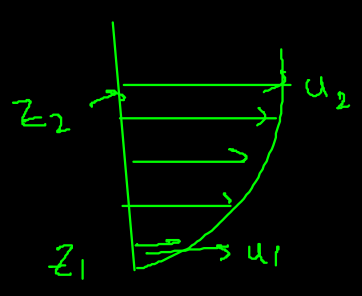
$$\Rightarrow \dot{m}_v = -\tau \cdot \frac{K_w}{K_m} \cdot \left(\frac{q_{v2} - q_{v1}}{u_2 - u_1}\right) \quad \text{— ②}$$

$$\tau = \rho_a \cdot \left(\frac{(u_2 - u_1)k}{\ln(z_2/z_1)}\right)^2$$

Ⓐ For wind

$$\frac{u}{u^*} = \frac{1}{k} \ln\left(\frac{z}{z_0}\right)$$

$$u^* = \sqrt{\tau/\rho_a}$$



z_0 = Roughness height

u_* = shear velocity

$$\Rightarrow \quad \dot{m}_v = -\rho_a \cdot \left(\frac{(u_2 - u_1)k}{\ln(z_2/z_1)} \right)^2 \cdot \frac{k_w}{k_m} \cdot \frac{q_{v2} - q_{v1}}{u_2 - u_1}$$

$$\Rightarrow \quad \dot{m}_v = \frac{\rho_a \cdot (u_2 - u_1) \cdot (q_{v1} - q_{v2})\, k^2}{\left(\ln\left(\frac{z_2}{z_1}\right)\right)^2} \cdot \frac{k_w}{k_m}$$

Von-Karman, $k \simeq 0.4$ and $\frac{k_w}{K_m} \simeq 1$

$$\Rightarrow \quad \boxed{\dot{m}_v = \frac{\rho_a (u_2 - u_1) \cdot (q_{v1} - q_{v2})\, k^2}{\left[\ln(z_2/z_1)\right]^2}}$$

(*) At $z_1 = z_0$ (Roughness height @ water surface).

$u_1 = 0$ and $q_1 = 0.622\, e_a/p$ | $e_a = 611 \exp\left(\frac{17.27T}{237 + T}\right)$

At $z_2 \Rightarrow q_2 = 0.622\, e_a/p$

$*(e_a = e_s \cdot R_h)$

$$\Rightarrow \quad m_v = \frac{\rho_a (u_2 - 0)(e_s - e_a) \cdot \frac{0.622}{p} \times k^2}{\left[\ln(z_2/z_0)\right]^2}$$

$$\Rightarrow \dot{m}_v = \underbrace{\rho_w \left(\frac{0.622 \rho_a k^2 u_2}{P \rho_w \left(\ln\left(\frac{z_2}{z_0}\right)\right)^2} \right)}_{B} (e_s - e_a)$$

$$\Rightarrow \boxed{\dot{m}_v = \rho_w B (e_s - e_a)}$$

→ Unit Area = 1

$$\Rightarrow \rho_w A E = \rho_w B (e_s - e_a) \Rightarrow \boxed{E = B(e_s - e_a)}$$

Combined Energy Balance and Aerodynamic.

$$\boxed{E = \frac{\Delta}{\Delta + \gamma} E_e + \frac{\gamma}{\Delta + \gamma} E_a}$$

$$\Rightarrow \boxed{E_e = \frac{R_n}{l_v \cdot \rho_w}} \quad \text{and} \quad \boxed{E_a = B(e_s - e_a)}$$

Δ = Gradient of saturated vapor pressure curve.

$$\Delta = \frac{4098 e_s}{(237.3 + T)^2} \qquad e_s = 611 \exp\left(\frac{17.27T}{237 + T}\right)$$

γ = Psychometric Constant

(*) We know that H_s = Sensible Heat flux. lost to air by convection.

Similarly $l_v \cdot \dot{m}_v$ = Vapour flux through the air by Convection
Heat

β = Bowen Ration = $\dfrac{\text{Sensible Heat flux}}{\text{Vapour Heat flux}} = \left(\dfrac{H_s}{l_v \dot{m}_v}\right)$

⇒ Now, $G = 0$ (Assume).

$$\Rightarrow R_n - H_s - G = l_v \cdot \dot{m}_v \Rightarrow R_n = \beta \cdot l_v \cdot \dot{m}_v + l_v \dot{m}_v$$

$$\Rightarrow \boxed{R_n = l_v \dot{m}_v \cdot (1 + \beta)}$$

Ⓐ How to get Bowen's Ratio ⇒ $\beta = ?$

$$\beta = \frac{H_s}{l_v \dot{m}_v} = \frac{-\rho_a C_p K_h \cdot dT/dz}{l_v \left(-\rho_a k_w \frac{dq}{dz}\right)}$$

K_h = heat diffusivity

$$= \frac{C_p K_h (T_2 - T_1)}{l_v K_w (q_2 - q_1)}$$

C_p = specific Heat

$$= \frac{C_p K_h \cdot P (T_2 - T_1)}{0.622\, l_v K_w (e_2 - e_1)}$$

$$\boxed{\beta = \frac{\gamma (T_2 - T_1)}{(e_2 - e_1)}}$$

Psychometric constant

$$\boxed{\gamma = \frac{C_p K_h P}{0.622\, l_v K_w}}$$

⊛ Priestley Taylor Method (1972) :- For large lakes.

* The energy balance considerations govern the evaporation.

→ The second term in combined method ≈ 30 % of the first one.

⇒ Combined evaporation = $\left(\alpha \cdot \dfrac{\Delta}{\Delta + \gamma}\right) E_e$

⇒ $\boxed{\alpha = 1.3}$ ⇒ For large water bodies.

Surface Water

⊛ As the rain falls, it gets stored in different components. The water gets released at different times.

$\boxed{P} \longrightarrow \boxed{\text{Storage}} \longrightarrow \boxed{\text{Flow}}$

↳ we try to model this

→ Interception → Gets evaporated and lost to atmosphere.

← Depression

* Water is stored on small depressions on the land.

$\boxed{\text{Overland flow}}$ starts only after the interception and Depression storage is filled.

- Surface storage, Channel storage → Satisfied → $\boxed{\text{stream flow}}$

- Soil Moisture → If there is sufficient supply
  ↳ Interflow

  ↓

- GW Storage → Through Base flow → Rivers

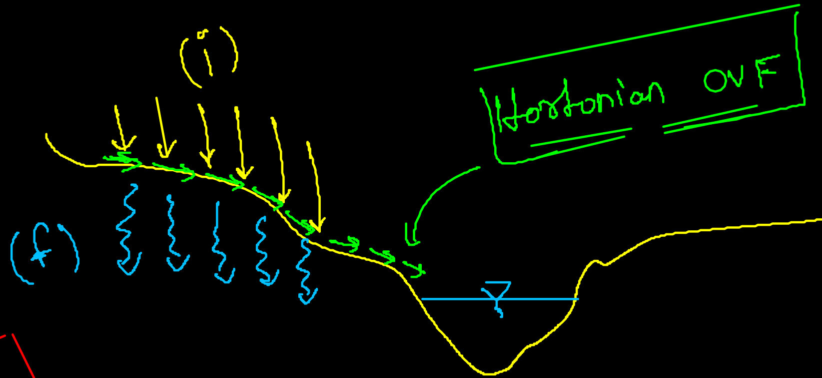
-

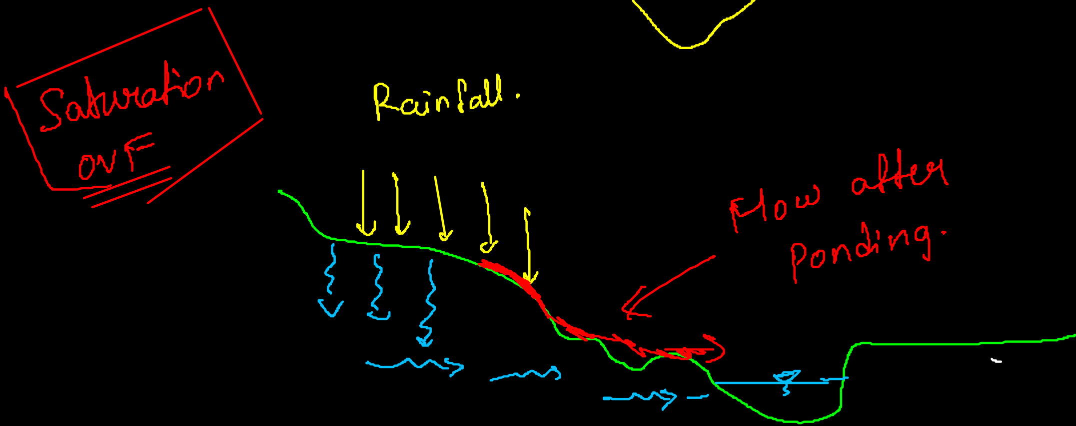
$\boxed{\text{Storages}}$

Retention	Detention.
↳ water is retained for long durat	→ Short Duration
→ Depleted by evaporation	→ Depleted by evaporation / flow out of that component
↳ Ex:- Soil Moisture ;	→ Detention Basins, Dams. etc . .

- Hortonian OVF v/s Saturation OVF

Runoff = Rainfall - Infiltration	Saturated from below
catchment is saturated from top.	Hilly areas → steep catchment slopes.
Impermeable area / low infiltration Semi-Arid / Arid areas	High Permeability.





* In most applications, we use Hortonian flow.

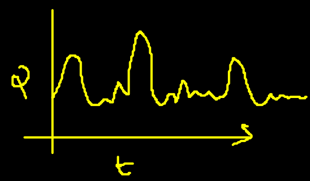
* Variable source area → Applicable for saturation OVF.
  - * At any given point of time, the amount of area that is contributing to runoff at outlet is different.
  - Not all parts of the catchment contribute to runoff at the outlet initially. The fraction of catchment contributing increases with time.

* <u>Streamflow hydrograph</u>:- The plot b/w Q and t. observed/ measured at a particular location.

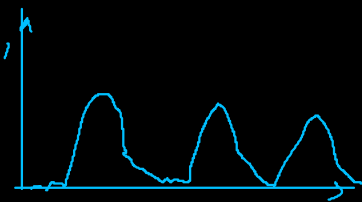
* Represents the response of a catchment to rainfall event.

Types of Hydrograph:-

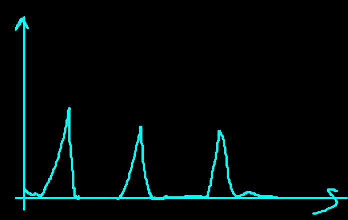
- Annual (Q v/s t for the whole year)
- Storm Hydrograph (Q v/s t, for a single storm).
- Perennial → Always flow river
  - → significant base flow



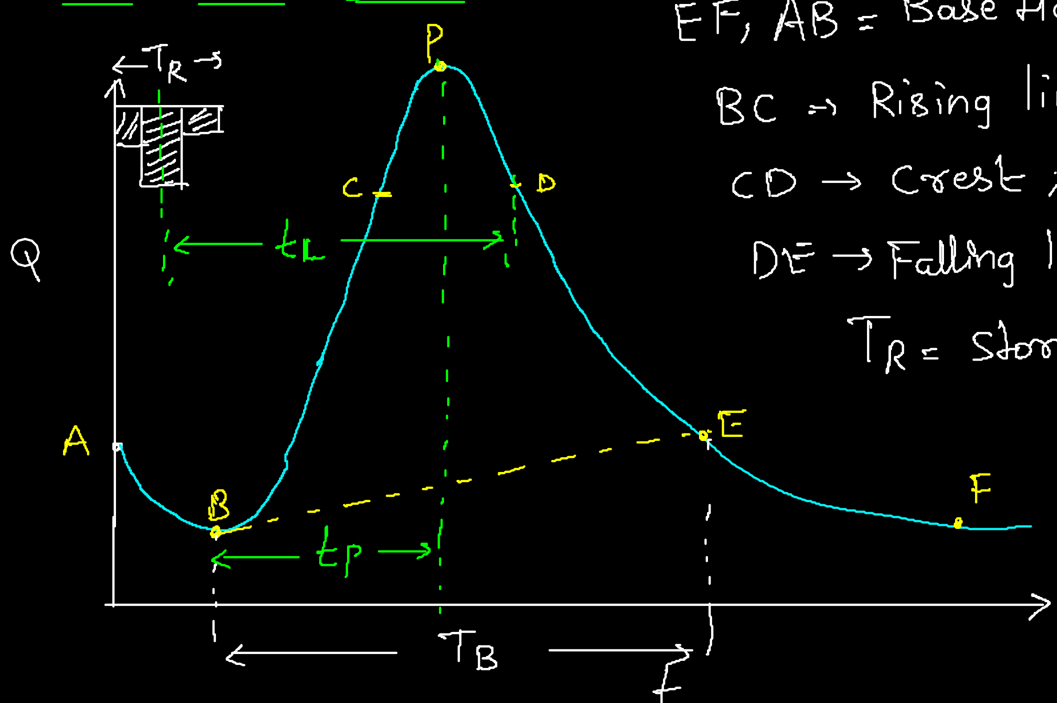
- Intermittent → Flow occurs intermittently,
  - → Partial GW contribution



- Ephemeral
  - → Spikes of flow
  - → No GW contribution.



Storm Hydrograph.



EF, AB = Base flow Recession.

BC → Rising limb.

CD → Crest ; P = Peak.

DE → Falling limb

T_R = Storm Duration

t_p = Time to peak.

T_B = time base

$\boxed{T_L = \text{Time lag (or Basin lag)}}$

⊛ **Rising Limb:-** Also known as the concentration curve.

⊛ Release of water due to gradual increase in the storage of the catchment →

⊛ Initially, it rises slowly b/c of high infiltration.

⊛ Later, there is a steep rise due to low infiltration as the soil gets saturated.

⊛ [The shape of rising limb depends on both catchment character and storm characteristics.]

⊛ **CREST**

⊛ Various portions of catchment contribute to outlet runoff

⊛ Peak will occur after the storm.

⊛ Multiple Peaks occur when there are two successive storms

⊛ **Falling Limb**

→ water contribution from depletion of different storages

→ only a function of Catchment char.

Recession curve (Horton, 1933)

↳ Normal depletion curve,

$$\boxed{Q_t = Q_0 \cdot e^{-(t-t_0)/K}}$$

Q_t = Flow at time 't' on the falling limb

Q_0 = Initial flow at begining of falling limb

t_0 = time at start of falling limb.

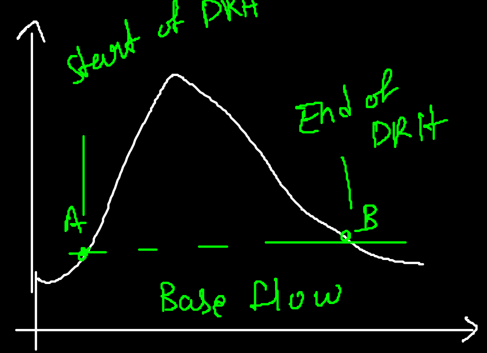
↳ Assumption:- Catchment is a linear reservoir

$\Rightarrow \boxed{Q = KS}$

Baseflow seperation

Str line method

* Applicable to ephemeral



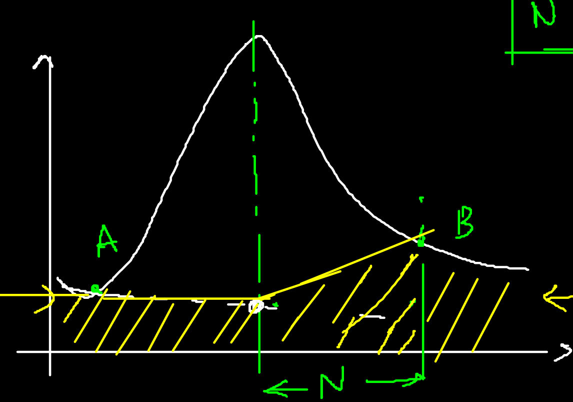
Fixed Base Method.

* For bigger catchments
* Flat catchments

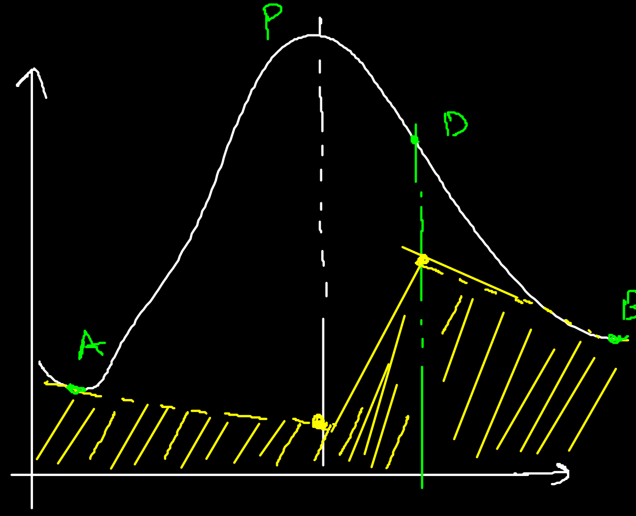
$$\boxed{N = 0.83 A^{0.2}}$$

N in days

A = Catchment area in km^2



③ Variable slope method



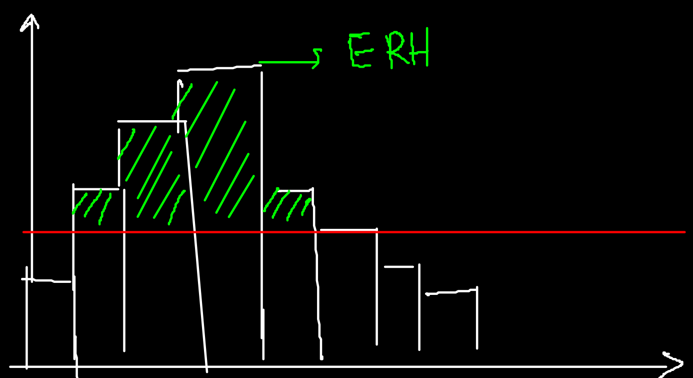
→ Extrapolate from A to forward

→ Extrapolate from B to backward

→ Applicable to moderate size catchments.

← Base flow

* Effective Rainfall Hyetograph (ERH)



ϕ- Index → Constant

rate of infiltration that would yield an ERH with a total depth equal to the depth of direct runoff (R_d)

$$\boxed{R_d = \sum_{m=1}^{M} (R_m - \phi \Delta t)}$$

R_m = observed Rainfall in m^{th} time interval

R_d = Depth of direct runoff

M = Total no. of intervals that actually contribute to runoff.

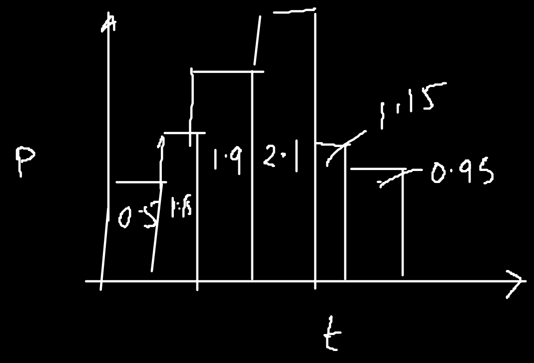
Ex:- ϕ - Index

(*) An isolated storm in a catchment produced a runoff of 3.5 cm.
Mass curve of the avg. rainfall is given.

T	0	1	2	3	4	5	6
cumulative RF	0.	0.50	1.65	3.55	5.65	6.80	7.75

(*) What is the ϕ - index. ?

T	Mass c	Depth
0	0.50	0
1	0.50	0.50
2	1.65	1.15
3	3.55	1.9
4	5.65	2.1
5	6.80	1.15
6	7.75	0.95



RF = 7.75 cm

Runoff = R_d = 3.5 cm

$$\Rightarrow \quad R_d = \sum_{i=1}^{M} (R_m - \phi \Delta t)$$

M = Total no. of intervals actually contributing to direct runoff

$\nleq$ Total no. of intervals in ERH

(1) Consider M = 1

$$\Rightarrow \quad 3.5 = (2.1 - \phi \cdot (1))$$

$$\Rightarrow \quad \cancel{\phi = -1.4} \text{ cm/hr}$$

↳ Not possible.

→ But the blocks above $\phi \neq M$

* Consider $M = 2$

$$\Rightarrow 3.5 = 2.1 - \phi(1) + 1.9 - \phi(1)$$

$$\Rightarrow 3.5 = 4 - 2\phi$$

$$\Rightarrow \phi = 0.25 \text{ cm/hr}$$ → But the no. of blocks above $\phi \neq M$

* Consider $M = 4$

$$3.5 = 2.1 - \phi(1) + 1.9 - \phi(1) + 1.15 - \phi(1) + 1.15 - \phi(1)$$

$$\Rightarrow 3.5 = 6.3 - 4\phi$$

$$\Rightarrow \phi = 0.7 \text{ cm/hr.}$$ ⊗ Blocks above $\phi \neq M$

* $M = 5 \Rightarrow \boxed{\phi = 0.75 \text{ cm/hr}}$ ✓ Blocks above $\phi = M$

---

* ϕ-Index does not account for Initial Abstractions
* W-Index accounts for them.

$$R_d = \sum_{m=1}^{M} (R_m - W\Delta t - I_a)$$

→ Additional term to account for the Initial abstraction

* Runoff coefficient

$$\boxed{C = \frac{R_d}{\sum_{m=1}^{M} R_m}}$$

Algorithm for Abstractions using Infiltration Eq^ns

- Applicable to the ungauged streams. (No flow data available)
- The abstractions due to interception & depression are accounted seperately by other methods.
- The algorithm will be developed using Green Ampt Eq^ns.

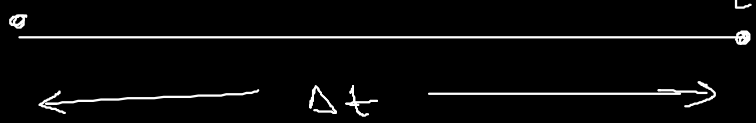
Given :- Rainfall Hyetograph; $\psi, K, \eta, \theta_e, \theta_i\ (S_e)$

Find :- Ponding time; Infiltration after ponding, ERH.

- Ponding time eq^n under constant intensity applicable
- In the absence of ponding, cumulative P = cumulative F
- The potential Infiltration rate ⇒ function of 'F'
- When $f(t) < i(t)$ ⇒ $\boxed{\text{Ponding occured}}$
- The intensity of rainfall $= i(t)$ b/w t and $t+\Delta t$
- The potential infiltration rate at the begining of any interval $= f_t$
- The cumulative infiltration at the beging of interval $= F_t$
- Corresponding quantities at the end of interval are

$\boxed{f_{t+\Delta t}}$ & $\boxed{F_{t+\Delta t}}$ and F_t ⇒ known. for each time interval.

④ i_t, f_t, F_t $\qquad$ $i_{t+\Delta t}, f_{t+\Delta t}$, $F_{t+\Delta t}$



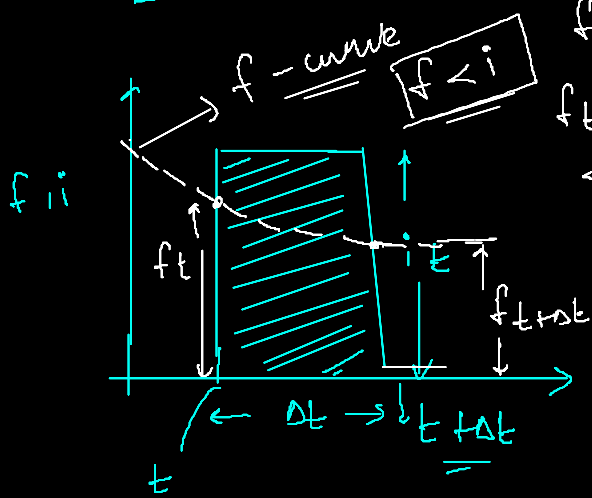
③ possible cases of ponding can happen at each time interval.

Case 1:- Ponding throughout the time interval.

Case 2:- No ponding throughout " " "

Case 3:- Ponding starts during the time interval

Case 1



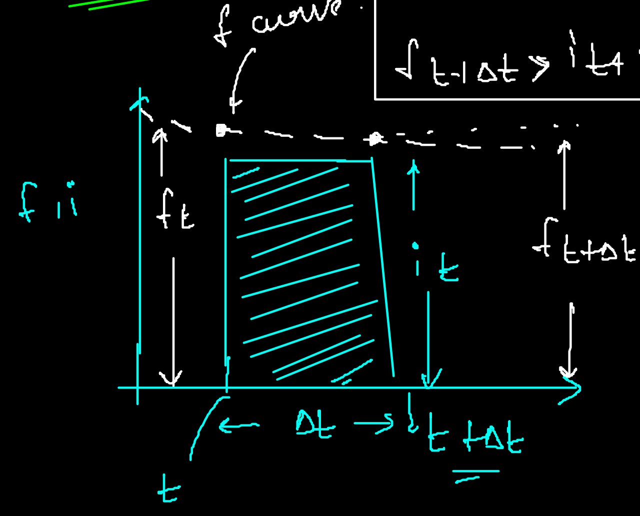
$f_t < i_t$

$f_{t+\Delta t} < i_{t+\Delta t}$

Case 2

$f_t > i_t$

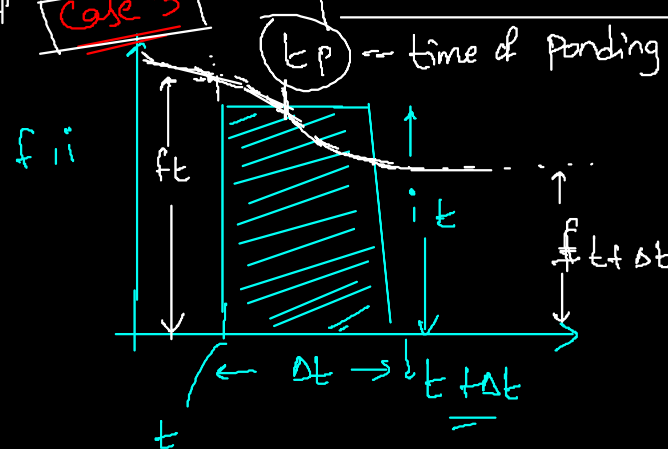
$f_{t+\Delta t} > i_{t+\Delta t}$

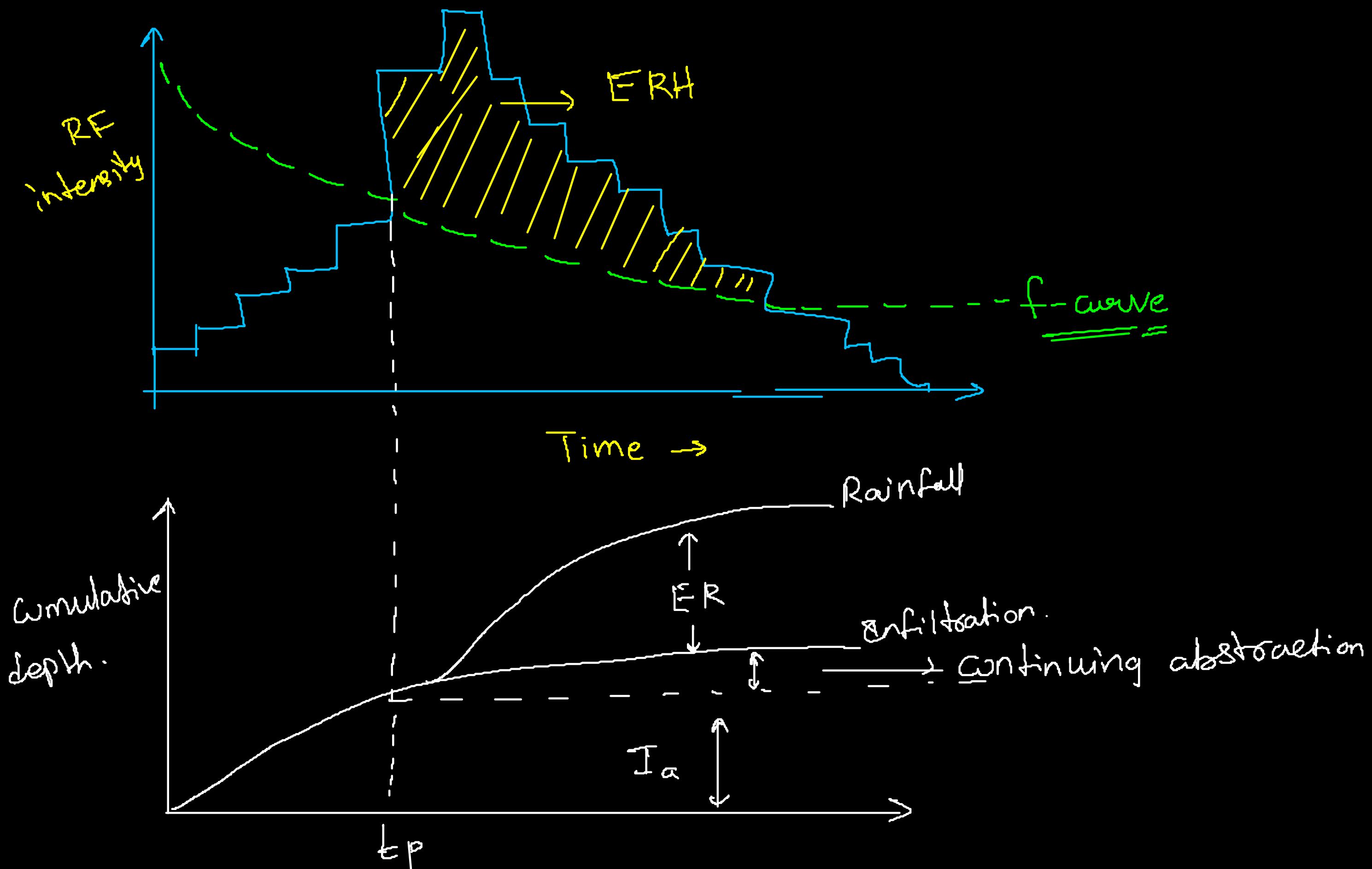


Case 3

$f_t > i_t$

$f_{t+\Delta t} \leq i_{t+\Delta t}$

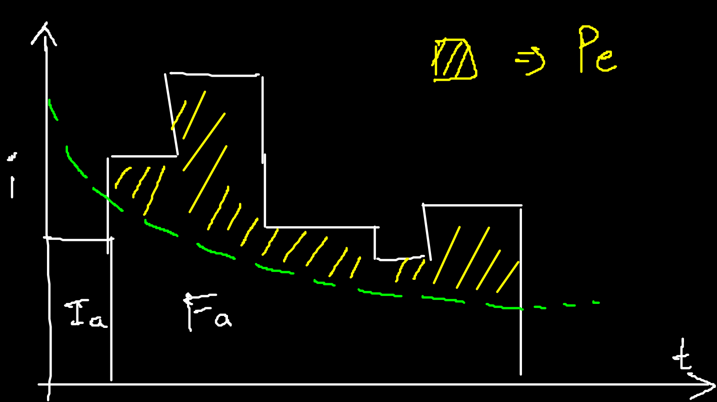




SCS – Curve Number Method

Ⓐ **Basic concept** :- Applicable to only isolated storm events

$$\boxed{\frac{\text{Actual Retention}}{\text{Potential Retention}} = \frac{\text{Actual Runoff}}{\text{Potential Runoff}}}$$



$$\boxed{P = P_e + F_a + I_a}$$

P = Total Rainfall depth

I_a = Initial abstraction

P_e = Excess Rainfall / Direct Runoff.

F_a = Depth of water retained ; $\boxed{S = \text{Max Retention}}$

$$\boxed{\frac{F_a}{S} = \frac{P_e}{P - I_a}}$$ and $\boxed{P = P_e + I_a + F_a}$

$$\Rightarrow \frac{P - P_e - I_a}{S} = \frac{P_e}{P - I_a}$$

$$\Rightarrow \frac{(P - I_a) - P_e}{S} = \frac{P_e}{P - I_a}$$

$$\Rightarrow (P - I_a)^2 - P_e(P - I_a) = P_e S$$

$$\Rightarrow \boxed{\frac{(P - I_a)^2}{(P - I_a + S)} = P_e}$$

I_a = Initial abs.

S = Potential retention.

We assume that $\boxed{I_a = 0.2S}$ ⇒ for small catchments

$$\Rightarrow \boxed{P_e = \frac{(P - 0.2S)^2}{(P + 0.8S)}}$$

⇒ This can be solved graphically using a curve number

⇒ Curve Number depends on the Retention storage (S) in the catchment

$$\Rightarrow \boxed{CN = \frac{1000}{10 + S}}$$

for P is in inches.
S is in inches

CN = No dimensions (0 to 100)

* CN = 100 ⇒ Completely impervious ⇒ $\boxed{S = 0}$

* CN = 0 ⇒ Completely pervious ⇒ $\boxed{S \to \infty}$

* The CN are given for different soil types and LU/LC. for Antecedent Moisture Condition (AMC-II)

→ To find the CN for AMC-I and AMC-III, we have to modify.

$$\boxed{CN(\text{I}) = \frac{4.2\, CN(\text{II})}{10 - 0.058\, CN(\text{II})}}$$

$$\boxed{CN(\text{III}) = \frac{23\, CN(\text{II})}{10 + 0.13\, CN(\text{II})}}$$

Last 5 day RF → (Inches).

		Dormant	Growing Season
AMC-I ⇒	Dry	< 0.5	< 1.4
AMC-II ⇒	Normal	0.5 – 1.1	1.4 – 2.1
AMC-III ⇒	Wet	> 1.1	> 2.1

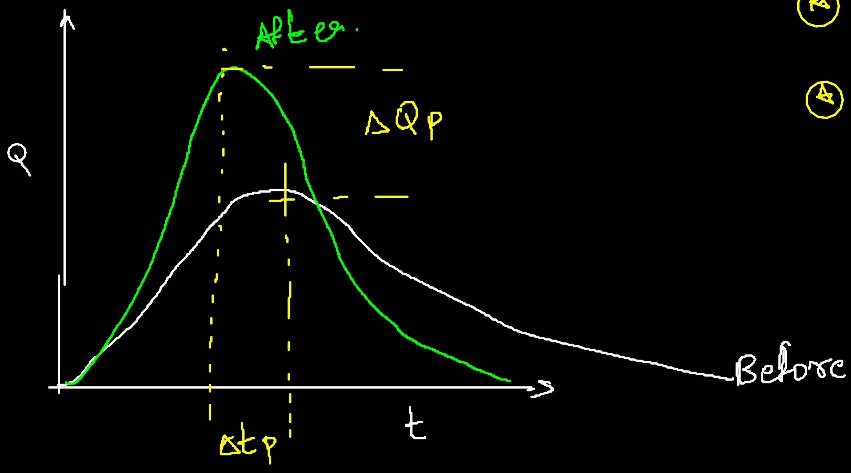
Urbanization Effects.

→ Catchment imperviousness increases.

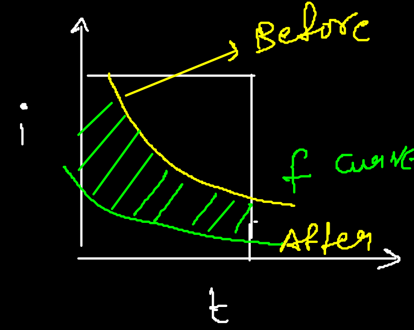
→ Less infiltration, more runoff

⇒ Increased and earlier peak flow.

→ Hydraulic efficiency of the catchment increases.



- Early peak
- Higher Runoff.



Surface flow modelling

- Find flow depth & velocity of the overland flow. (Sheet flow)

→ After RF water begins to pond

→ Initially, this water moves as sheet flow (≈ 100 to 200 ft)

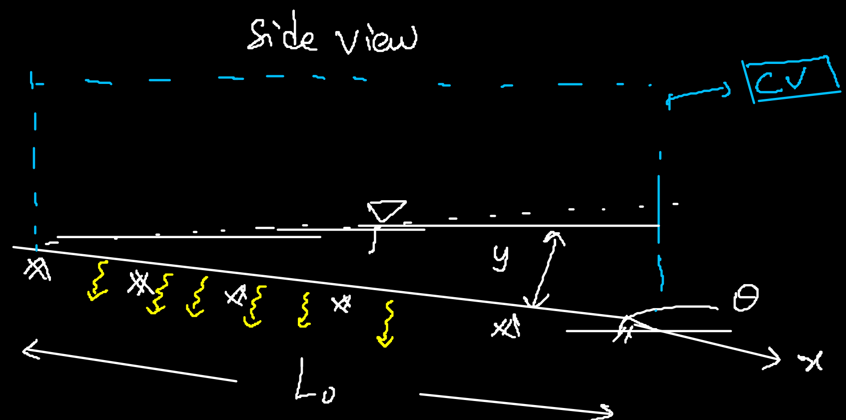
→ After that, small channels start to appear.

→ These small channels combine into recognisable channels.

- overland flow

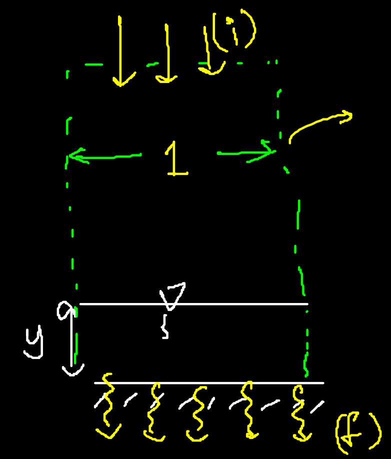
θ = Angle b/w the ground and the horizontal.

y = ⊥ar depth to the ground



⊛ L_0 = Length of ground below flow, V = outlet flow velocity.

Front view



⊛ Assume steady state

$$\frac{dB}{\cancel{dt}} = \cancel{\frac{d}{dt}\iiint_{C.V}} \beta\rho\, d\forall + \iint_{C.S} \beta\rho\, \vec{V}\cdot d\vec{A}$$

↳ zero

B = Mass

$\beta = 1$;

ρ = constant

$$\Rightarrow \quad 0 = \iint_{C.S} \vec{V}\cdot d\vec{A}$$

outflows

$$\Rightarrow \quad f L_0 \cos\theta - i L \cos\theta + V y = 0$$

Inflow

$$\Rightarrow \quad V y = \boxed{q_0 = (i - f) L_0 \cos\theta}$$

⊛ Now the momentum Eq^n; for uniform, laminar:

$$\Rightarrow \quad V = \frac{g S_0 y^2}{3\nu}$$

For uniform, $S_0 = S_f = (h_f/L)$

$\Rightarrow$ Darcy Weisbach $\rightarrow$ $$\boxed{h_f = \frac{fL}{4R}\cdot\frac{V^2}{2g}}$$

f = friction factor

$96/R_e$

$\Rightarrow$ For a unit width of sheet flow, $R = \dfrac{\text{Area}}{\text{Wetted Perim}} = \underline{\underline{y}}$

$\boxed{\text{Valid for } R_e \le 2000}$ $\rightarrow$ Laminar.

⊛ Now, $$\boxed{y = \frac{fV^2}{8gS_o}}$$

But $q_o = Vy \Rightarrow V^2 = \dfrac{q_o^2}{y^2}$

$$\Rightarrow \boxed{y = \left(\frac{f q_o^2}{8 g S_o}\right)^{1/3}}$$

$\Leftarrow$ Depth of sheet flow, Laminar case

⊛ But what if the flow is turbulent?

↳ General expression (both Laminar & Turbulent)

$\rightarrow$ Use Manning's Equation

$$V = \frac{1.49}{n} R^{2/3} S_f^{1/2}$$

(FPS system)

Put $R = y$, $S_f = S_o$ and $q_o = Vy$

$$\Rightarrow \boxed{y = \left(\frac{n q_0}{1.49\, S_0^{1/2}}\right)^{3/5}} \Leftarrow \text{Valid for Turbulent flow}$$

⊛ General case :-

$$\boxed{y = \alpha q_0^{m}}$$

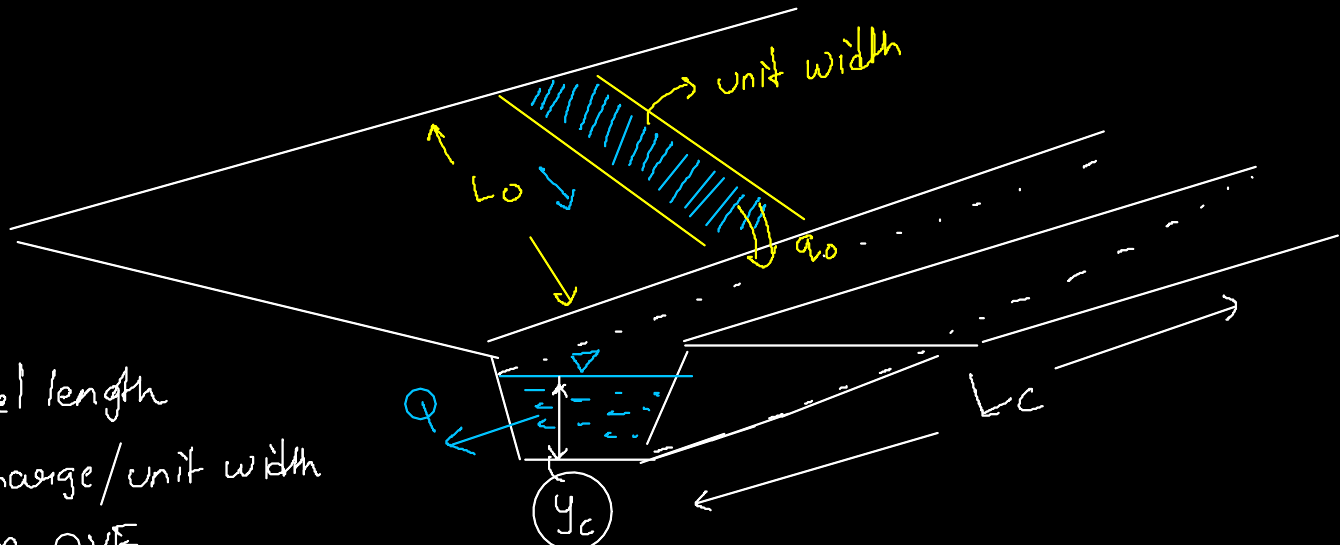
⊛ Laminar flow :- $\alpha = \left(\frac{f}{8 g S_0}\right)^{1/3}$ and $m = 2/3$

⊛ Turbulent flow ; $\alpha = \left(\frac{n}{1.49\, S_0^{1/2}}\right)^{3/5}$; $m = 3/5$

↳ FPS

For SI ; $\alpha = \left(\frac{n}{S_0^{1/2}}\right)^{3/5}$

Channel flow



L_c = channel length

q_0 = Discharge/unit width from OVF

Q = channel discharge

⊛ We want to find Q, y_c at different points along L_c.

$$Q = \frac{1.49}{n} R^{2/3} S_0^{1/2} \cdot V \qquad \Big|\ S_f \approx S_0$$

- The discharge in the channel due to overland flow contribution

$$\boxed{Q = q_0 L_c}$$

- Use Newton-Raphson Method to solve for Mannings depth.

↳ Let the error $= f(y_j) = \underbrace{Q_j - Q}_{\text{minimize this}}$ [Q_j = Guess after jth iteration]

⇒ (1) Select an arbitrary value of y_j (Initial guess)

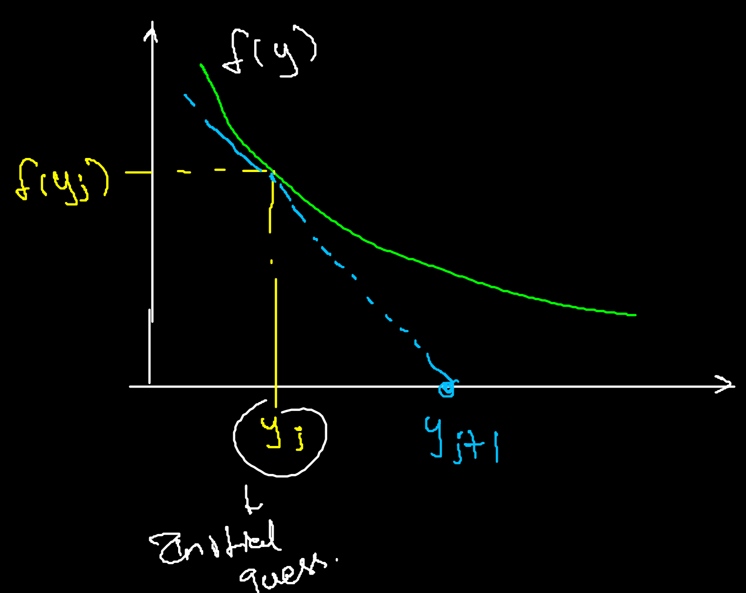
(2) Then calculate Q_j using Mannings Eqn

(3) Compute and estimate error function.

(4) Check for convergence.

- In NR method

$$\boxed{y_{j+1} = y_j - \frac{f(y_j)}{(df/dy)_j}}$$



- It extrapolates the tangent to get the next value

$$f(y_j) = Q_j - Q$$

$$= \frac{1.49}{n} R^{2/3} S_o^{1/2} - Q$$ → constant

$$\Rightarrow \left(\frac{df}{dy}\right)_j = \frac{1.49}{n} S_o^{1/2} \left(\frac{2}{3} \cdots \right)$$

$$\Rightarrow \boxed{\left(\frac{df}{dy}\right)_j} = Q_j \boxed{\left(\frac{2}{3R} \frac{dR}{dy} + \frac{1}{A} \frac{dA}{dy} \right)_j}$$ → channel slope function

$$\Rightarrow \boxed{y_{j+1} = y_j - \frac{1 - Q/Q_j}{\left[\left(\frac{2}{3R} \frac{dR}{dy} + \frac{1}{A} \frac{dA}{dy} \right)_j\right]}}$$ → channel slope function.

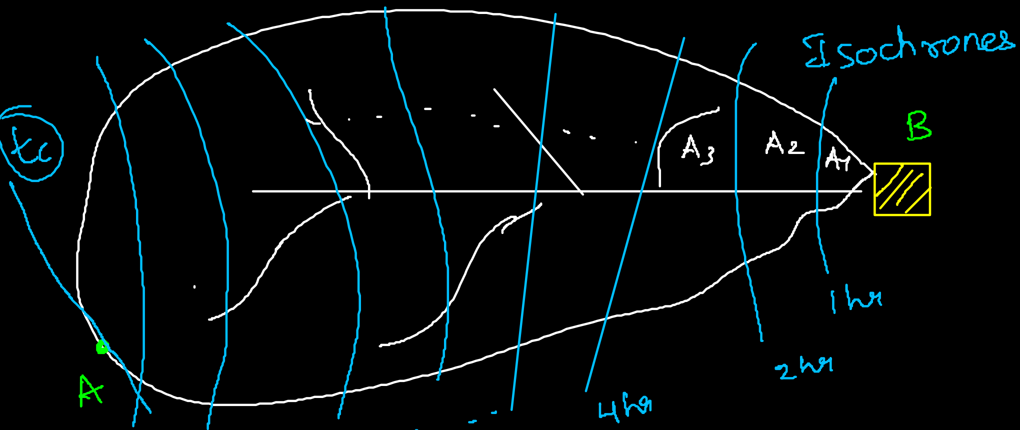
⇒

(A) **Travel time**:- The travel time of flow from one point on a watershed to another can be deduced from

$$\boxed{t = \sum_{i=1}^{I} \frac{\Delta l_i}{V_i}}$$

V_i = Incremental velocity in i^{th} interval.

(A) **Time of concentration**:- Time of travel for water to travel from the remotest location to the catchment outlet.



If A is the remotest location, time of travel b/w A and B = $\boxed{t_c}$

→ Isochrone is defined as the line joining equal times of travel in a catchment.

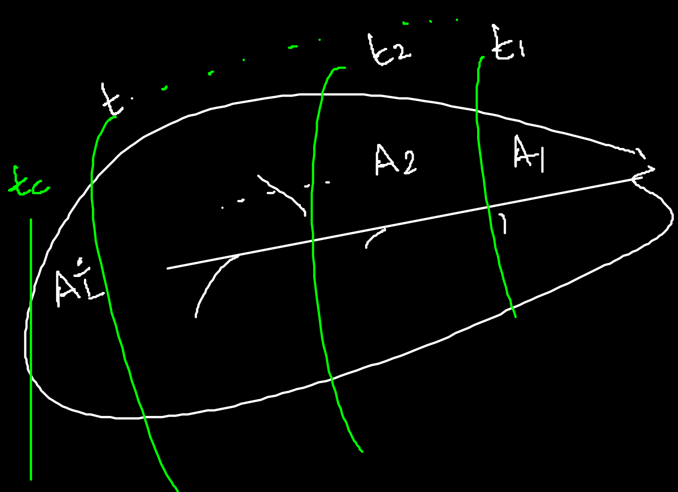
→ A_1 takes 1 hr to start contributing to outlet

A_2 takes 2 hr - - - - - - - -

A_3 - - - - - - - - - - - -

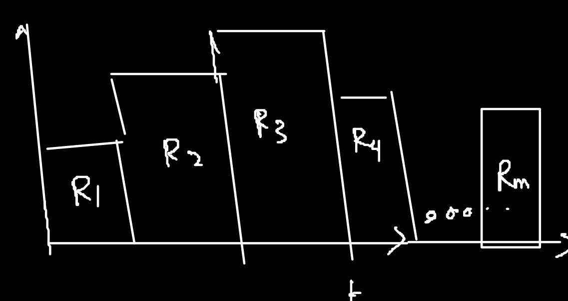
⊛ DRH from time Area Diagram. (Time Area curve)

→ curve which gives inter isochronal areas v/s travel time



⊛ Let $A_1, A_2, A_3 \cdots A_I$ be the inter-isochronal areas.

⊛ Let $R_1, R_2 \ldots R_N$ be the effective rainfall.



⊛ Let $Q_1, Q_2 \ldots Q_N$ be the DRH ordinates,

* Total No. of ordinates in TAD = I (i)

" " " in ERH = M (m)

" " " DRH = N (m)

* DRH ordinate

Q_n ⇒ Given using Discrete Convolution Eq^n.

$$\boxed{Q_n = \sum_{i=1}^{n \le M} \left(R_i \, A_{(n-i+1)} \right)}$$

→ Translates the input R_i to Q_n through time

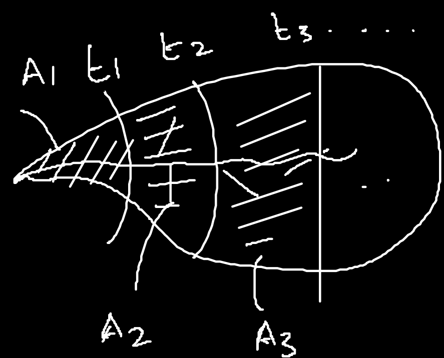
Note :- i → (1) to $\boxed{(n \le M)}$

n = DRH ordinate index

M = Total no. of impulses in ERH.

* when $n = 1$

⇒ $\boxed{Q_1 = R_1 A_1}$



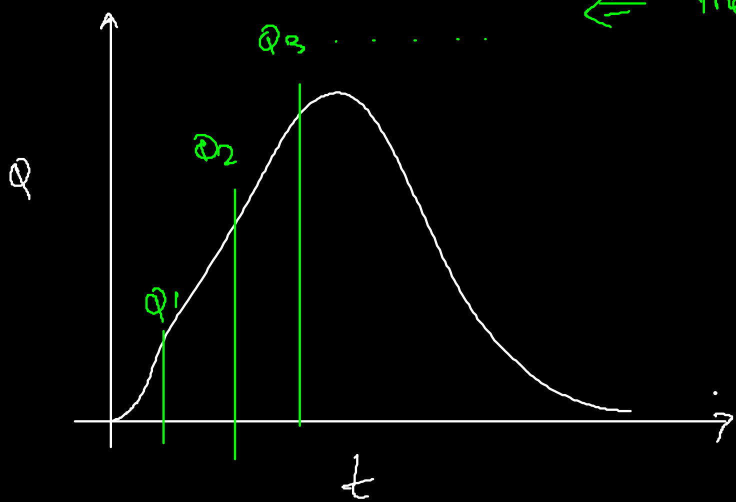
⇒ The contribution from the catchment is fully from A_1 before time t_1.

* when $n = 2$: $Q_2 = R_1 A_2 + R_2 A_1$

↳ The Runoff response at end of 2^{nd} hour = Rainfall that fell on A_2 during first hour (R_1) + RF that fell on A_1 during R_2.

When n = 3

$$Q_3 = R_1A_3 + R_2A_2 + R_3A_1 + \cdots$$



← The ordinates are $Q_1, Q_2 \ldots Q_3 \ldots Q_4 \ldots$

(A) In the Rainfall - Runoff process.

- Translation in Time
- Attenuation due to storage

$\boxed{\text{UNIT HYDROGRAPH Accounts for both}}$ → More accurate

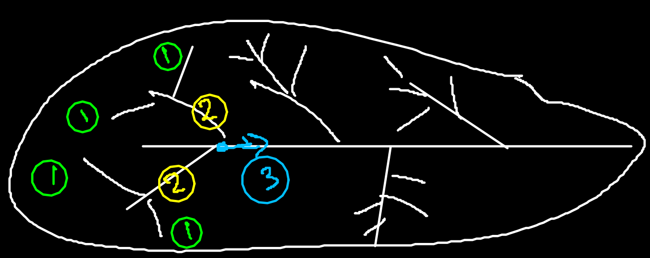
(A) Limitation of TAD

- Accounts only for translation.

Geomorphological Parameters

(A) Hortons stream ordering

- Smallest recognizable channel → $\boxed{\text{order 1.}}$
- when 2 channels of order 'i' join ⇒ New channel order (i+1) is formed.



1 + 1 ⇒ 2
1 + 2 ⇒ 2

2 + 2 = 3
⋮

* Bifurcation Ratio :- R_B is $\left(\frac{N_i}{N_{i+1}}\right) \approx (3 \text{ to } 5)$

N_i = No. of i^{th} order channels.

N_{i+1} = No. of $(i+1)^{th}$ order channels.

* Higher R_B → More u/s channel ⇒ High Drainage.

* Length Ratio, $\boxed{R_L = \left(\frac{L_{i+1}}{L_i}\right)}$ ⇒ Law of stream lengths

L_i = Avg. length of i^{th} order streams.

* Higher L_i → Good drainage

* Area Ration ; $\boxed{R_A = \left(\frac{A_{i+1}}{A_i}\right)}$

A_i = Avg. catchment area drained by i^{th} order stream.

→ If R_A is high → Greater area drained by d/s channel
→ Good drainage.

* Drainage Density (D) $= \left(\dfrac{\sum_{i=1}^{I}\sum_{j=1}^{N_i} L_{ij}}{A_I}\right)$

L_{ij} = Length of j^{th} stream of order i

⇒ $\dfrac{\text{Total length of streams}}{\text{Total Area.}}$

$\boxed{\text{If D is high ⇒ Good drainage}}$

$\boxed{L_o = \frac{1}{2D}}$ ⇐ Avg. length of overland flow