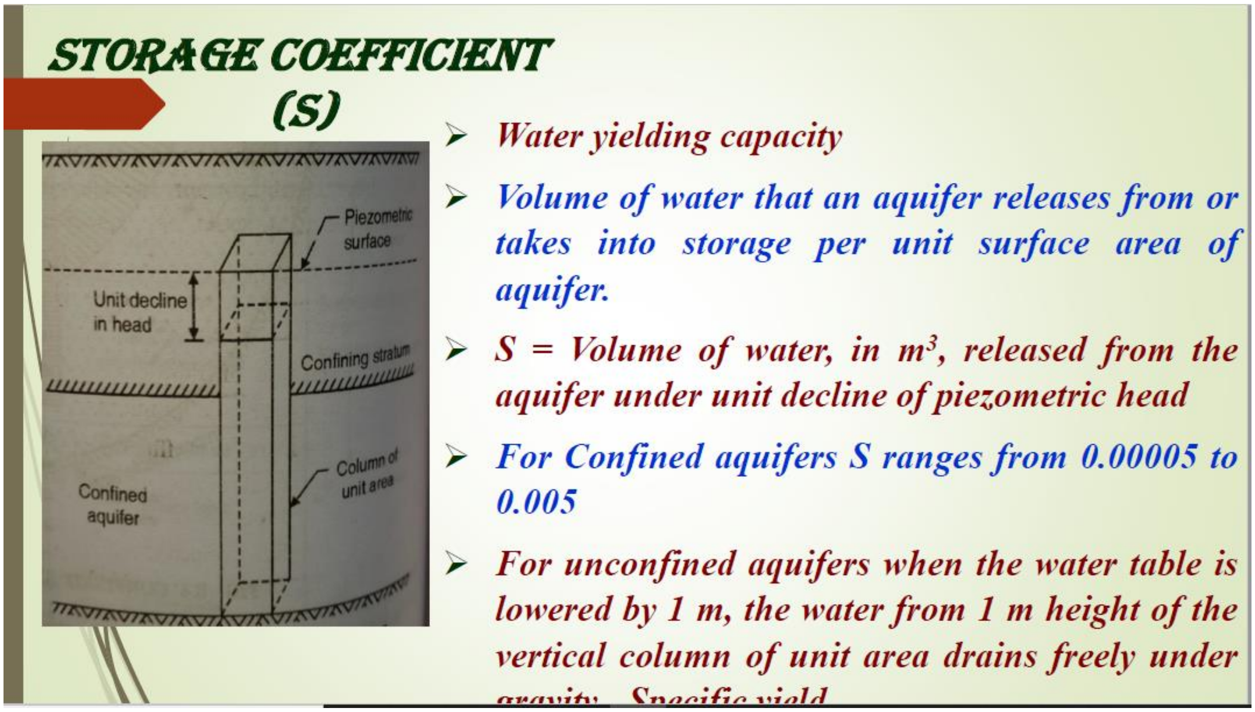
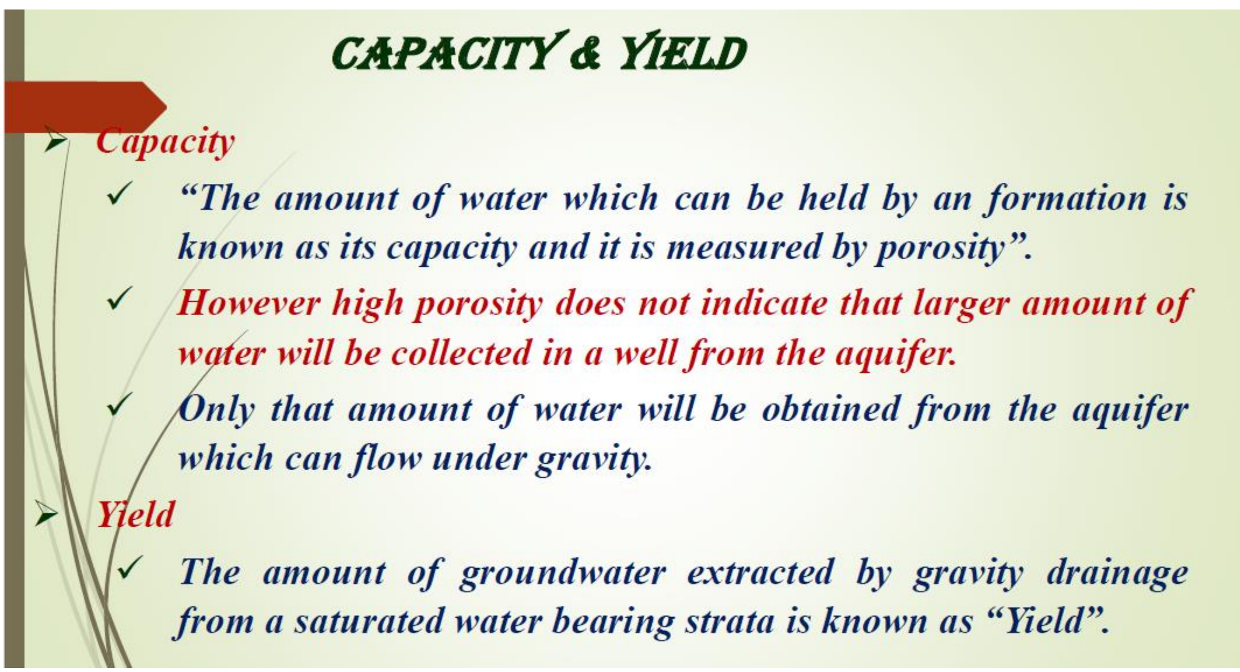
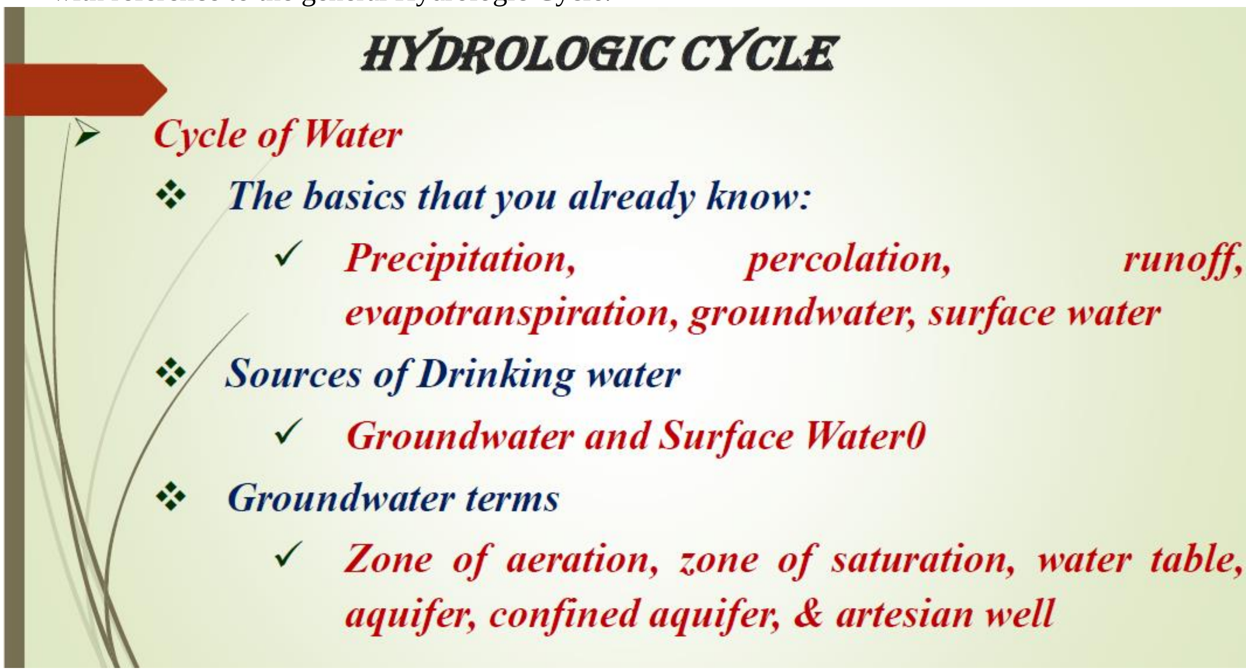
Unit 1

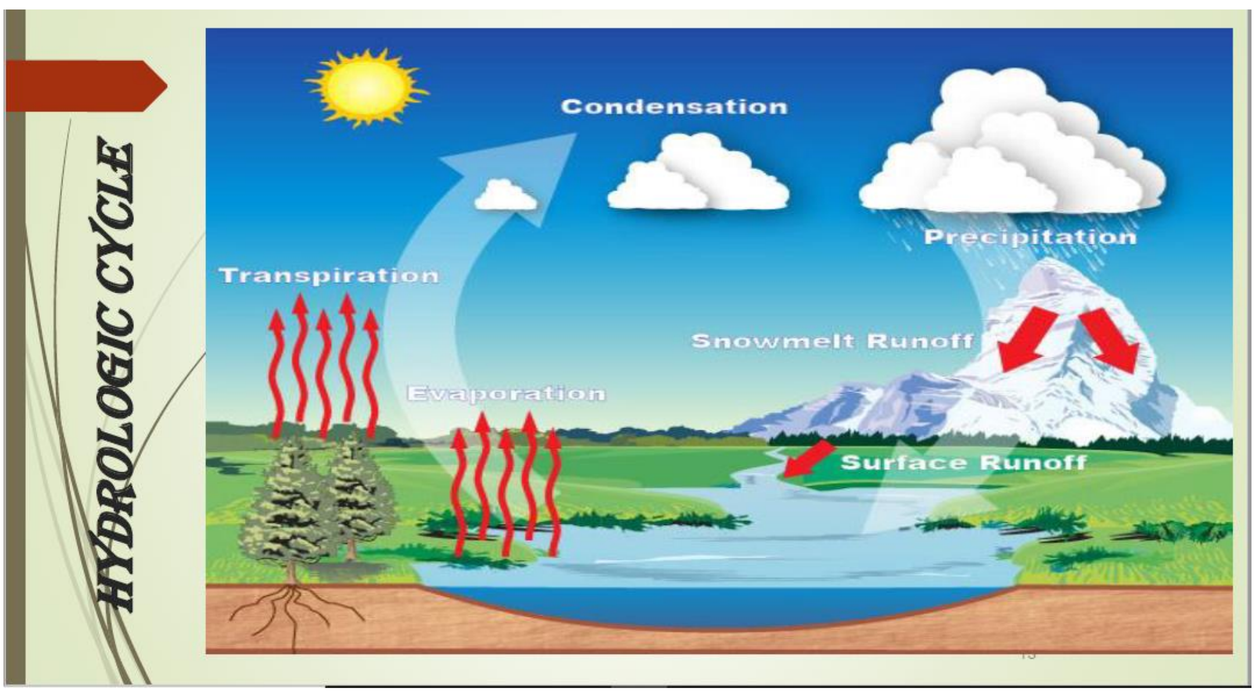
1. Define the terms (i) Storage Coefficient, (ii) Capacity and (iii) Yield.





2. With the help of a neat sketch identify the various sources of storage and movement of groundwater with reference to the general Hydrologic Cycle.



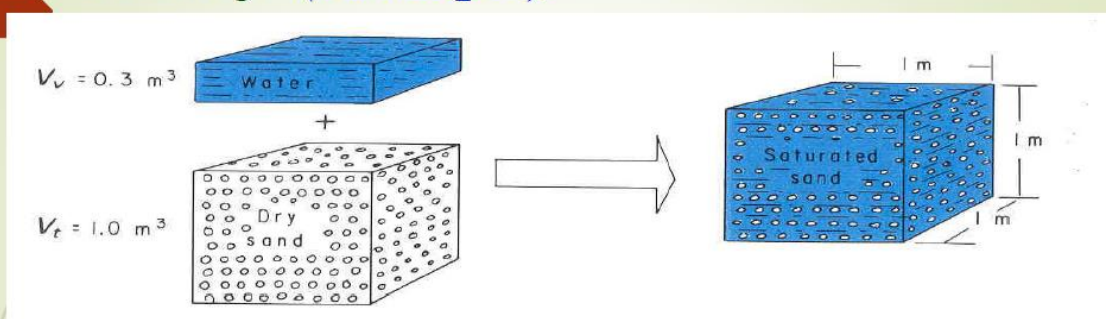


3. Define the terms (i) Porosity, (ii) Specific Yield and (iii) Specific Retention.

Porosity (Expressed as %)

- *Ratio of openings (voids) to the total volume of a soil*
- $n = (V_{t} - V_{s})/V_{t} = V_{v}/V_{t}$
- *Vt= total volume of the soil or rock*
- *Vs= Volume of solids in the sample*
- *Vv= Volume of openings (voids)*

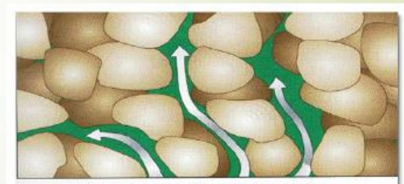
Porosity (Example)



❖ *Porosity* $n = (V_{t} - V_{s})/V_{t} = V_{v} / V_{t} = 0.3 / 1.0 = 30\ \%$

POROSITY AND PERMEABILITY

- *Porosity - the percentage of rock that consists of voids or openings*
  - ✓ *Volume of empty space in a rock*
  - ✓ *A rock's ability to hold water*
  - ✓ *Loose sand has ~30-50% porosity*
  - ✓ *Compacted sandstone may have only 10-20% porosity*
- *Permeability - the capacity of a rock to transmit fluid through pores and fractures*
  - ✓ *Interconnectedness of pore spaces*

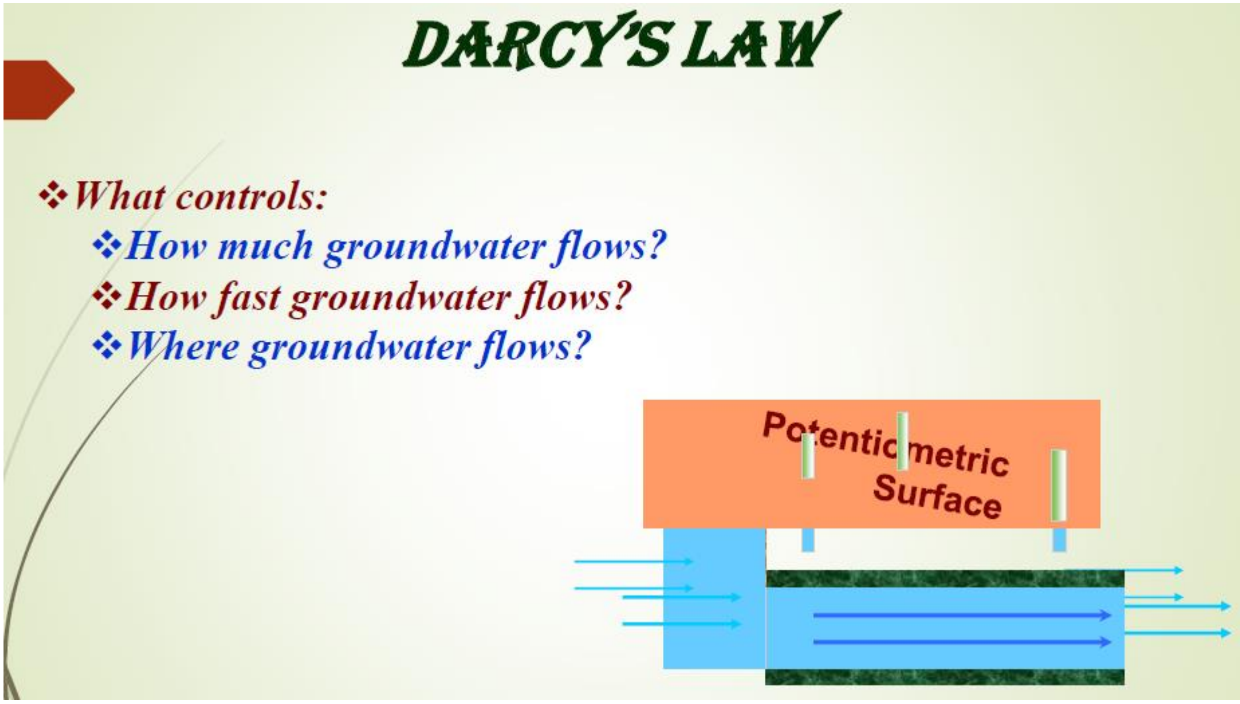


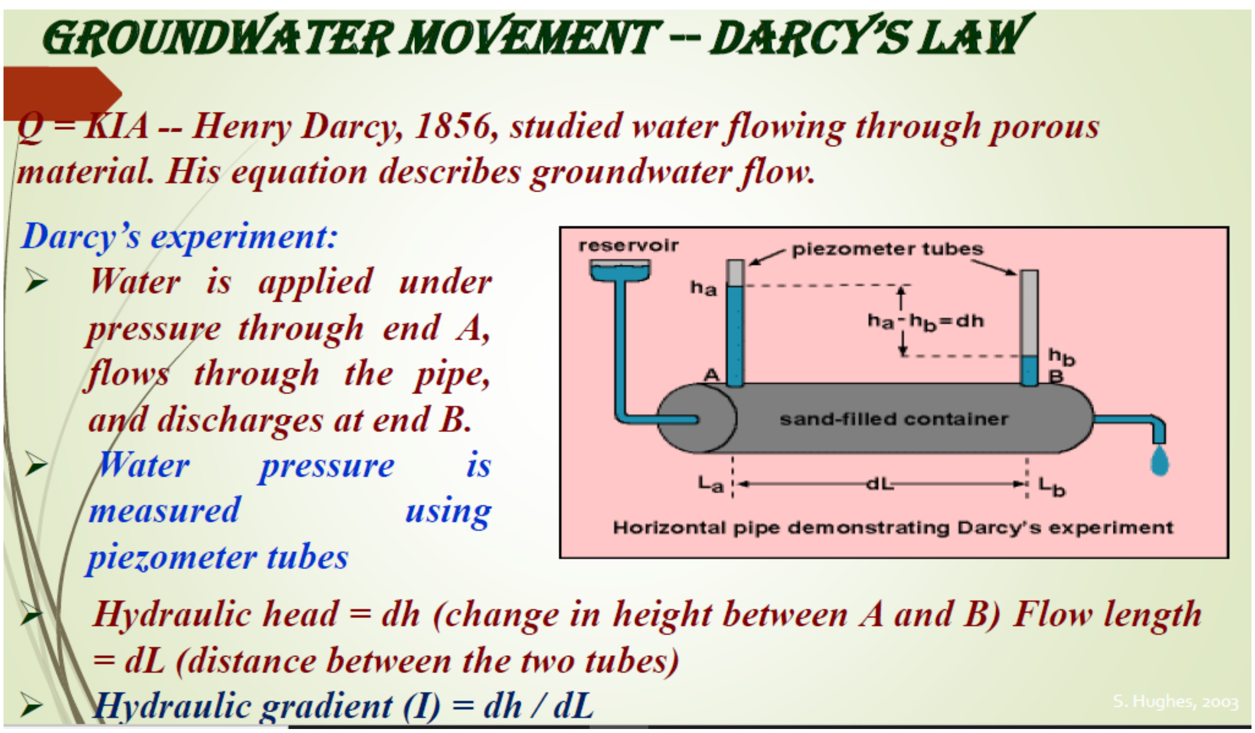
Connected pores give a rock permeability.

- ***SPECIFIC YIELD (S_y) is the ratio of the volume of water drained from a rock (due to gravity) to the total rock volume. Grain size has a definite effect on specific yield. Smaller grains have larger surface area/volume ratio, which means more surface tension. Fine-grained sediment will have a lower Sy than coarse-grained sediment.***
- ***SPECIFIC RETENTION (S_r) is the ratio of the volume of water a rock can retain (in spite of gravity) to the total volume of rock.***
- ***Specific yield plus specific retention equals porosity (often designated with the Greek letter phi):***

$$Sr + Sy = \Phi$$

4. State Darcy's law for movement of groundwater in an aquifer. What are the limitations of the same?





Groundwater Movement

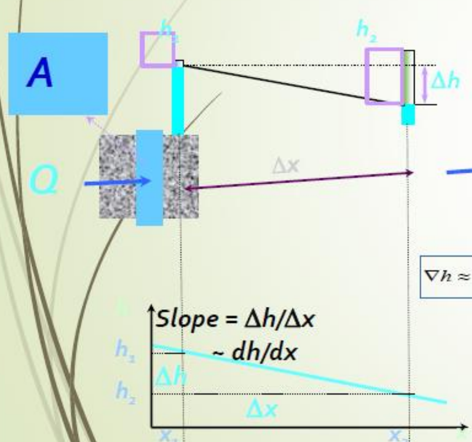
Porosity and hydraulic conductivity of selected earth materials

Material	*Porosity (%)*	*Hydraulic Conductivity (m/day)*
Unconsolidated		
Clay	***45***	***0.041***
Sand	***35***	***32.8***
Gravel	***25***	***205.0***
Gravel and sand	***20***	***82.0***
Rock		
Sandstone	***15***	***28.7***
Dense limestone or shale	***5***	***0.041***
Granite	***1***	***0.0041***

Darcy's Law

Henry Darcy's Experiment (Dijon, France 1856)

Darcy investigated ground water flow under controlled conditions



$$\boxed{Q \propto \Delta h, \quad Q \propto 1/\Delta x, \quad Q \propto A}$$

Q: Volumetric flow rate [$L^{3/T}$]

A: Cross Sectional Area (Perp. to flow)

K: The proportionality constant is added to form the following equation:

$\nabla h \approx \frac{\Delta h}{\Delta x}$ *: Hydraulic Gradient*

$$\boxed{Q \propto A\frac{\Delta h}{\Delta x} \quad \Rightarrow \quad Q = -K\,A\frac{\Delta h}{\Delta x}}$$

K units [L/T]

LIMITATIONS OF DARCY'S LAW

- *Flow is Laminar.*
- *Reynolds Number defined as $N_R = v\,D/\nu$, where v = velocity, D= A representative length and ν = kinematic viscosity of the fluid is <5.*
- *The soil media is isotropic (K independent of direction).*
- *Soil is homogeneous.*

5. Write a note on (i) Anisotropy and (ii) Heterogeneity w.r.t. ground water flow through aquifers.

2.5 Anisotropy and heterogeneity

The well-flow equations presented in this manual are based on several assumptions, one of which is that aquifers and aquitards are homogeneous and isotropic. This means that the hydraulic conductivity is independent of where it is measured within the formation and also independent of the direction of measurement (Figure 2.4.A). The individual particles of geological formations are seldom spherical, so when deposited under water they tend to settle on their flat sides. Such a formation can still be homogeneous, but the hydraulic conductivity varies with the direction of measurement (Figure 2.4.B). In this particular case the hydraulic conductivity K_h measured in the horizontal plane is significantly greater than the hydraulic conductivity K_v measured in the vertical plane. This phenomenon is called anisotropy. In alluvial formations the K_h/K_v ratios normally range from 2 to 10, but values as high as 100 do occur.

The lithology of geological formations generally varies significantly in both horizontal and vertical planes. Consequently, the hydraulic conductivity now depends on the position within the formation. Such a formation is called heterogeneous. Figure 2.4.C is an example of layered heterogeneity. If the hydraulic conductivity of the individual layers also varies in the direction of

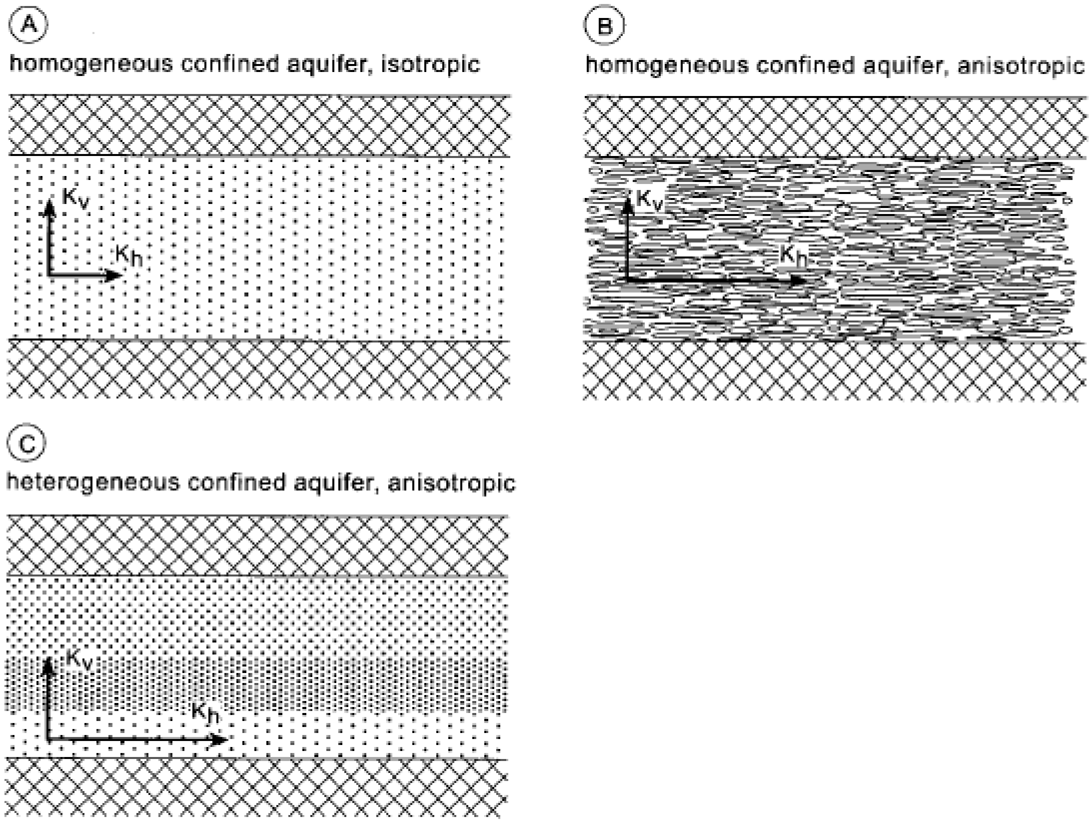
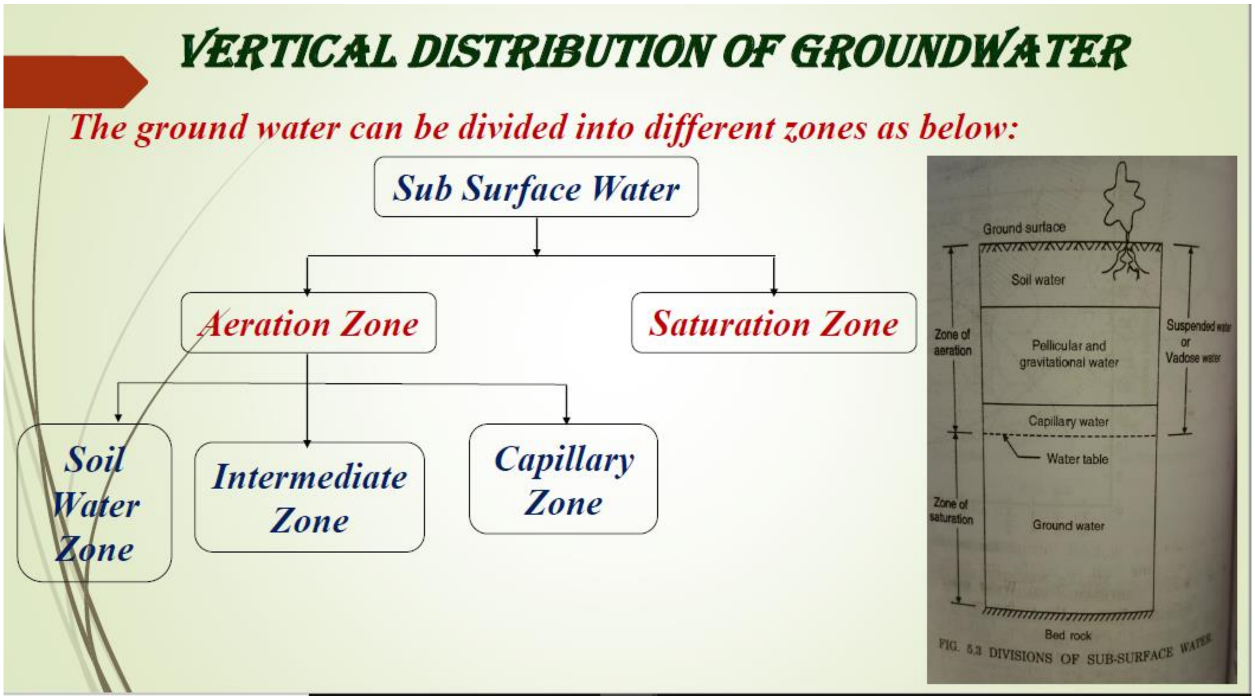
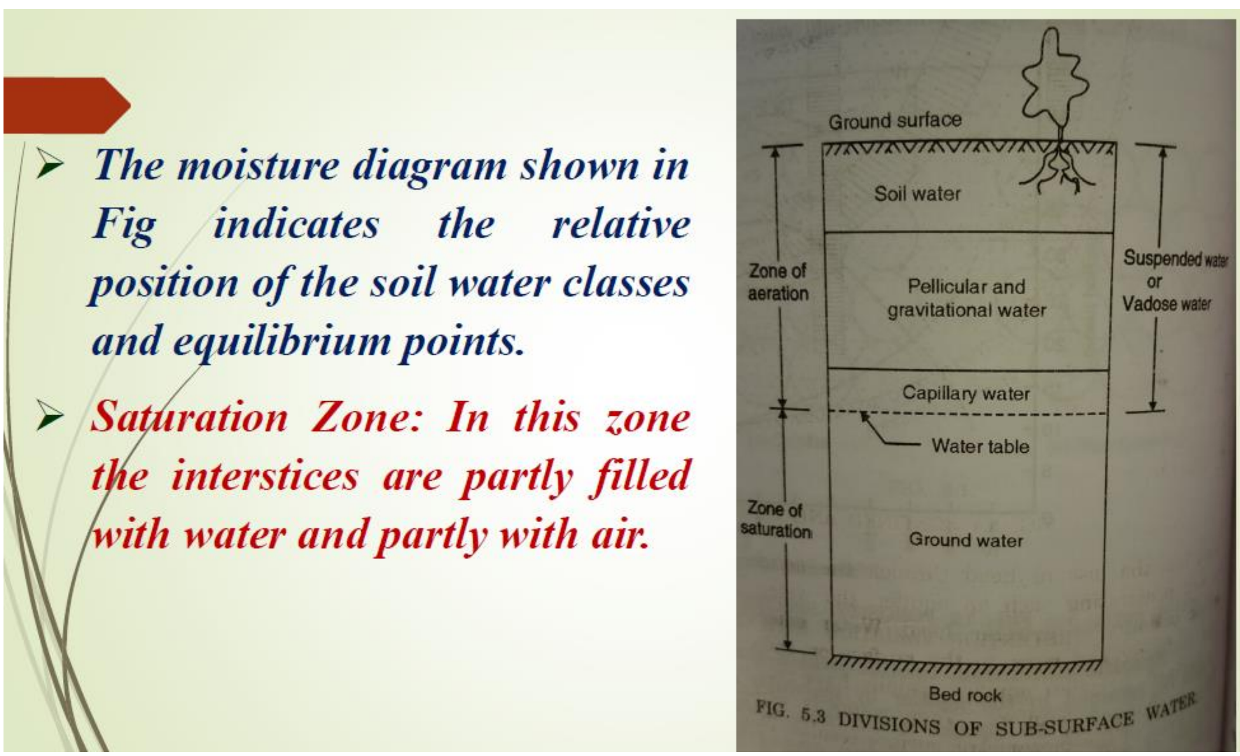


Figure 2.4 Homogeneous isotropic aquifer, homogeneous anisotropic aquifer, and heterogeneous anisotropic aquifer

measurement, the formation is, moreover, anisotropic. Heterogeneity appears in forms other than that shown in Figure 2.4.C: individual layers may pinch out, their grain size may vary in the horizontal plane, they may contain lenses of other grain sizes, or they may be discontinuous because of faulting.

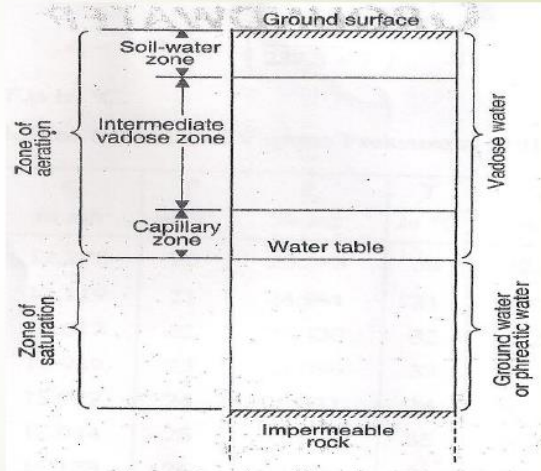
6. Discuss how groundwater is distributed vertically in the form of different zones.





DIVISION OF SUB-SURFACE WATER

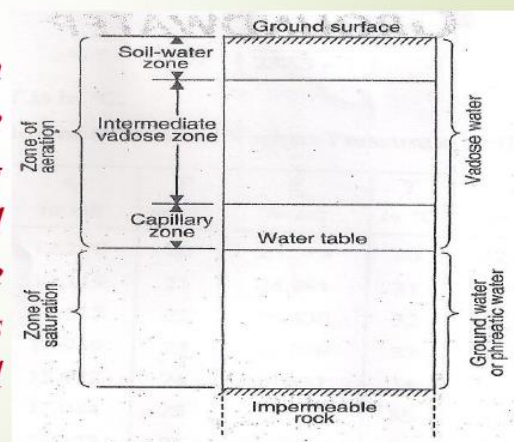
- **Soil Water Zone**
- *This zone extends "from the ground surface down through the major root-zone of vegetations". Its thickness varies with the type of soil and vegetation*



- **Intermediate Zone**

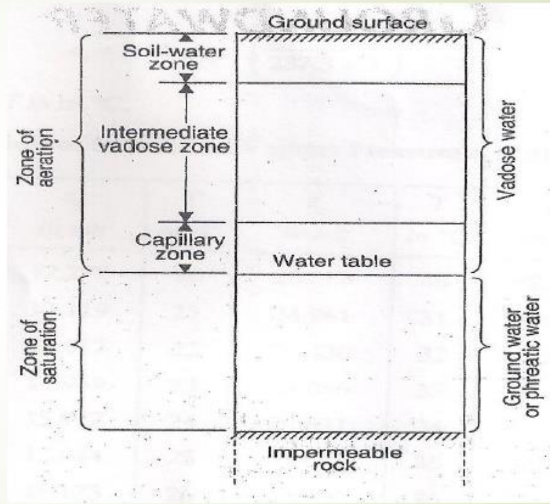
DIVISION OF SUB-SURFACE WATER

- *This zone extends "from the lower edge of the soil water zone to the upper limit of the capillary zone".*
- *This zone varies in thickness from zero, when the grounding zone merge with a high water table approaching to the ground, to several hundred meter under deep water table condition. This zone contains pellicular as well as gravitational water.*



DIVISION OF SUB-SURFACE WATER

- ***Capillary Zone***
- ***It extends "<u>from the water table upto the limit of capillary rise of water</u>".***



7. Distinguish between Aquifer, Aquifuge and Aquitard giving examples for each of them.

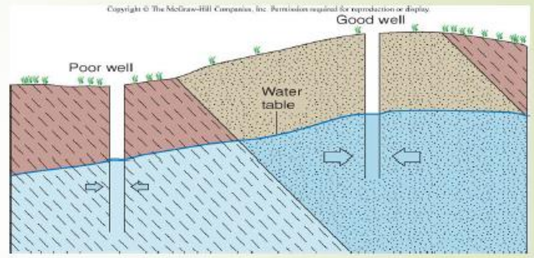
GROUND WATER

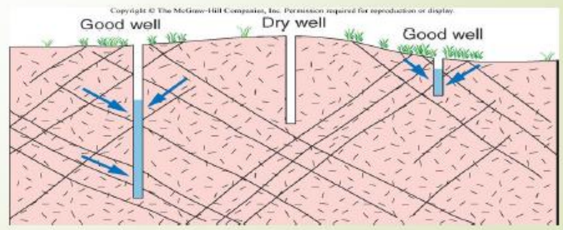
- ***Aquifer***
  - ***Aquifer - Groundwater Reservoir and Water-bearing Formation.***
  - ***A rock unit that will yield water in a usable quantity to a well or spring.(saturated geological formation, containing and transmitting significant quantities of water under normal field quantities); rock: unconsolidated sediments***
  - ***"Geologic formation containing sufficient saturated permeable material that yields significant quantities of water is defined as aquifers".***
  - ***This implies an ability to store and to transmit water.***
  - ***The sand and gravel aquifers produce large quantities of water, most of which is replenished by seepage from streams into alluvial formations.***
  - ***All developed aquifers probably consist of unconsolidated rocks, chiefly gravel and sand.***

AQUIFERS AND AQUITARDS

Aquifer - body of saturated rock or sediment through which water can move easily

- *Sandstone*
- *Conglomerate*
- *Well-jointed limestone*
- *Highly fractured rock*





GROUND WATER

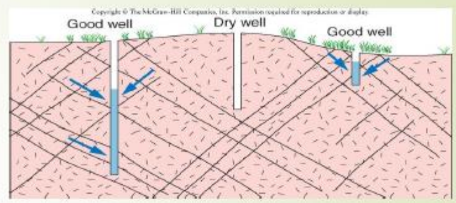
- *Aquifers are generally extensive and may be overlain or underlain by a Confining Bed which may be defined as a relatively impermeable material stratigraphically adjacent to one or more aquifers.*
- *Various types of confining Beds:*
  - ✓ *Aquiclude - Formation containing water → Do not transmit significant quantities*
  - ✓ *Saturated but relatively impermeable material that does not yield appreciable quantities of water to wells; Clay is an example*

GROUND WATER

- ✓ *Formations with low permeability…includes both aquiclude and aquifuge*
- ✓ *Aquifuge - Formation → doesnot contain nor transmit*
- ✓ *Relatively impermeable formation neither contining nor transmitting water ; Solid Granite belongs to this category.*

AQUIFERS AND AQUITARDS

Aquitard - Saturated but poorly permeable stratum that impedes groundwater movement and does not yield water freely to wells, that may transmit appreciable water to or from adjacent aquifers and, where sufficiently thick, may constitute an important groundwater storage zone;



Sandy Clay is an example rock/sediment that retards ground water flow due to low porosity and/or permeability

- *Shale, clay, unfractured crystalline rocks*

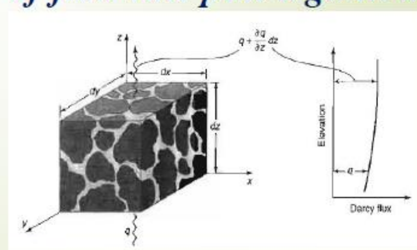
Unit 2

1. Explain the control volume approach for development of groundwater flow equations.

GENERAL GROUNDWATER FLOW EQUATIONS

- *Consider the control volume (CV) for a saturated flow as shown in the fig.*
- *The sides, which define the control surface (CS), have lengths dx, dy and dz in the coordinate directions.*
- *The total volume of the CV is dxdydz and the volume of water flowing into or out of the CV is θ dxdydz, where 'θ' is the moisture content.*
- *Rate of outflow of GW thro' the CS is = Rate of change of GW stored in the CV. V = Velocity Vector and the volume of flow rate past a given area 'dA' is V.dA, where 'A' is the area vector.*

Control Volume for development of the continuity equation in porous medium.



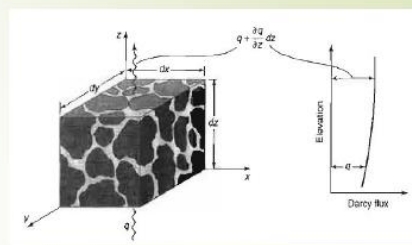
GENERAL GROUNDWATER FLOW EQUATIONS

- *The general control volume equation for continuity is applicable and is as Follows:*

$$0 = \frac{d}{dt}\int_{CV} \rho . d\forall + \int_{CS} \rho\vec{V} . dA \text{ - - (1) or}$$

$$\int_{CS} \rho V . dA = -\frac{d}{dt}\int_{CV} \rho . d\forall$$

where d$\forall$ isthe elemental volume given by dxdydz.



- *The time rate of change of mass stored in the CV is defined as the rate of change of fluid mass storage. Expressed as,*

$$\frac{d}{dt}\int_{CV} \rho . d\forall = \rho S_S \frac{\partial h}{\partial t}(dxdydz) + \rho W(dxdydz)$$

where,SS is the specific storage and W is the flow into or out of the CV. W is given by: $W = Q/(dxdydz)$

GENERAL GROUNDWATER FLOW EQUATIONS

$\int_{CS} \rho V . dA = -\frac{d}{dt}\int_{CV} \rho . d\forall$ *RHS - Time rate of change of mass stored in CV:*

$\frac{d}{dt}\int_{CV} \rho . d\forall = \rho S_S \frac{\partial h}{\partial t}(dxdydz) + \rho W(dxdydz)$ *- (2)* $W = Q/(dxdydz)$

- *The term* $\rho S_S \frac{\partial h}{\partial t}(dxdydz)$ *indicates the mass rate of water produced by:*
  1) *An expansion of the water under change in density and*
  2) *The compaction of the porous medium due to change in the porosity.*
- *LHS: The inflow of water through the CS at the bottom of CV = qdxdy and the outflow at the top of CV "[q+(∂q/ ∂z)dz]dxdy". So the net outflow in the vertical direction is "*$\rho dxdydz \frac{\partial q}{\partial z}$*" - - (3a).*

  "$\frac{\partial q}{\partial z}$" *is denoted as "*$q_z$*"*
- *Since the inflow and out flow will occur in all three directions the term is sum of the net flows in all the 3 directions , x, y and z.*

S. Hughes, 2003

GENERAL GROUNDWATER FLOW EQUATIONS

- *Net outflow in the vertical (z) direction is "*$\rho dxdydz \frac{\partial q}{\partial z}$*" - - (3a)*

$$\int_{CS} \rho V . dA = \rho dxdydz \frac{\partial q}{\partial x} + \rho dxdydz \frac{\partial q}{\partial y} + \rho dxdydz \frac{\partial q}{\partial z} \text{ - - (3)}$$

- *Substituting (2) & (3) in eq(1):*

$$0=\rho S_S \frac{\partial h}{\partial t}(dxdydz) + \rho W(dxdydz) + \rho dxdydz \frac{\partial q}{\partial x} + \rho dxdydz \frac{\partial q}{\partial y} + \rho dxdydz \frac{\partial q}{\partial z} \text{ - - (4a)}$$

- *Dividing by* $\rho(dxdydz)$ *(4a) becomes:*

$$S_S \frac{\partial h}{\partial t} + \frac{\partial q}{\partial x} + \frac{\partial q}{\partial y} + \frac{\partial q}{\partial z} + W = 0 \text{ OR}$$

$$S_S \frac{\partial h}{\partial t} + q_x + q_y + q_z + W = 0 \text{ - - (4)}$$

GENERAL GROUNDWATER FLOW EQUATIONS

- *Darcy' Law:* $q_s = -K_s \frac{\partial h}{\partial s}$ *where q_s is the flux in s-direction and K_s is the coefficient of permeability in s-direction.*
- *Using Darcy's law eqn (4) can be written as:*

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) = S_s \frac{\partial h}{\partial t} + W \text{ - - - (5)}$$

- *Eq(5) is the equation for three-dimensional transient flow through a saturated anisotropic porous medium.*
- *For a homogeneous, isotropic medium* $K_x = K_y = K_z = K$

$$S_S \frac{\partial h}{\partial t} - K\left[\frac{\partial}{\partial x}\left(\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(\frac{\partial h}{\partial z}\right)\right] + W = 0 \text{ - - (5a)}$$

$$\left[\frac{\partial}{\partial x}\left(\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(\frac{\partial h}{\partial z}\right)\right] = \frac{S_S}{K}\frac{\partial h}{\partial t} + \frac{W}{K} \text{ - - (5b)}$$

GENERAL GROUNDWATER FLOW EQUATIONS

- *For steady flow* $\frac{\partial h}{\partial t} = 0$. *Hence* $\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{W}{K}$ *- - (5c)*
- *For a horizontal confined aquifer of thickness 'b',* $S = S_S b$ *and the transmissivity* $T = Kb$. *With* $W = 0$, *the equation for the unsteady flow through an isotropic homogeneous 2-dimensional flow through a confined aquifer will be:*

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{S}{T}\frac{\partial h}{\partial t} \text{ - - (6)}$$

- *Eqn (6) for radial flow will be: (using the relation* $r = \sqrt{x^2 + y^2}$*)*

$\frac{\partial^2 h}{\partial r^2} + \frac{1}{r}\frac{\partial h}{\partial r} = \frac{S}{T}\frac{\partial h}{\partial t}$ *- - (7) where r is the radial distance from pumped well and t is the time since the beginning of pumping.*

- *Eqn (7) is known as "Diffusion Equation"*

2. State Dupuit- Forchheimer assumptions for two-dimensional ground water flow.

ELEMENTARY GROUNDWATER FLOW

1) *The flow is horizontal at any vertical cross-section.*
2) *The velocity is constant over the depth.*
3) *The velocity is calculated using the slope of the free surface as the hydraulic gradient.*
4) *The slope of the water table is relatively small.*

3. Considering the general continuity equation for a derive the "Diffusion Equation" two-dimensional radial flow in to a well located in a confined aquifer.

GENERAL GROUNDWATER FLOW EQUATIONS

- *For steady flow* $\frac{\partial h}{\partial t} = 0$. *Hence* $\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{W}{K}$ *- - (5c)*
- *For a horizontal confined aquifer of thickness 'b', $S = S_S b$ and the transmissivity $T = Kb$. With $W = 0$, the equation for the unsteady flow through an isotropic homogeneous 2-dimensional flow through a confined aquifer will be:*

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{S}{T}\frac{\partial h}{\partial t} \quad \text{- - (6)}$$

- *Eqn (6) for radial flow will be: (using the relation* $r = \sqrt{x^2 + y^2}$*)*

$\frac{\partial^2 h}{\partial r^2} + \frac{1}{r}\frac{\partial h}{\partial r} = \frac{S}{T}\frac{\partial h}{\partial t}$ *- - (7) where r is the radial distance from pumped well and t is the time since the beginning of pumping.*

- *Eqn (7) is known as "Diffusion Equation"*

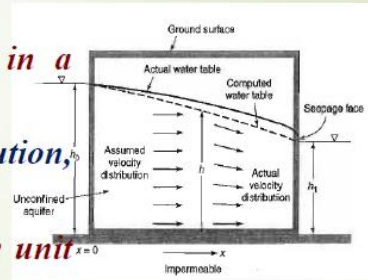
4. Why Laplace equation cannot be used for flow through an unconfined aquifer? Explain how the same is overcome by Dupuit's assumptions.

UNCONFINED AQUIFER

- *Solution of Laplace eqn for this case is not possible since the water table which is a flow line is also an unknown bobdary.*
- *To obtain a solution, Dupuit has made the following assumptions:*

i. *The velocity of flow is proportional to the tangent of the hydraulicgradient instead of the sine as defned in the eqn :* $v = -K\frac{\partial h}{\partial s}$

ii. *The flow to be horizontal and uniform everywhere in a vertical section.*

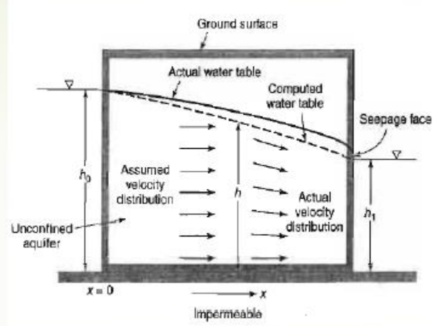


- *These assumptions, although permit obtaining a solution, limit the applicability of the results.*
- *For a unidirectional flow as shown, the discharge per unit width q at any vertical section is given by:* $v = -Kh\frac{dh}{dx}$

where K = hydraulic conductivity, h = ht. of the water table and x is the direction of flow.

UNCONFINED AQUIFER

$$v = -Kh\frac{dh}{dx}$$



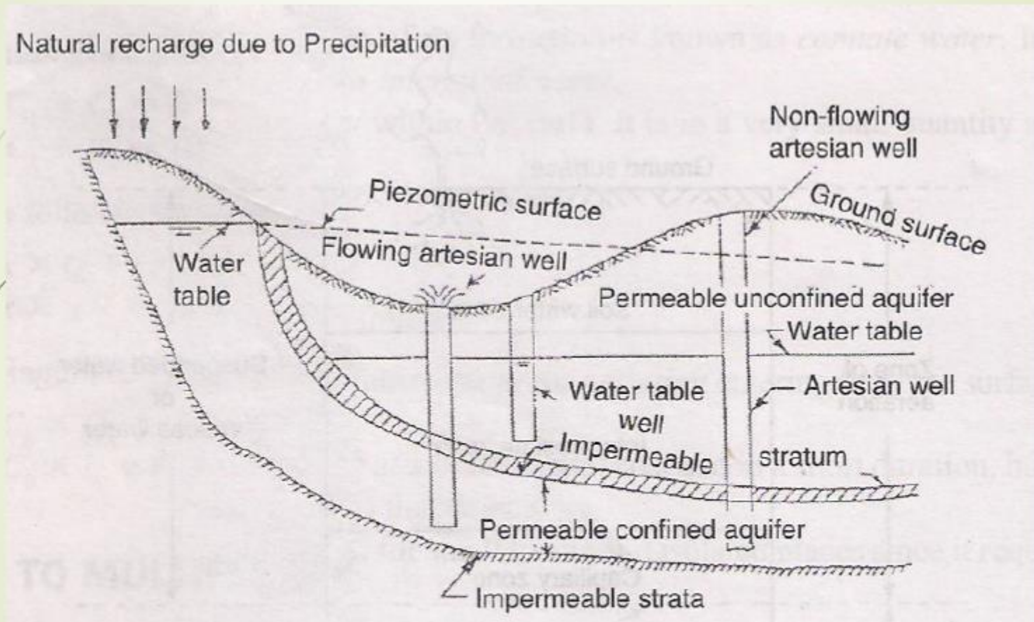
- *Integrating, we get* $qx = -\frac{K}{2}h^2 + C$ *and, if h =* h_0 *where x = 0, then the Dupuit eqn for unconfined aquifer will be:*

$$qx = \frac{K}{2x}\left(h_0^2 - h^2\right) \quad \text{-- (3)}$$

- *For flow between two fixed boundaries of water of constant heads h0 and h1 as shown, the water table slope at u/s boundary of the aquifer, (neglecting capillary zone)* $\frac{dh}{dx} = -\frac{q}{Kh_0}$

5. Derive equation steady state radial flow in to wells in any aquifier.

WELLS



GENERAL GROUNDWATER FLOW EQUATIONS

$\int_{CS} \rho V . dA = -\frac{d}{dt}\int_{CV} \rho . d\forall$ ***RHS - Time rate of change of mass stored in CV:***

$\frac{d}{dt}\int_{CV} \rho . d\forall = \rho S_S \frac{\partial h}{\partial t}(dxdydz) + \rho W(dxdydz)$ *- (2)* $W = Q/(dxdydz)$

- ***The term*** $\rho S_S \frac{\partial h}{\partial t}(dxdydz)$ ***indicates the mass rate of water produced by:***
  1) ***An expansion of the water under change in density and***
  2) ***The compaction of the porous medium due to change in the porosity.***
- ***LHS: The inflow of water through the CS at the bottom of CV = qdxdy and the outflow at the top of CV "[q+(∂q/ ∂z)dz]dxdy". So the net outflow in the vertical direction is*** "$\rho dxdydz\frac{\partial q}{\partial z}$" *- - (3a).*

"$\frac{\partial q}{\partial z}$" ***is denoted as*** "$q_z$"

- ***Since the inflow and out flow will occur in all three directions the term is sum of the net flows in all the 3 directions , x, y and z.***

GENERAL GROUNDWATER FLOW EQUATIONS

- ***Net outflow in the vertical (z) direction is "$\rho dxdydz\frac{\partial q}{\partial z}$" - - (3a)***

$$\int_{CS}\rho V.dA = \rho dxdydz\frac{\partial q}{\partial x} + \rho dxdydz\frac{\partial q}{\partial y} + \rho dxdydz\frac{\partial q}{\partial z} \text{ - - (3)}$$

- ***Substituting (2) & (3) in eq(1):***

$$0=\rho S_S\frac{\partial h}{\partial t}(dxdydz) + \rho W(dxdydz) + \rho dxdydz\frac{\partial q}{\partial x} + \rho dxdydz\frac{\partial q}{\partial y} + \rho dxdydz\frac{\partial q}{\partial z} \text{ - - (4a)}$$

- ***Dividing by $\rho(dxdydz)$ (4a) becomes:***

$$S_S\frac{\partial h}{\partial t} + \frac{\partial q}{\partial x} + \frac{\partial q}{\partial y} + \frac{\partial q}{\partial z} + W = 0 \text{ OR}$$

$$S_S\frac{\partial h}{\partial t} + q_x + q_y + q_z + W = 0 \text{ - - (4)}$$

GENERAL GROUNDWATER FLOW EQUATIONS

- ***Darcy' Law: $q_s = -K_s\frac{\partial h}{\partial s}$ where q_s is the flux in s-direction and K_s is the coefficient of permeability in s-direction.***
- ***Using Darcy's law eqn (4) can be written as:***

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right) = S_s\frac{\partial h}{\partial t} + W \text{ - - - (5)}$$

- ***Eq(5) is the equation for three-dimensional transient flow through a saturated anisotropic porous medium.***
- ***For a homogeneous, isotropic medium $K_x = K_y = K_z = K$***

$$S_S\frac{\partial h}{\partial t} - K\left[\frac{\partial}{\partial x}\left(\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(\frac{\partial h}{\partial z}\right)\right] + W = 0 \text{ - - (5a)}$$

$$\left[\frac{\partial}{\partial x}\left(\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(\frac{\partial h}{\partial z}\right)\right] = \frac{S_S}{K}\frac{\partial h}{\partial t} + \frac{W}{K} \text{ - - (5b)}$$

GENERAL GROUNDWATER FLOW EQUATIONS

- *For steady flow* $\frac{\partial h}{\partial t} = 0$*. Hence* $\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{W}{K}$ *- - (5c)*
- ***For a horizontal confined aquifer of thickness 'b', $S = S_S b$ and the transmissivity $T = Kb$. With $W = 0$, the equation for the unsteady flow through an isotropic homogeneous 2-dimensional flow through a confined aquifer will be:***

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{S}{T}\frac{\partial h}{\partial t} \quad \text{- - (6)}$$

- ***Eqn (6) for radial flow will be: (using the relation $r = \sqrt{x^2 + y^2}$)***

$\frac{\partial^2 h}{\partial r^2} + \frac{1}{r}\frac{\partial h}{\partial r} = \frac{S}{T}\frac{\partial h}{\partial t}$ ***- - (7) where r is the radial distance from pumped well and t is the time since the beginning of pumping.***

- ***Eqn (7) is known as "Diffusion Equation"***

S. Hughes, 2003

ANALYTICAL SOLUTIONS

- *Again for steady flow* $\frac{\partial h}{\partial t} = 0$ *and eqn (7) becomes:*

$$\frac{1}{r}\frac{\partial}{\partial r}\left(r\frac{\partial h}{\partial r}\right) = 0 \quad \text{- - (8)}$$

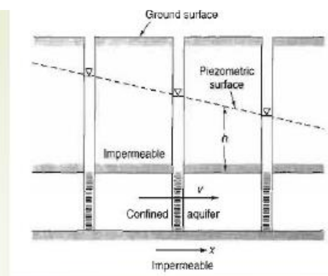
- ***Eqn (8) can be adopted for different aquifers with proper boundary conditions.***
- ***For solution of any problem, idealization of of aquifer and the boundary conditions of the flow system is necessary.***
- ***Results may only approximate the field conditions.***
- ***Known deviations from assumptions allow analytical solutions to be modified to obtain an answer that otherwise would not have been possible.***
- ***A common assumption regarding the aquifer is that it is homogeneous and isotropic.***

ANALYTICAL SOLUTIONS

- *Often aquifers can be assumed to be infinite in areal extent; if not boundaries are assumed to be*
  1) *Impermeable, such as underlying or overlying rock or clay layers, dikes, faults, or valley walls; or*
  2) *Permeable, including surface water bodies in contact with the aquifer, ground surfaces where water emerges from underground, and wells.*
- *As a first attempt, we will consider Steady, Unidirectional flow in confined and unconfined aquifers.*
- *Flow conditions differ for confined and unconfined aquifers and hence they need to be considered separately with flow in one direction.*

CONFINED AQUIFER

$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=\frac{W}{K}$$



- ***The general equation for steady groundwater flow in isotropic homogeneous porous medium is:***

$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=\frac{W}{K} \text{ - - (5c)}$$

- ***Now the equation for steady groundwater flow with a velocity 'v' in the x-direction in the confined aquifer (with $W = 0$) will be:***

$\frac{\partial^2 h}{\partial x^2} = 0$ *- - (1), the solution of this equation is* $h = C_1 x + C_2$ *- - (1a) where h = the head above a given datum and C_1 & C_2 are constants of integration.*

- ***Assuming $h = 0$ when $x = 0$ we get $C_2 = 0$ and $C_1 = \frac{\partial h}{\partial x}$.***
- ***Further, from Darcy' law we have $\frac{\partial h}{\partial x} = -(v/K)$ and thus $C_1 = \frac{\partial h}{\partial x}$.***
- ***Substituting the value of the constants in (1a) we obtain $h = -\frac{vx}{K}$ - - (2)***

6. From the first principles arrive at the analytical solution for steady radial flow through a confined aquifer.

ANALYTICAL SOLUTIONS

- *Again for steady flow* $\frac{\partial h}{\partial t} = 0$ *and eqn (7) becomes:*

$$\frac{1}{r}\frac{\partial}{\partial r}\left(r\frac{\partial h}{\partial r}\right) = 0 \quad \text{-- (8)}$$

- *Eqn (8) can be adopted for different aquifers with proper boundary conditions.*
- *For solution of any problem, idealization of of aquifer and the boundary conditions of the flow system is necessary.*
- *Results may only approximate the field conditions.*
- *Known deviations from assumptions allow analytical solutions to be modified to obtain an answer that otherwise would not have been possible.*
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- *Often aquifers can be assumed to be infinite in areal extent; if not boundaries are assumed to be*
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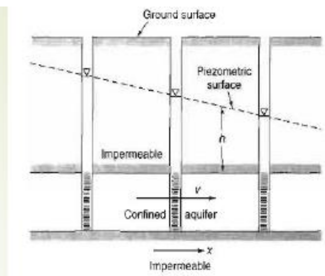
CONFINED AQUIFER

$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=\frac{W}{K}$



- ***The general equation for steady groundwater flow in isotropic homogeneous porous medium is:***

$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=\frac{W}{K} \text{ - - (5c)}$$

- *Now the equation for steady groundwater flow with a velocity 'v' in the x-direction in the confined aquifer (with $W = 0$) will be:*

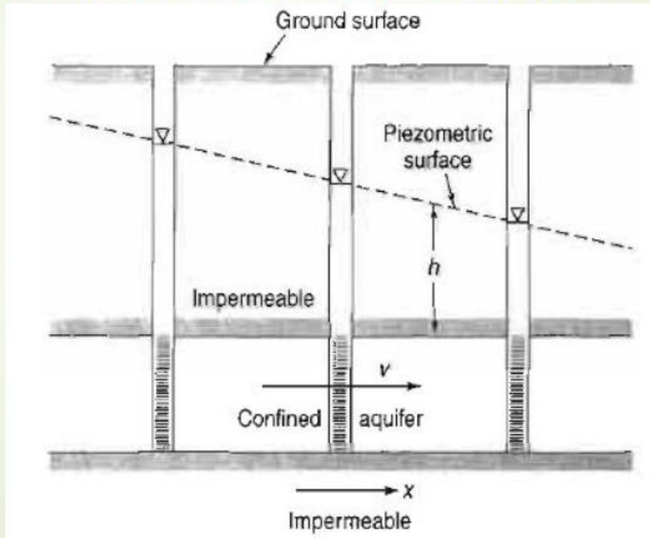
$\frac{\partial^2 h}{\partial x^2} = 0$ *- - (1), the solution of this equation is* $h = C_1 x + C_2$ *- - (1a) where h = the head above a given datum and C_1 & C_2 are constants of integration.*

- *Assuming $h = 0$ when $x = 0$ we get $C_2 = 0$ and $C_1 = \frac{\partial h}{\partial x}$.*
- *Further, from Darcy' law we have $\frac{\partial h}{\partial x} = -(v/K)$ and thus $C_1 = \frac{\partial h}{\partial x}$.*
- *Substituting the value of the constants in (1a) we obtain* $h = -\frac{vx}{K}$ *- - (2)*

CONFINED AQUIFER

$$h = -\frac{vx}{K} \text{ - - (2)}$$

- *This states that the head decreases linearly with the flow in the x-direction.*



Unit 3

1. Explain Cooper-Jacob method for solving the unsteady radial flow into a well in confined aquifer.

COOPER-JACOB METHOD

- *Cooper and Jacob noted that for small values of 'r' and large values of 't', u is small, so that the series in Theis eqn become negligible after the first two terms.*
- *As a result, the drawdown can be expressed by the asymptote:*

$$s = \frac{Q}{4\pi T}\left(-0.5772 - \ln\frac{r^2 S}{4Tt}\right) \text{ - - (3)}$$

- *Rewriting and changing to log to base 10, the eqn reduces to:*

$$s = \frac{2.303Q}{4\pi T}\log\frac{2.25Tt}{r^2 S} \text{ - - (3a)}$$

- *From eqn. (3a) it follows that sVs Log t forms a straight line.*
- *Plotting this line and projecting this to s = 0, where $t = t_0$, we have,*

$$0 = \frac{2.303Q}{4\pi T}\log\frac{2.25Tt_0}{r^2 S}$$

COOPER-JACOB METHOD

- *But we know that Log(1) = 0 and hence we can write that*

$$\frac{2.25Tt_0}{r^2S} = 1 \text{ -- (4) Resulting } s = \frac{2.25Tt_0}{r^2} \text{ -- (4a)}$$

- *A value for T can be obtained by noting that if $t/t_0 = 10$, then $Log(t/t_0) = 1$ and therefore replacing 's' by 'Δs' , where is the drawdown difference per log cycle of 't', eqn (3a) becomes:*

$$s = \frac{2.303Q}{4\pi T}\log\frac{2.25Tt}{r^2S} \text{ -- (3a) } T = \frac{2.303Q}{4\pi\Delta s} \text{ -- (5)}$$

- *The procedure is first to solve for 'T' with eq(5) and then to solve for S with eq(4a).*
- *The straight line approximation for this method should be restricted to small values of 'u' (<0.01) to avoid large errors.*

UNSTEADY (NONEQUILIBRIUM) FLOW - CONFINED AQUIFER

- *When a well penetrating into an extensive confined non-leaky aquifer is pumped at a constant rate, the influence of discharge extends outward with time.*
- *The rate of decline of head times the storage coefficient summed over the area of influence equals the discharge.*
- *Since water comes from a reduction of storage within the aquifer, the head will continue to decline as long as the aquifer is effectively infinite; the flow will be unsteady or transient.*
- *The rate of decline, however, decreases continuously as the area of influence expands.*
- *The applicable equation is :*

$$\frac{\partial^2 h}{\partial r^2} + \frac{1}{r}\frac{\partial h}{\partial r} = \frac{S}{T}\frac{\partial h}{\partial t}$$

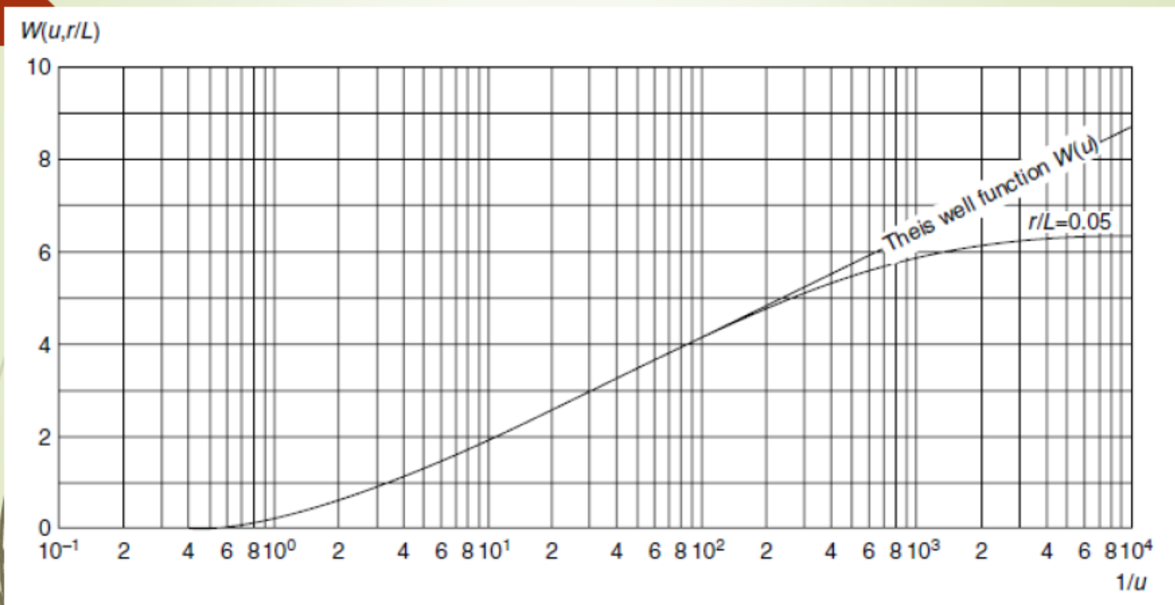
UNSTEADY (NONEQUILIBRIUM) FLOW - CONFINED AQUIFER

- *The applicable equation is :* $\frac{\partial^2 h}{\partial r^2} + \frac{1}{r}\frac{\partial h}{\partial r} = \frac{S}{T}\frac{\partial h}{\partial t}$
- *Where 'h' is head, 'r' is radial distance from the pumping well, 'S' is the storage coefficient, 'T' is the transmissivity and 't' is the time since beginning of pumping.*
- *Theis obtained a solution for this equation based on the analogy between groundwater flow and heat conduction.*
- *By assuming that the well is replaced by a mathematical sink of constant strength and imposing the boundary conditions $h = h_0$ for $t = 0$, and $h \rightarrow h_0$ as $r \rightarrow \infty$ for $t \geq 0$, the solution is:*

- $S = \frac{Q}{4\pi T}\int_u^\infty \frac{e^{-u}}{u}\, du = \frac{Q}{4\pi T} W(u)$ *- (1a) Here W(u) is called "Well Function"*
- ***OR*** $S = \frac{Q}{4\pi T}\left[-0.5772 - \ln u + u - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} - \frac{u^4}{4.4!} + \cdots\right]$ *- - (1)*
- *where 's' is drawdown, 'Q' is constant well discharge and* $u = \frac{r^2 S}{4Tt}$, *, S being the storage coefficient.*
- *Eqn. (1a) is referred as Theis eqn and is widely applied in practice and is preferred over the equilibrium eqn because:*
  - i. *The value of storage coefficient S can be determined,*
  - ii. *Only one observation well is required,*
  - iii. *A shorter period of pumping is generally necessary and*
  - iv. *No assumption of steady state flow conditions is required.*
- *However, the assumptions inherent in eqns (1) should be emphasized because they are often overlooked in applying the non-equilibrium eqn and there by can lead to erroneous results. The assumptions are:*

2. What is “Well Function”?

WELL (HANTUSH) FUNCTION FOR LEAKY AQUIFERS



LEAKY AQUIFERS

- *Hantush function exhibits an S shape and, for large values of 1/u, a horizontal straight-line segment indicating steady state.*
- ***For steady-state drawdown in a leaky aquifer, the following equation was developed by De Glee:***
- $s_m(r) = \frac{Q}{2\pi T} K_0\left(\frac{r}{L}\right)$ *- - (2) where s_m (r) is the steady state stabilized drawdown in meters and $K_0(r/L)$ is the dimensionless modified Bessel Function of the second kind and of zero order (Henkel Function).*
- ***Hantush noted that if r/L is small (<0.05) and L > 3D, eq (2) can, for all practical purposes, be approximated by:***
- $s_m(\boldsymbol{r}) = \frac{2.3Q}{2\pi T} \log \frac{1.12L}{r}$ *- - (3)*

3. Write a note on the Recovery Test.

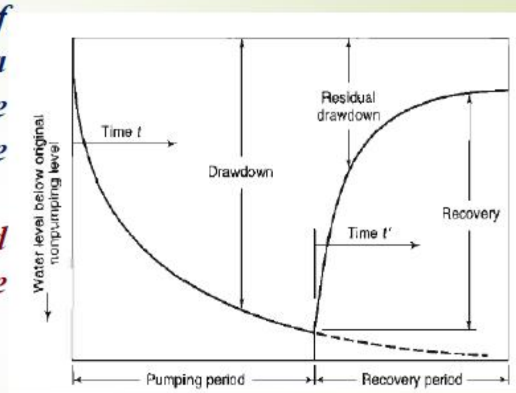
RECOVERY TEST

- ***At the end of a pumping test, when the pump is stopped, the water levels in the pumping and observation wells begin to rise. This is referred as "Recovery" of groundwater levels.***
- ***Measurements of drawdown below the original static level (prior to pumping) during the recovery period are known as "Residual Drawdowns".***
- ***Measurements of residual drawdowns enable the determination of transmissivity.***
- ***The rate of discharge to the well during recovery is assumed to be constant and is equal to the mean pumping rate, whereas pumping rates often vary and are difficult to control accurately in the field.***

RECOVERY TEST

- ***If a well is pumped for a known period of time and then shutdown, the drawdown thereafter will be identically the same as if the discharge had been continued and a hypothetical recharge well with the same flow were superimposed on the discharge well at the instant the discharge is stopped.***
- ***Considering this principle, Theis showed that the residual drawdown s' can be written as:***
- $s' = \frac{Q}{4\pi T}\left[W(u) - W\left(u'\right)\right]$ *- (6)*
- *where* $u = \frac{r^2 S}{4Tt}$ *- (6a) and* $u' = \frac{r^2 S}{4Tt'}$ *- (6b),*

t' and t are as defined in the below figure.



Drawdown and Recovery Curves in an Observation Well

RECOVERY TEST

- *For small 'r' and large 't", the well functions can be approximated by the first two terms of the below eqn.*

$$S = \frac{Q}{4\pi T}\left[-\mathbf{0.5772} - \ln \boldsymbol{u} + \boldsymbol{u} - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} - \frac{u^4}{4.4!} + \cdots \right] \text{-- (1)}$$

- *Thus eqn(6) can be written as*

$$s' = \frac{2.303Q}{4\pi T} \log \frac{t}{t'} \text{-- (7)}$$

- *Thus, a plot of residual drawdown s' versus log(t/t') forms a straight line, the slope of which equals (2.3Q/4T) so that for $\Delta s'$, the residual drawdown per log cycle of (t/t'), the transmissivity becomes,*

$$\boldsymbol{T} = \frac{\mathbf{2.303}\boldsymbol{Q}}{\mathbf{4}\boldsymbol{\pi}\Delta s'}$$

- *However, no comparable value of the storage coefficient can be determined by this recovery test method.*

4. What are the inherent assumptions in the non-equilibrium equation of Theis?

ASSUMPTIONS INHERENT IN THEIS (NONEQUILIBRIUM) EQUATION

1. *The aquifer is homogeneous, isotropic, of uniform thickness, and of infinite areal extent.*
2. *Before pumping, the piezometric surface is horizontal.*
3. *The well is pumped at a constant rate.*
4. *The pumped well penetrates the entire aquifer, and the flow is every where horizontal within the aquifer to the well.*
5. *The well diameter is infinitesimal so that storage within the well can be neglected.*
6. *Water is removed from storage is discharged instantaneously with decline of head.*

5. Write a note on characteristic well losses.

CHARACTERISTIC WELL LOSSES

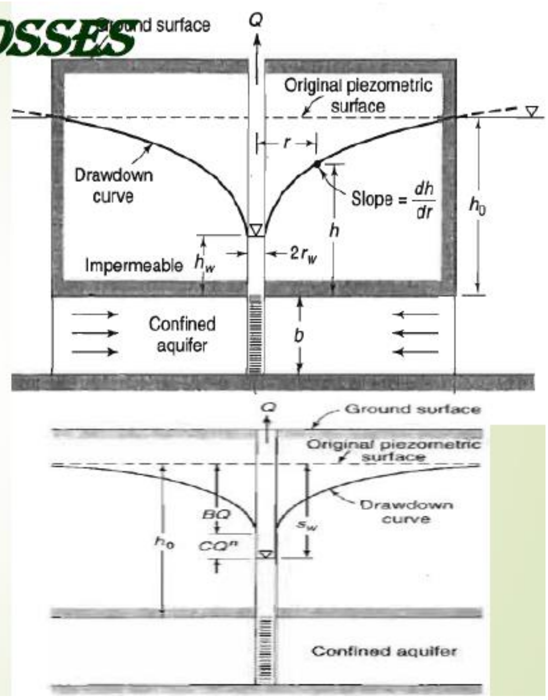
- *The steady state flow equations for confined and unconfined aquifers indicate that the discharge inversely varies with the radial distance from the pumping well according to logarithmic proportion.*
- *Confined aquifer:* $Q = 2\pi T\,(h_2 - h_1)/(\ln(r_2/r_1))$
- *Unconfined aquifer:* $Q = \pi K(h_2^2 - h_1^2)/(\ln(r_2/r_1))$
- *ie.,the drawdown in the well is due to the logarithmic variation of flow through the aquifer. In reality, the drawdown in the well is not only due to the flow thro' the aquifer but also due to the flow thro' the well screen and the flow within the well.*
- *Hence the observed drawdown in the well comprises of that due to logarithmic variation and due to the turbulent flow since the flow through the screen and that within the well are turbulent in nature.*
- *Taking into account the well loss the total drawdown s_w at the well may be written for steady state flow as:*

CHARACTERISTIC WELL LOSSES

- $s_w = \frac{Q}{2\pi T}\ln\frac{r_0}{r_w} + CQ^n$, *where C is a constant governed by the radius, construction and condition of the well.*
- *Letting* $B = \frac{\ln(r_0/r_w)}{2\pi T}$ *we get,*

$$s_w = BQ + CQ^n \quad \text{- - (1)}$$

- *As shown in the fig, the total drawdown s_w consists of the formation loss BQ and the well loss CQ^n.*



CHARACTERISTIC WELL LOSSES

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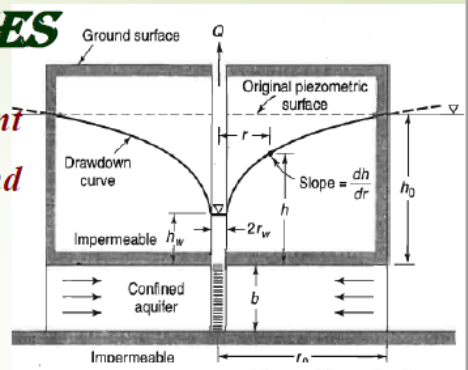
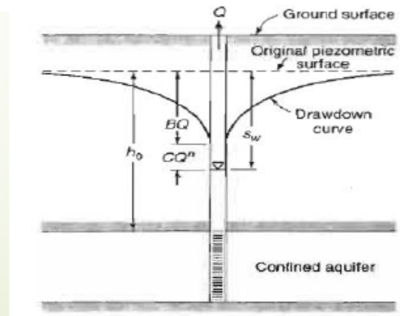
- *As shown in the fig, the total drawdown s_w consists of the formation loss BQ and the well loss CQ^n.*
- *The well loss can be a substantial fraction of the total drawdown when pumping rates are large.*





CHARACTERISTIC WELL LOSSES

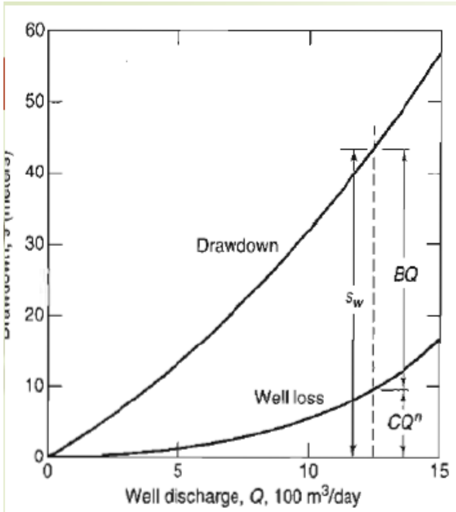
- *From the steady state equation it can be noted that Q varies inversely with $\ln(r_0/r_w)$, if all other variable are held consta*

$$Q = 2\pi Kb\frac{h_0 - h_w}{\ln(r_0/r_w)}$$

- *This shows that the discharge varies only a small amount with well radius.*
- *For example, doubling a well radius increases the discharge only 10%. When the comparison is extended to include well loss, however, the effect is significant.*
- *Doubling the well radius doubles the intake area, reduces entrance velocity to almost half, and (if n=2) cuts the frictional losses to less than a third.*
- *For axial flow within the well, the area increases four times, reducing this loss to less than even greater extent.*

CHARACTERISTIC WELL LOSSES

- *It is apparent that the well loss can be substantial fraction of the total drawdown when the pumping rates are large, as shown in the fig.*
- *With proper design and development of new wells, well loss can be minimized.*
- *Clogging or deterioration of well screens can increase the well losses in old wells.*
- *The criteria for arriving at the well loss coefficient C is given in the table below.*
- *For axial flow within the well, the area increases four times, reducing this loss to less than even greater extent.*



Variation of Total Drawdown

Relation of Well Loss Coefficient to Well Condition

Well loss coefficient C (min^2/m^5)	Well condition
< 0.5	Properly designed and developed
0.5 to 1.0	Mild deterioration or clogging
1.0 to 4.0	Severe deterioration or clogging
> 4.0	Difficult to restore well to original capacity

6. Explain the image well theory as applied to groundwater hydraulics.

Boundaries & Image Wells

Superposition

Consider this now-familiar scenario: if we pump from a well at a constant rate for a certain amount of time, an observation well at some distance away (but not too far away) will experience a certain amount of drawdown. What happens to the observation well if we have more than one pumping well in the general vicinity (Figure 9-1)?

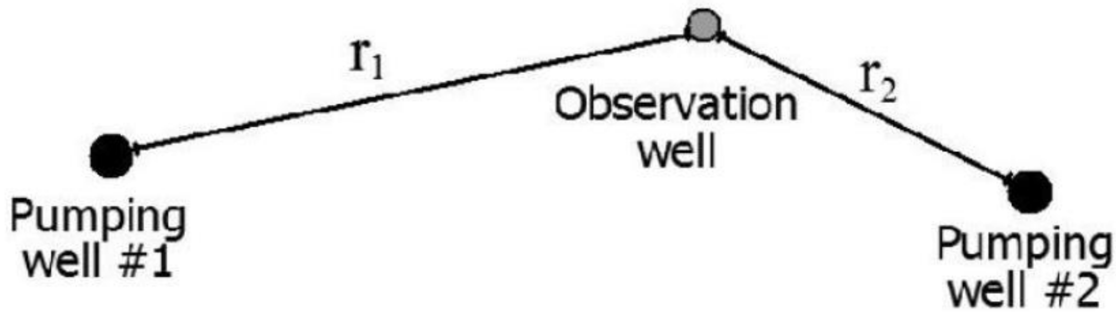


Figure 9-1. Map view of observation well influenced by multiple pumping wells (© Uliana, 2001, 2012).

Each pumping well influences the observation well, and the resulting drawdown in the observation well will be greater. The effect of pumping from one well on the cone of depression developed by a second well is called **well interference**. Fortunately for us, all the equations that we are dealing with are basically linear, so it turns out that the total drawdown at the observation well is simply the sum of the drawdowns induced by each pumping well. This is the principal of **superposition**, that the effects of the two wells sum together to get a total drawdown.

Because of this, it is relatively easy to calculate the drawdown in an observation well influenced by multiple pumping wells, provided that you know T and S and are dealing with a specific point in time. All you do is calculate the drawdown from pumping well #1 at distance r_1, then calculate the drawdown from pumping well #2 at r_2, then sum the two drawdowns to get the total drawdown.

Use of Image Wells

This concept of superposition has a direct application in calculating the drawdown from multiple wells. It also has a second application in dealing with hydraulic boundaries in the aquifer. If you recall, we have two basic hydraulic boundaries – a no-flow boundary and a constant head boundary – and each of these has a specific influence on the drawdown response of the observation well (see Figures 8-2 and 8-3). The effect on drawdown from a boundary in an aquifer is basically the same as the effect of a second pumping well in an unbounded aquifer. Therefore, we can use something called **image wells** to calculate the effect of a boundary on drawdowns from a pumping well.

The basic idea behind image wells is that, if we know the location of the boundary and we know the transmissivity and storativity of our aquifer, we can mathematically place an imaginary pumping well in our aquifer an equal distance from the boundary, but on the *opposite* side of the boundary, and use those two wells and the principal of superposition to calculate the drawdowns in the aquifer (the same way we calculate drawdown from two wells pumping from the same aquifer). An example of a pumping well near a no-flow boundary is shown in figure 9-2.

The pumping well is located some distance $r_{boundary}$ away from the edge of the alluvial aquifer. The boundary created by the uplifted block of granite reduces the amount of water that is available to flow to the pumping well in exactly the same way as a second pumping well in an unbounded aquifer an equal distance from the boundary pumping at the same rate (Figure 9-3). So, if we want to calculate the drawdown at the observation well near the boundary, we would calculate the drawdown from the real well ("Pumping well") at distance r_{real}, then calculate the drawdown from the imaginary well ("Image well") at distance r_{image}, and add the two drawdowns together.

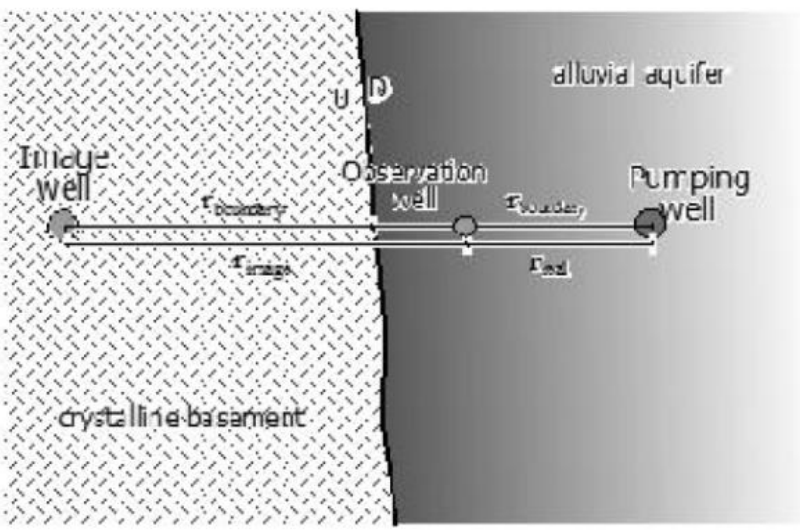


Figure 9-2. Map view of pumping well at no-flow boundary showing image well (© Uliana, 2012).

Use of Image Wells

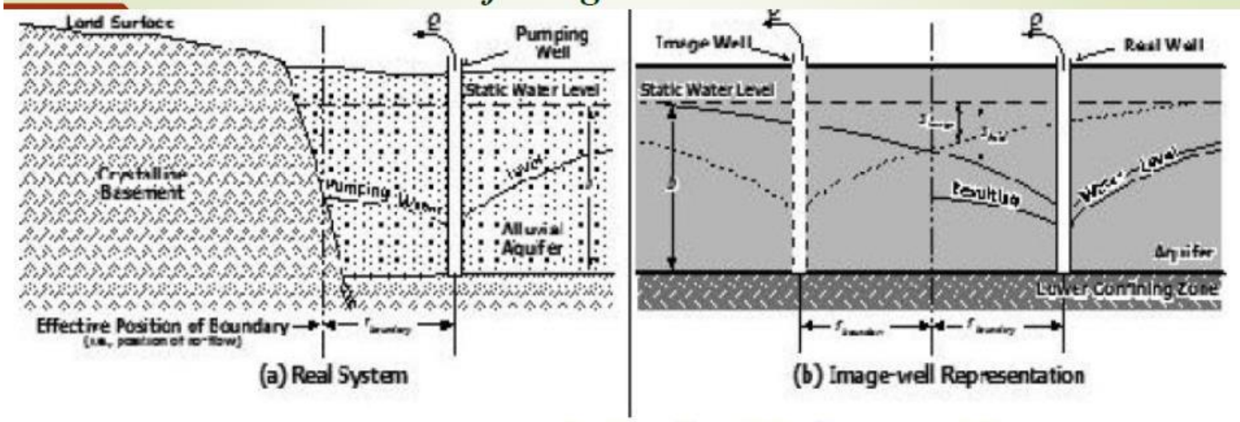


Figure 9-3. Cross-section of a system similar to that in figure 9-2 (© Uliana, 2012).

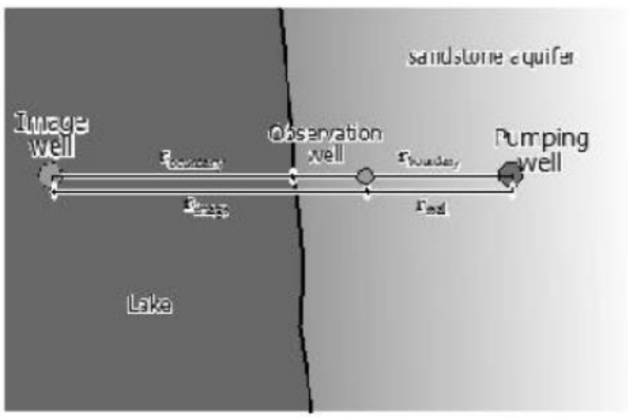


Figure 9-4. Map view of pumping well at constant head boundary showing image well (© Uliana, 2012).

We can do the same thing with a constant head boundary, except that a constant head boundary is supplying additional water to the system and resulting in less drawdown. In that case, the image well is actually an <u>injection well</u>, and the additional drawdown from the image well is <u>subtracted</u> from the pumping well drawdown. (see Figures 9-4 and 9-5).

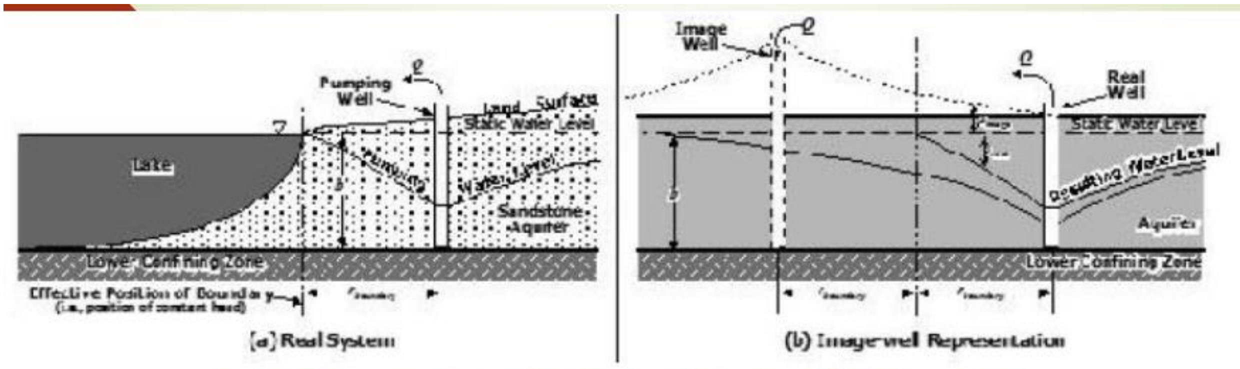


Figure 9-5. Cross-section of a system similar to that in figure 9-4 (© Uliana, 2012).

7. Describe the methodology for analyzing the flow through a leaky aquifer.

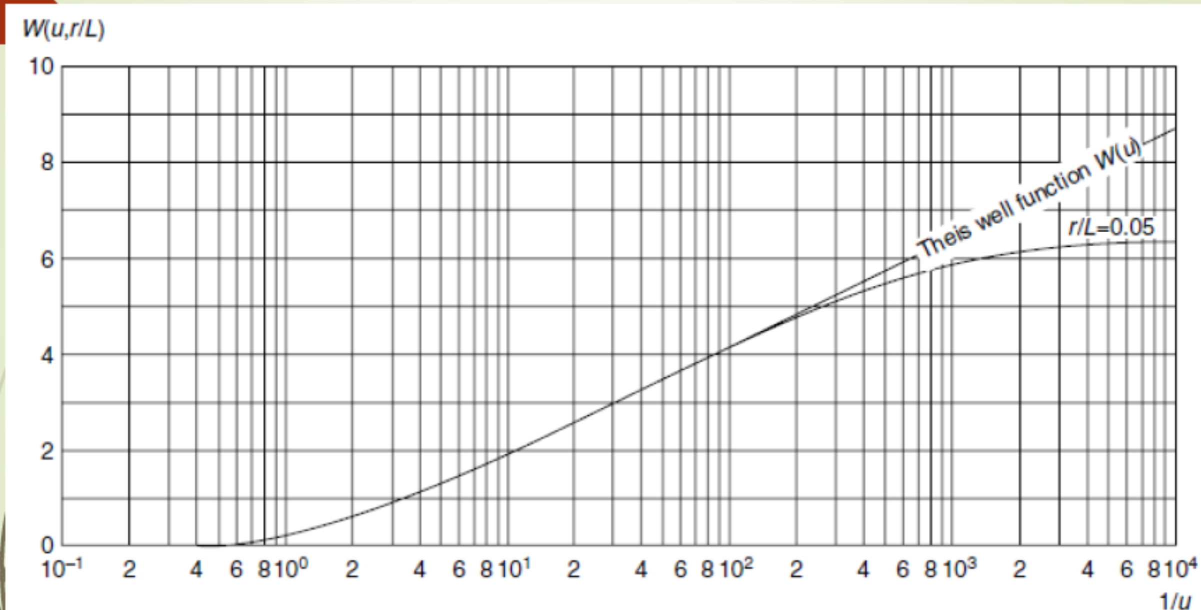
LEAKY AQUIFERS

- ***The drawdown due to pumping in a leaky aquifer is given by the following eqn.:***

$s(r,t) = \frac{Q}{4\pi T}\int_{u}^{\infty}\frac{1}{y}e^{\left(-y-\frac{r^2}{4L^2y}\right)}dy$ ***or*** $s(r,t) = \frac{Q}{4\pi T}W(u, r/L)$ ***- (1) where W(u, r/L) is the dimensionless Well Function (Hantush Function),*** $L = \sqrt{(Tc)}$ ***is the leakage factor or characteristic length in meters, c (D′/K′) is the hydraulic resistance of the aquitard in days,*** D' ***= saturated thickness of the aquitard in meters and*** K' ***= Vertical hydraulic conductivity the aquitard in m/d.***

- ***The above eqn has the same form as the Theis eqn, but there are two parameters in the integral: u and r/L.***
- ***Eqn (1) approaches Theis eqn for large values of L, when the exponential term (r2/4L2y) approaches zero.***

WELL (HANTUSH) FUNCTION FOR LEAKY AQUIFERS



LEAKY AQUIFERS

- *Hantush function exhibits an S shape and, for large values of 1/u, a horizontal straight-line segment indicating steady state.*
- *For steady-state drawdown in a leaky aquifer, the following equation was developed by De Glee:*
- $s_m(r) = \frac{Q}{2\pi T} K_0\left(\frac{r}{L}\right)$ *- - (2) where s_m (r) is the steady state stabilized drawdown in meters and $K_0(r/L)$ is the dimensionless modified Bessel Function of the second kind and of zero order (Henkel Function).*
- *Hantush noted that if r/L is small (<0.05) and L > 3D, eq (2) can, for all practical purposes, be approximated by:*
- $s_m(r) = \frac{2.3Q}{2\pi T} \log \frac{1.12L}{r}$ *- - (3)*

AQUIFER TEST IN LEAKY AQUIFER

- *The physical properties of a leaky aquifer can be found by developing the time-drawdown relationship based on the following equation:*
- $s(r,t) = \frac{Q}{4\pi T} W(u, r/L)$ *- (1)*
- *From the Hantush Well Function type curve it can be noted that the function exhibits an inflection point and for large values of 1/u, a horizontal straight line segment indicating steady state flow.*
- *With this condition, it was shown that the following relationships hold for the inflection point.*

AQUIFER TEST IN LEAKY AQUIFER

1. *The drawdown value s_p is given by:* $s_p = 0.5 s_m$
2. *The u_p value is given by:* $u_p = \frac{r^2 S}{4Tt_p} = \frac{r}{2L}$
3. *The slope of the curve at the inflection point Δs_p per log cycle of time is given by:* $\Delta s_p = \frac{2.3Q}{4\pi T} e^{-r/L}$
4. *At the inflection point, the relation between the drawdown and the slope of the curve is given by:* $2.3 \frac{s_p}{\Delta s_p} = e^{r/L} K_0(r/L)$

8. Explain how a well screen is designed.

SCREENS

- *For consolidated formations where the material surrounding the well is stable, groundwater can enter directly into an uncased well.*
- *In case of unconsolidated wells screens are required for stabilizing the sides of the hole, prevent sand movement into the well and allow a maximum amount of water to entr the well with a minimum hydraulic resistance.*
- *In cable tool method of drilling, screens are normally placed by "Pullback Method".*
- *After casing is in place, the screen is lowered inside and the casing is pulled up to near the top of the screen.*
- *A lead packer ring on the top of the screen is flared outward to form a seal between the inside of the casing and the screen.*

SCREENS

- *For rotary method of drilling without casing, screens are lowered into place as drilling mud is diluted and again sealed by a lead packer to an upper permanent casing.*
- *Screens are also sometimes placed by the "Bail-down method", involving bailing out material below the screen until the screen section is lowered to the desired aquifer depth.*
- *In earlier days well casings used to be perforated by a special cutting knife to serve the purpose of screen. However, this technique now a days is out dated and is discontinued because of the large irregular openings created, small percentage of open area obtained.*
- *Such process also makes it difficult to control the entry of sand with water during pumping.*

SCREENS

- *The current practise is to use a perforated casing, constructed by sawing, machining, or torch-cutting slots in the casing.*
- *Slot openings range from 1 to 6 mm, with larger slots the maximum % of open area is about 12%.*
- *Openings by sawing or machining can be properly sized whereas, torch-cut slots tend to be large, irregular and conducive to entry of sand.*
- *A major factor in controlling the head loss through a perforated well screen is the % of open area. For practical purposes a minimum open area of 15% is desirable.*
- *Manufactured screens are preferred to pre-perforated casing because of the ability to tailor the opening sizes as per the aquifer requirements and the large % of open area that can be achieved.*

SCREENS

- Several types of screens are available - Punched, Stamped, Louvered, Wire-wound perforated pipe and Continuous - slot wire wound screens.
- Screens are available in a range of diameters; selection of screen diameter will be on the basis of desired well yield and the aquifer thickness.
- A parameter that decides the slot size is the screen entrance velocity which depends on the hydraulic conductivity of the aquifer material.

$$v_s = \frac{Q}{c\pi d_s L_s P}$$

- Where v_s = optimum screen entrance velocity, Q = well discharge, c = Clogging coeff. (normally considered as 0.5 ie., 50% of clogging), d_s = screen diameter, L_s = Screen length and P = % of open area in the screen (available from manufacturer's specification).
- Thus for a given aquifer material, aquifer thickness, well yield and type of screen, the appropriate diameter and length of well screen can be selected.

SCREENS

- Screens are made of a variety of metals and metal alloys, plastics, concrete, asbestos - cement, fibreglass - reinforced epoxy, coated base metals and also wood.
- Because a well screen is susceptible to corrosion and incrustation, nonferrous metals, alloys and plastics are preferred for longevity and efficient operation.
- A significant characteristic of well screen is its slot size.
- If the uniformity coefficient of the aquifer sample for naturally developed well (without gravel pack) is 5 or less, the slot size should retain 40 to 50 % of the aquifer material.
- Where a well screen is surrounded by an artificial gravel pack, the size of the screen opening is governed by the size of gravel.